Importance de la nature et du rôle des discontinuités au sein des ophiolites lors du développement d'un orogène

Maryse Ohnenstetter

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UNIVERSITE DE NANCY I
U.E.R. - S.T.M.C.M.
LABORATOIRE DE PÉTROLOGIE

THÈSE
présentée à
L'UNIVERSITE DE NANCY I

pour l'obtention du grade de

DOCTEUR-ES SCIENCES NATURELLES

par

Madame Maryse OHNENSTETTER

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IMPORTANCE DE LA NATURE ET DU ROLE DES DISCONTINUITES AU SEIN DES OPHIOLITES LORS DU DEVELOPPEMENT D'UN OROGENE

Soutenue publiquement le 26 novembre 1982 devant la Commision d'examen composée de :

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Rapporteurs	M. G. LENSCH (Université de Sarrbrücken,R.F.A.)
	M. M.J. O'HARA (Université d'Aberystwyth, G.B.)
Examinateurs	M. W.L. BROWN (Université de Nancy I)
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	M. S.M.F. SHEPPARD (C.R.P.G. - CNRS)

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A la mémoire de

Raphaël GROSDIDIER

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les publications communes. Ajoutons que, au hasard des congrès (à Chypre, en Angleterre et en Italie) ou sur le terrain (dans les Alpes) j'ai pu bénéficier d'échanges fructueux avec de nombreuses personnalités qui ont ainsi contribué au développement de ma pensée, en particulier sur le problème ophiolitique.

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SITUATION DU SUJET

OBJECTIFS ET METHODES

"Mais un fait reste certain : les roches demeurent et les hypothèses passent car même celles qui nous paraissent les plus valables ne représentent qu'un stade d'une perpétuelle recherche vers une meilleure approximation et une meilleure interprétation des phénomènes naturels. Et nous devons avoir le souci constant de désigner les roches de façon telle que la lecture des travaux scientifiques soit toujours possible, dans le présent et dans le futur, quel que soit le sort réservé aux hypothèses du moment".

M. ROUBAULT (1961)

SITUATION DU SUJET. OBJECTIFS ET MÉTHODES

Dans le cadre d'une thèse ou de tout travail de recherche, les résultats obtenus, s'ils dépendent des techniques employées reflètent aussi les conceptions régnant lors de la mise en oeuvre du projet. A la suite des travaux de recherche précédents qui ont conduit à distinguer deux types majeurs d'ophiolites (ROCCI et al., 1975) issus probablement de contextes géotectoniques différents (OHNENSTETTER et OHNENSTETTER, 1975, 1976), l'unicité du modèle ophiolitique ne pouvait être que locale : elle correspond, dans mon esprit, à un processus magmatique qui entraîne des mécanismes de fusion et fractionnements successifs à hautes et basses pressions rendant compte de la présence de résidus, cumulats et liquides. Dès lors, dans l'hypothèse de la tectonique globale, plusieurs modèles de mise en place primaire des ophiolites restaient à définir d'un point de vue pétrologique et structural. Les étudier, c'était contribuer à la connaissance du magmatisme dans des zones à fort gradient géothermique et tectoniquement actives, en distension ou en décrochement, avant d'être incorporées aux continents lors de collisions.

I - SITUATION DU SUJET

1 - Historique - Définition de l'ophiolite

AMSTUTZ en 1980 a rappelé que le terme ophiolite, créé par BRONGNIART (1821) était associé à un mélange de quatre types de roches principaux péridotites, gabbros, diabases et cherts exposés dans les "terrains ophiolitiques" et "formations ophiolitiques".

Les autres définitions fréquemment mentionnées concernent celles de STEINMANN (1927) et de la "Penrose Conference" (1972) dont il est aisé de démontrer la filiation de la seconde par rapport à la première. Les distinctions entre ces deux définitions, qui portent principalement sur la présence ou non de complexe filonien, de chromitite, pyroxénolite ou de roche intrusive felsitique sodique recoupent en fait des discontinuités pétrologiques et structurales aisément observées d'un assemblage ophiolitique à un autre.

Il semble donc souhaitable de revenir à une définition simple qui considère les ophiolites comme une association de roches regroupant des péridotites, des termes grenus intrusifs, des laves et des formations hypabyssales à l'inter-

face entre cumulats et laves. Ces dernières sont recouvertes par des sédiments. L'absence locale d'un des membres de la trilogie de STEINMANN (1927) peut être due à une dénudation tectonique, efficace seulement sur une portion de la chaîne.

Par analogie avec la croûte océanique actuelle, on dit souvent que les ophiolites sont constituées d'une portion de lithosphère représentée par des péridotites le plus souvent de caractère résiduel et d'une croûte classiquement composée de cumulats, roches hypabyssales et laves recouvertes par des sédiments. De cette analogie (BRUNN, 1959; GASS, 1968 ; JUTEAU, 1970 ; ROCCI, 1980; pour l'historique, cf. JUTEAU, 1975), il s'est dégagé une vue unitaire des ophiolites qui seraient créées au droit d'une dorsale médio-océanique et incorporées ensuite au continent. Une autre conséquence fut que les études structurales et pétrologiques menées sur les ophiolites devaient donner des images fidèles des processus se déroulant dans une dorsale.

Néanmoins, simultanément au développement de cette hypothèse, fort féconde d'après le volume d'articles publiés depuis dix ans, d'autres travaux concernant surtout la géochimie des ophiolites ont souligné les divergences à propos de certains magmatismes ophiolitiques et océaniques. Une origine dans un arc insulaire fut alors proposée pour les assemblages ophiolitiques (MIYASHIRO,1973) qui constitue la deuxième grande hypothèse quant à leur environnement géotectonique. Si une nette ressemblance peut être mise en évidence entre les manifestations effusives des arcs immatures et de certaines ophiolites, en fait, cette seconde hypothèse pose d'importants problèmes pour élaborer des reconstitutions paléogéographiques en tenant compte des études géologiques et géophysiques.

C'est alors qu'il convient d'effectuer un bilan sur les ophiolites afin de juger de la validité de ces hypothèses. Celui-ci sera facilité par la mise en oeuvre de travaux analytiques visant à mettre en évidence et à expliquer la nature des discontinuités observées.

2 - Méthodes

Lorsque l'on se propose de créer des modèles pétrogénétiques sur la mise en place primaire des ophiolites, plusieurs techniques peuvent être employées simultanément lors de l'étude des différents termes de l'ophiolite.

En effet, du point de vue structural, le manteau peut présenter des caractéristiques liées à son ascension dans une zone d'accrétion, à son éloignement par rapport à celle-ci ou encore à sa mise en place tectonique sur la mar-

ge continentale. Du point de vue magmatique, des preuves de fusion, contamination, fractionnement et circulation des magmas peuvent y être recherchées à différentes pressions et températures. C'est la confrontation des études pétrologiques et structurales qui permet de préciser la chronologie des processus se déroulant dans le manteau, l'origine possible des magmas primaires et le rôle actif ou passif du manteau.

La croûte océanique dans sa partie basse est composée de roches grenues ayant cristallisé à partir d'un liquide issu du manteau après un cheminement plus ou moins long. Là encore des études structurales, pétrologiques et géochimiques peuvent être menées de front pour décider de la morphologie des chambres, du mécanisme d'injection des magmas dans ces chambres, de leur départ vers la surface et de la nature du fractionnement. Les formations hypabyssales, en écran entre cumulats et laves, peuvent aussi constituer des contraintes aux modèles géochimiques : en particulier, la composition des liquides émis en surface comparée à celle calculée à partir des cumulats permet de rechercher la nature des magmas parentaux en équilibre avec le manteau. Par ailleurs, les zones d'injection des dykes peuvent donner des indications sur les zones d'extension fossiles. Dans ce contexte, la stratigraphie des laves et sédiments facilite la recherche de la position du vecteur gravité. Enfin, au sein d'une chaîne de montagne, les ophiolites sont toujours associées à des termes volcaniques et/ou sédimentaires dont l'histoire permet d'éclairer la durée et le lieu de l'accrétion.

De ces considérations, il ressort qu'entreprendre une étude de terrain géographiquement limitée aurait conduit à une vue partielle des ophiolites, choisir d'analyser une partie de l'assemblage, c'eût été méconnaître le caractère unitaire de la séquence ophiolitique, enfin, entreprendre seulement un travail général sur plusieurs massifs ou effectuer une bibliographie exhaustive, c'eût été soulever les problèmes et ne pas les résoudre. C'est pourquoi, le travail présenté comporte des études détaillées menées simultanément qui portent sur la structure et/ou la pétrologie d'un membre de l'assemblage ophiolitique (cumulats de divers massifs, péridotites du Monte Maggiore) et des travaux à objectifs plus étendus qui concernent la nature des fractionnements de haute et basse pressions et la mise en évidence de discontinuités pétrologiques et structurales au sein et entre les différents assemblages ophiolitiques. La confrontation de ces diverses approches permet de mieux comprendre les mécanismes de l'accrétion ophiolitique et la signification des principaux types d'ophiolites dans un cadre géologique plus vaste.

II - PRESENTATION DU MEMOIRE

Elle a été conditionnée d'une part par le choix des thèmes prioritaires de recherche et d'autre part par la décision de regrouper les articles parus récemment. Un axe de recherche important fut la mise en évidence de discontinuités pétrologiques et structurales entre les sutures de même qu'au sein de celles-ci. Un autre thème fut axé sur la nature des liens magmatiques d'abord entre cumulats et liquides et ensuite entre le manteau et le reste de l'assemblage. Ces différents thèmes ont nécessité la mise en oeuvre de plusieurs techniques et collaborations. Pour ma part, j'ai procédé sur le terrain à l'analyse structurale et pétrologique de quelques massifs étudiés en détail : massif de péridotites du Monte Maggiore, massifs de cumulats d'Iran. Cette technique m'a permis de soulever des questions quant à la migration des magmas du manteau vers la croûte et à leur chronologie, la liaison entre fusion partielle, fractionnement et contamination dans le manteau et la structure interne des massifs ophiolitiques. Au laboratoire, j'ai procédé à l'accumulation et au traitement des données analytiques à partir de massifs échantillonnés ou non par moi-même, à la séparation de minéraux et calculs de fractionnement. Ainsi, il m'a été possible d'apporter des précisions sur la nature des phénomènes magmatiques observés et de quantifier les fractionnements dans plusieurs séries ophiolitiques-types reconnues précédemment.

Le mémoire comprend quatre parties dont les deux premières suffisent à estimer la place des ophiolites dans la reconstitution des chaînes de montagnes tandis que les dernières sont davantage dévolues à l'explication des phénomènes magmatiques et quelquefois structuraux à l'intérieur des principaux assemblages ophiolitiques. Chaque partie est précédée d'un chapitre introductif. A lui seul, il permet d'obtenir une bonne image du problème traité : discontinuités, reconstitutions tectoniques des sutures ophiolitiques, fractionnement dans les chambres magmatiques et processus liés à la remontée diapirique du manteau, en le replaçant par rapport aux connaissances internationales actuelles et aussi par rapport aux autres parties de la thèse. Les articles regroupés ne font qu'illustrer le thème présenté. Précisons que chaque partie ou leur introduction peut être lue séparément, indépendamment des autres. Enfin, un résumé final conclusif permet d'associer les grands thèmes présentés et d'en montrer leur unité.

Dans une première partie, on met en évidence les discontinuités pétrologiques et structurales qui constituent la base des réflexions à propos de la ge-

Les PRINCIPAUX THEMES TRAITES dans les QUATRE PARTIES

	DISCONTINUITES	Partie	MODELISATION	Partie
LIQUIDES Roches hypabyssales Laves	Chronologie	1-3	FRACTIONNEMENT	3
	Geochimie Site Géotectonique	1	Magmas parentaux	
EXTRAITS Cumulats	Chronologie	3		
	Datation	3		
	Ordre de Cristallisation	1-3	FRACTIONNEMENT	3
	Structure Forme des chambres	1	FONCTIONNEMENT des chambres	
			Injection de magma	1
			Températures	3
RESTITES Manteau	Chronologie intra-mantellaire	4	FUSION	3
			FRACTIONNEMENT	3-4
			CONTAMINATION	4
	Structure	1-2 4	MONTEE DU DIAPIR Zone de migration des magmas	4
	Géochimie	3		
	Minéralogie	4	Conditions P et T	

	CLASSIFICATION STRUCTURALE ET PETROLOGIQUE		VALIDITE DE LA CLASSIFICATION	
TERRAINS ASSOCIES OROGENE	Stratigraphie intra-ophiolitique	1	OUVERTURE ET FERMETURE	
	Stratigraphie post-ophiolitique	2	Intra-ophiolitique	1-2 4
	Durée et lieu d'accrétion	2	Intra-plaques	2

nèse des ophiolites. On peut toutefois spécifier que les variations pétrologiques furent démontrées avant les discontinuités structurales, ces dernières s'intégrant en fait parfaitement dans une conception non unitaire des ophiolites. Ces constatations ont permis de préciser dans un second chapitre le rôle des ophiolites et leur signification lors du développement des orogènes, en particulier en bordure du bassin méditérranéen. Dans la troisième partie, la nature des fractionnements liée aux chambres magmatiques sous-crustales de même que la pétrographie et la minéralogie des assemblages sont exposées. On y a développé la chronologie des processus magmatiques et souligné les divergences quant à la nature des magmas parentaux et des mécanismes de fractionnement entre les principaux types d'ophiolites représentés par la Corse et le Vourinos. Cette troisième partie rend compte des discontinuités pétrologiques observées dans la croûte. La quatrième partie, si elle permet aussi d'opposer la nature des résidus des principaux massifs ophiolitiques, s'est proposée de suivre les évènements tectoniques et pétrologiques lors de l'ascension d'un diapir dans un environnement de ride médio-océanique : en particulier, la chronologie des migrations magmatiques liées à des processus de fusion, fractionnement et contamination a été dégagée à partir de l'étude d'une écaille de manteau en Corse.

La conclusion replace les mécanismes d'évolution intra-ophiolitique dans leur cadre géologique : il permet de discuter de la genèse des ophiolites à partir d'exemples types qui tiennent compte des classifications pétrologiques et structurales. On constatera que les sites géotectoniques, bien que théoriquement nombreux, se limitent à quelques types principaux de bordures de plaques dont le jeu relatif entraîne plus ou moins un renouvellement du manteau. Dans ce contexte, l'évolution de la plupart des ophiolites de Méditerranée orientale est inéluctable et liée au développement d'un orogène. Celles du type de Méditerranée occidentale correspond aux premières étapes de la formation d'un océan.

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PREMIÈRE PARTIE

LES DISCONTINUITES PÉTROLOGIQUES ET STRUCTURALES DE LA LITHOSPHÈRE OCÉANIQUE

"Most well preserved and well described ophiolites belong to one class, that a minority belong to another, and it is the class of the minority that most resembles oceanic crust".

J.R. CANN (1976)

PREMIÈRE PARTIE

LES DISCONTINUITES PÉTROLOGIQUES ET STRUCTURALES DE LA LITHOSPHERE OCEANIQUE

INTRODUCTION

Chapitre I - DU PRINCIPE DES DISCONTINUITES

Chapitre II - DISCRIMINATION GEOCHIMIQUE DES LAVES ET DES CUMULATS

Chapitre III - GEOMETRIE INTERNE DES OPHIOLITES. LEUR CLASSIFICATION STRUCTURALE

Chapitre IV - ORGANISATION VARIABLE DU LITAGE DANS LES CUMULATS

Chapitre V - DISCONTINUITE STRATIGRAPHIQUE AU SEIN D'UNE SUTURE

Chapitre VI - CONCLUSION

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PREMIÈRE PARTIE

LES DISCONTINUITES PÉTROLOGIQUES ET STRUCTURALES DE LA LITHOSPHÈRE OCEANIQUE

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En étudiant les ophiolites, on est tour à tour confronté aux problèmes d'unicité et de discontinuité ou non des assemblages. La pérennité et l'invariabilité de l'association, péridotite, cumulat et lave témoignent en fait des processus propres à tout magmatisme : une fusion partielle laisserait un résidu - les péridotites appauvries- tandis que les magmas en route vers la surface donneraient les cumulats et liquides dérivés. Le caractère inéluctable du magmatisme se traduit dans la structure par l'existence de couches de lave, gabbro et péridotite à vitesse sismique différente que l'on peut respectivement comparer aux couches 2et 3 de la croûte océanique et du manteau supérieur (couche 4 de BOTTINGA, 1974). Rappelons que c'est sur cette base que repose, essentiellement, l'analogie entre lithosphère océanique et ophiolites (BRUNN, 1959). Ce principe étant admis, la mise en évidence de discontinuités ne peut se fonder que sur des variations de la nature géochimique et de l'organisation interne des ophiolites dont l'épaisseur des différents termes et leur géométrie.

I - DU PRINCIPE DES DISCONTINUITES

Nos études précédentes ont mis en évidence une discontinuité majeure dans la nature pétrologique des ophiolites méditerranéennes (ROCCI et al., 1975). La dualité des ophiolites implique des manteaux résiduels plus ou moins appauvris : lherzolitique pour le type I en Méditerranée Occidentale, harzburgitique pour le type II en Méditerranée Orientale (rejoignant en cela la classification de NICOLAS et JACKSON, 1972), des ordres de cristallisation différents dans les cumulats avec plagioclase précoce dans le type I, tardif dans le type II, et des liquides riches en TiO_2 (tholéiite à olivine)ou pauvres en cet élément (tholéiite à quartz) respectivement dans les types I et II. Le caractère universel de cette classification apparaît dès lors qu'on la recherche parmi les autres ceintures ophiolitiques d'âge et de lieu différents, (OHNENSTETTER et al., 1975). La distinction pétrographique des assemblages ophiolitiques juxtaposés qui s'accorde avec la distribution géographique, a naturellement conduit à envisager des sites géotectoniques particuliers pour les deux principaux types d'ophiolites : celles du type I seraient issues de ride médio-océanique, tandis que celles du type II seraient formées au voisinage d'un arc insulaire (OHNENSTETTER et OHNENSTETTER, 1976).

	SUCCESSION	TYPE I	TYPE II
	Tectonites	Lherzolites	Harzburgites
1. —	*Peridotites à cumulats*	Harzburgites plagifères, lherzolites	Lherzolites, wehrlites, dunites, chromitites
	Zone de transition	Troctolites	Pyroxénolites
2. —	*Gabbros*	Gabbros à clinopyroxène (euphotides) Ferrogabbros	Norites ou gabbros à olivine Diorites
	Trondjhemites	rares	Tonalites, granophyres abondants
3. —	*Roches hypabyssales*	rares filons, dolérites massives tholéitiques	Complexes filonien développé, souvent à tendance calco-alcaline
4. —	*Laves*	Tholéites à olivine	Tholéites à quartz ± andésites

Fig. I.1 - Tableau de la succession des deux types d'ophiolites téthysiennes ; tiré de ROCCI et al. (1975).

Dès lors, poursuivre des recherches sur la nature et la constitution des ophiolites, c'était, entre autre, accepter comme principe de base le caractère fondamental des discontinuités pétrologiques qu'on pouvait s'attendre à être accompagnées de discontinuités structurales, en raison d'hypothétiques discontinuités géotectoniques. Dans la pratique, cela revient à admettre que chaque assemblage ophiolitique observé selon une coupe perpendiculaire à la chaîne de montagnes est caractéristique d'une suture, mais celle-ci n'en constitue pas forcément la patrie d'origine. Plusieurs sutures parallèles juxtaposées dans une chaîne peuvent avoir des origines communes ou distinctes. De plus, chacune peut varier longitudinalement par suite des écarts de temps lors du début de l'ouverture, de la vitesse d'extension, de sa durée, et des accidents transverses comme c'est le cas au Chili (DE WIT et STERN, 1981) ou en Oman (obs. pers.). Ce simple énoncé permet de juger que la mise en évidence des discontinuités est un projet vaste qui, plus que d'autres, fait appel à de nombreuses références bibliographiques. Bien évidemment la synthèse qui peut en découler, une fois analysés les mécanismes d'accrétion propres aux principaux types d'ophiolites, permet de dégager la genèse de celles-ci. Ce dernier aspect sera exposé en conclusion. On se bornera dans ce chapitre à présenter les travaux personnels qui ont relevé des discontinuités géochimiques et structurales entre les sutures et au sein de certaines d'entre elles. L'explication de la diversité géochimique des basaltes, en tenant compte des fractionnements sera à rechercher dans la troisième partie et

aussi dans la quatrième qui traite du manteau, tandis que la géologie des sutures sera abordée dans la deuxième partie.

II - DISCRIMINATION GEOCHIMIQUE DES LAVES ET CUMULATS

1 - Discrimination géochimique des laves

Etant donné l'unicité de chaque assemblage ophiolitique, la mise en évidence de variations géochimiques entre les ophiolites peut se faire à partir des trois principaux membres de la trilogie. Dans ce cadre, les liquides observés à la partie supérieure des ophiolites, qui représentent des témoins des fractionnements et fusions successifs se déroulant à plus grande profondeur, constituent un excellent champ d'investigation. La plupart des études sur la discrimination des ophiolites portent sur eux, de même que sur les liquides émis actuellement en différents points du globe, afin de proposer des zones géotectoniques possibles propres à leur genèse.

Ce sont les travaux de PEARCE et CANN (1971, 1973) qui ont ouvert la voie dans ce domaine. L'étude de la répartition des éléments peu sensibles à l'altération comme les incompatibles (Ti, Zr, Y), l'évolution de leur rapport dans les différentes séries, leur confrontation avec un élément compatible ont progressivement permis de distinguer dans des diagrammes appropriés les principaux types de basaltes actuels émis en bordure et à l'intérieur des plaques et des liquides ophiolitiques (PEARCE et CANN, 1971, 1973 ; PEARCE, 1975, 1980). On peut souvent prévoir le comportement des incompatibles en considérant la répartition du titane dans les ophiolites. Il a permis d'opposer un groupe riche en Ti d'un autre plus pauvre en cet élément (ROCCI et al., 1975 ; SUN et NESBITT, 1978 ; UPADHYAY et NEALE, 1979 ; SHERVAIS, 1982), respectivement issu de bassins de type médio-océanique ou d'un environnement d'arc insulaire. Appliquée aux cumulats, une même distinction apparaît (SERRI, 1981) qui recoupe les classifications basées sur l'ordre de cristallisation des minéraux (ROCCI et al., 1975 ; CHURCH et RICCIO, 1977 ; OHNENSTETTER et al, 1979).

Le premier article présenté dans cette partie porte sur la discrimination des laves médio-océaniques (OFT) et des arcs insulaires immatures (IAT) à partir du diagramme Ti/Cr-Ni qui permet d'éviter certains artefacts liés aux fractionnements de basses pressions. 97 % des OFT et 93 % des IAT tombent respectivement de part et d'autre d'une ligne tracée de façon empirique. Chaque groupe de laves défini se place soit dans un champ, soit dans l'autre : les ba-

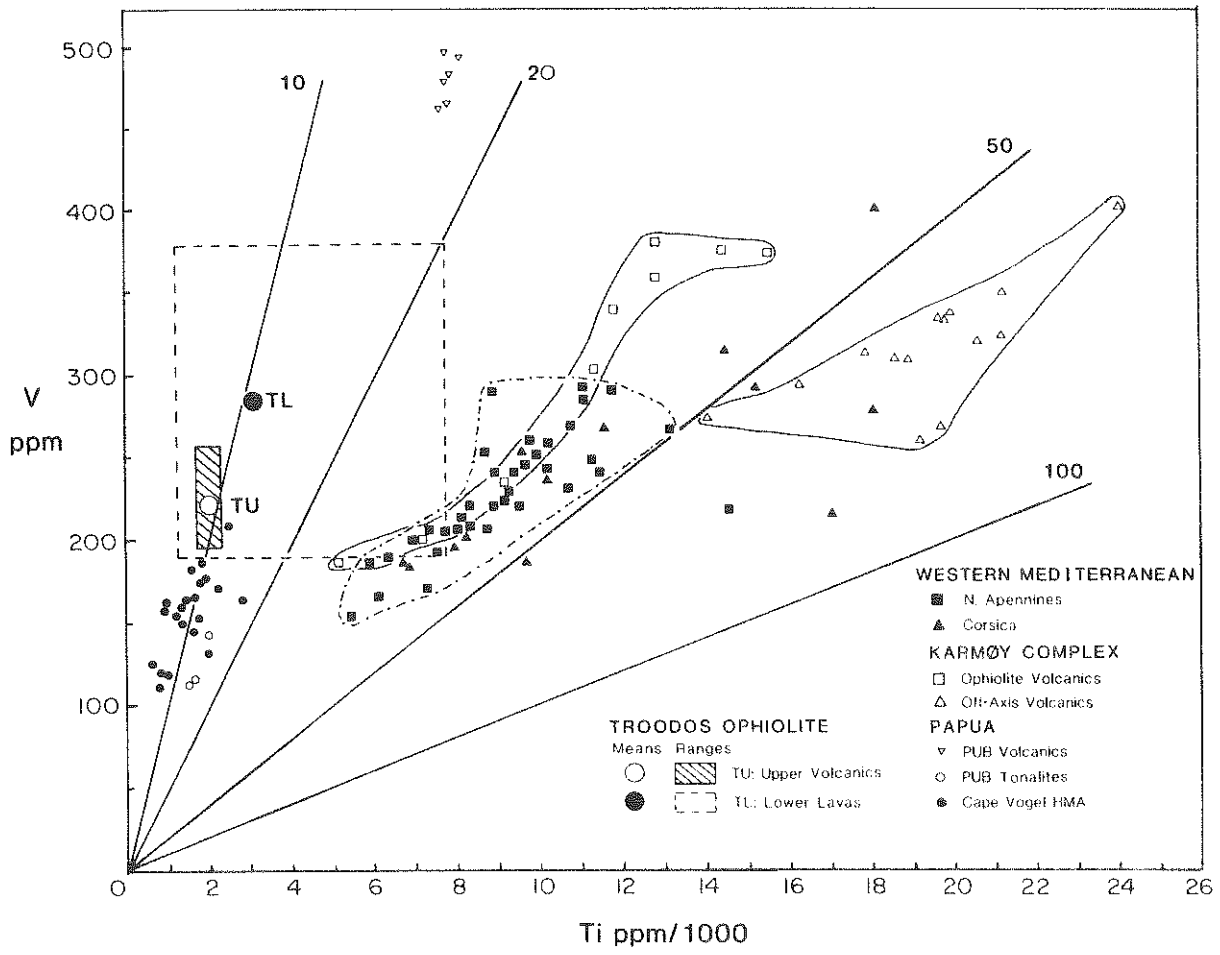

Fig. I.2 - Répartition de Ti et V dans certains massifs ophiolitiques tiré de SHERVAIS (1982). Les ophiolites de Méditerranée Occidentale et de Karmoy ont un rapport de Ti/V compris entre 50 et 55. Leur fractionnement, voisin de celui des MORB, implique surtout OL + Pl. Les volcanites "off axis" de Karmoy ont une origine dans une île intra-océanique. Pour les ophiolites de Méditérranée Orientale, les laves de la ceinture ultramafique de Papouasie (PUB) et les andésites riches en Mg de Cape Vogel (HMA), Ti/V 0 implique une formation près d'une plaque convergente (résumé d'après SHERVAIS, 1982).

saltes ophiolitiques dans le domaine des OFT peuvent être issus d'un océan ou d'un bassin marginal ; ceux qui sont aussi associés avec des basaltes ayant des caractères d'IAT proviennent d'un environnement de marge active : arc et bassin marginal ou arc et croûte océanique avant-arc. De plus cet article souligne que dans de nombreuses ophiolites ayant des caractères d'arcs, les laves les plus jeunes,par exemple les "upper pillow lavas" du Troodos sont plus appauvris en

éléments incompatibles et souvent picritiques. La genèse des laves ophiolitiques serait alors expliquée par une fusion à plusieurs stades (SMEWING et POTTS, 1976) libérant des masses de magma qui évoluent avec le temps (BECCALUVA et al., 1979;I.1 BAROZ et al. III.3 ; OHNENSTETTER II.3).

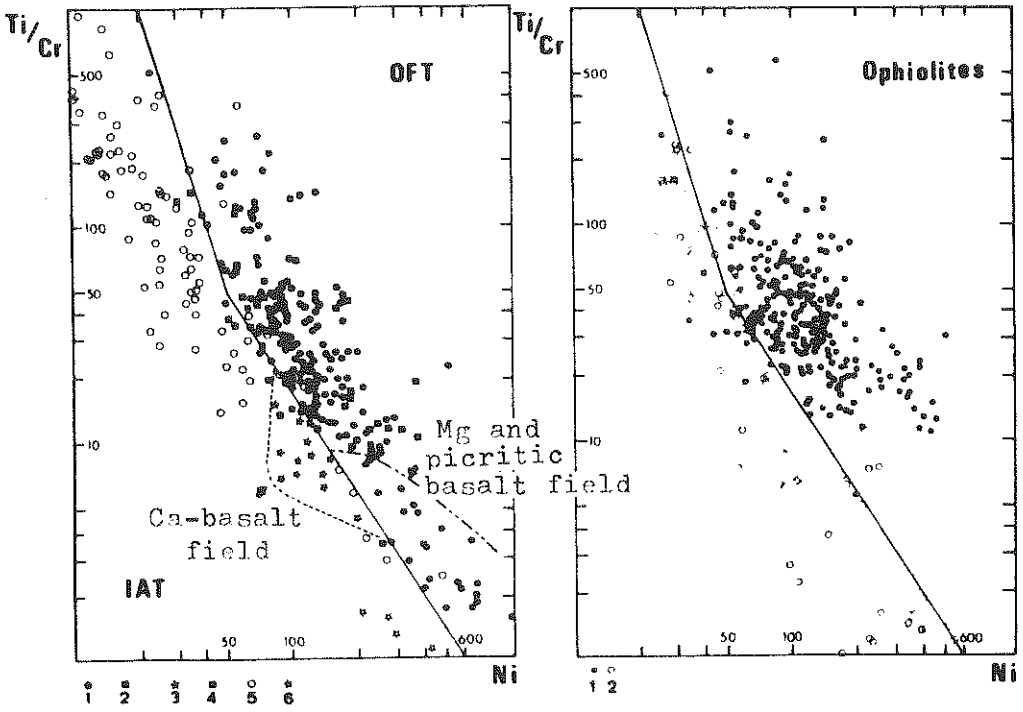

Fig. I.3 - Diagramme de discrimination Ti/Cr - Ni tiré de BECCALUVA et al., 1979. Il regroupe à gauche des roches basaltiques 40 % < SiO_2 < 56 % des arcs immatures, IAT (5) et des rides médio-océaniques, OFT (1). Les basaltes calciques du Leg 37 sont indiqués (3), des basaltes de bassins marginaux (2) et en avant de l'arc (4), de même que les boninites (6). A droite, les laves des ophiolites tombent de part et d'autre d'une droite empirique.

L'utilisation d'éléments incompatibles les moins sensibles à l'altération comme P, Ti, Zr et Y et de quelques éléments à fort coefficient de partage comme Ni et Cr permet simplement dans les ophiolites de reconnaître les massifs qui ont de fortes affinités soit avec les basaltes des dorsales médio-océaniques soit avec les tholéïtes pauvres en potassium des arcs immatures. Une distinction entre ophiolites de segments océaniques $\pm$ anormaux, de bassins marginaux intra-océanique ou intra-continental nécessite l'emploi d'autres groupes d'éléments susceptibles de mettre en évidence des discontinuités géochimiques comme dans les sites géotectoniques actuels.

Parmi les éléments hygromagmatophiles (TREUIL et JORON, 1975), on peut définir les éléments faiblement attractifs (low field strenght : LFS) dont le rapport rayon/charge est supérieur à 2 : Cs, Rb, K, Ba, Sr, Th et U.. Ces éléments LFS couplés avec La et Ce constituent le groupe des ions larges et lithophiles (Large ion lithophile : LIL) défini par SCHILLING, (1973).

Les concentrations et les rapports entre LFS et HFS permettent de séparer nettement les basaltes des arcs immatures de ceux des océans et d'estimer l'influence de la zone de subduction sur la composition des laves des bassins marginaux (SAUNDERS et al., 1980) qui serait transitionnelle entre océan et arc (GILL, 1976 ; WEAVER et al., 1979 ; SAUNDERS et al., 1980).

L'enrichissement en LFS, dans le cas de massif ophiolitique dont le comportement des majeurs et des éléments HFS est voisin de celui des arcs insulaires permet de préciser les variations spatio-temporelles de la composante magmatique liée à une zone de subduction. Si la composition des laves ophiolitiques est proche de celle des basaltes médio-océaniques pour la plupart des éléments, des concentrations plus élevées en éléments LFS dans certains massifs pourraient traduire une origine dans un bassin marginal. De ce point de vue, les ophiolites de la chaîne alpine seraient les témoins d'un océan et celles de Sarmiento, d'un bassin marginal. Une discussion quant à l'origine de l'enrichissement en éléments LFS peut être soulevée à partir de la comparaison des bassins marginaux intra-océanique et intra-continental. Dans le cas du massif de Sarmiento, qui présente les mêmes concentrations en éléments LFS qui celui de Bransfield en domaine océanique, les teneurs élevées de ces éléments pourraient être reliées soit à une zone de subduction, soit à un environnement plus continental (SAUNDERS et al., 1980).

Enfin, tout au long des dorsales médio-océaniques, les éléments lithophiles de même que les rapports isotopiques de Sr et Pb ont facilité la séparation de segments normaux et anormaux, ces derniers étant caractérisés par un magmatisme proche de celui à l'origine des basaltes transitionnels et alcalins. Cette distinction refléterait une hétérogénéité mantellaire (SCHILLING, 1973 ; BRYAN et al., 1976 ; WHITE et SCHILLING, 1978 ; WOOD et al., 1979). En ce qui concerne les ophiolites de la chaîne alpine des segments de croûte peuvent être séparés en fonction d'une part de la latitude comme dans les rides actuelles et d'autre part en fonction de l'âge (OHNENSTETTER M., 1980, I.4), l'ouverture des

bassins pouvant être reliée à un magmatisme plus transitionnel (VENTURELLI et al., 1981).

2 - DISCRIMINATION PETROLOGIQUE DES CUMULATS

Le comportement des liquides dans des diagrammes basés soit sur la concentration de SiO_2 soit sur la valeur du rapport $Mg/Mg+Fe^{2+}$ traduit l'existence de fractionnements dans les ophiolites. A priori, des extractions ont eu lieu à faible profondeur comme en témoigne la présence de cumulats. Leur proportion et succession peuvent donner des indications sur la composition et l'évolution des différents magmas parentaux (3ème partie). En conséquence, l'ordre de cristallisation qui dépend de la nature du magma entrant dans la chambre, peut servir à caractériser les différentes suites magmatiques entre les sutures et au sein de celles-ci.

CRYSTALLIZATION ORDER	ol-pl-cpx-opx	ol-cpx-pl-opx	ol-cpx-opx-pl	ol-opx-cpx-pl	
		Diopsidperidotitisches Variationsprofil	Pyroxenitisches V.	Enstatitperidotitisches V.	*Bötticher 1969*
	Type I (Troctolitic)	Type II (Pyroxenitic)			*Rocci & Ohnenstetter 1975*
	Class D	B	C	A	*Church & Riccio 1977*
	Type I	II-A	II-B	II-C	*This study*
CUMULATE	Pl-dunite Pl-wehrlite troctolite Ol-gabbro gabbro Mt-gabbronorite ferrogabbro leucoferrogabbro albitite	dunite ± chromitite wehrlite clinopyroxenite Ol-gabbro gabbronorite Mt-gabbronorite diorite granophyre	dunite+chromitite wehrlite websterite gabbronorite Q-diorite granophyre	dunite++chromitite wehrlite+Pl orthopyroxenite norite Q-noritic-gabbro Q-diorite granophyre	
	Type I High TiO_2 Ol-Tholeiite(MORB)	Type II mainly Q-Tholeiite (with C.A. affinity) Low TiO_2			*Rocci & Ohnenstetter 1975*
	High Ti β	Low Ti β			*Sun & Nesbitt 1978*
LIQUID	OFT	IAT	boninitic affinity		*Beccaluva & al. 1979*
	High TiO_2 (HT)	Low TiO_2 (LT)	Very Low TiO_2 (VLT)		*Bébien & al. 1980*
	High HFS.E.	Low HFS.E.	Very Low HFS.E.		*This study*

Tab. I.1 - Classification pétrographique des cumulats basée sur l'ordre de cristallisation. La nature des liquides associés.

TYPE I		TYPE II		
Magmatisme voisin des rides médio-océaniques		Magmatisme proche de celui des arcs insulaires		
Océanisation avec marges continentales passives	Bassins marginaux	Bassins marginaux ± proche de l'arc		
		TYPE II-A	TYPE II-B	TYPE II-C
Corse	Sargaz Abshur	Kellaki	Papouasie	Neyriz
Alpes	Bay of island	Troodos	Troodos	Troodos
Apennins	Pindos	Esfandageh	Nurabad, Iran	Nain
	Othrys	Solemanieh	Betts Cove	Betts Cove
		Bay of island	Serpi Köy	Syrie
	avec une contamination crustale possible :	Pindos	North Cape	North Cape
		Krapa, Vourinos		Nurabad, Iran
	Chili	Pozanti-Kersanti		
	Hokkaido	Hatay		
	Guevgueli	Cirali-Tekivora		
		Nouvelle-Calédonie		
		Othrys		
EXTENSION ET DECROCHEMENTS		AUGMENTATION DE LA CONTAMINATION EN LFS →		
		ROLE ACCRU DES DECROCHEMENTS		
Exemples actuels :		Exemples actuels : le Pacifique ouest près du continent		
Mer Rouge	Détroit de Bransfield			
Océan Atlantique	Mer d'Ecosse			

Tab. I.2 - Exemples de massifs ophiolitiques illustrant la classification proposée et leur environnement géotectonique compte tenu des équivalents actuels.

Les minéraux à considérer sont peu nombreux : Pl, Cpx et Opx si l'on tient compte que Ol et Sp sont toujours précoces en raison de l'expansion du volume primaire de l'olivine vers la surface et de la compatibilité des éléments entrant dans le spinelle. Parmi les six possibilités de suites minéralogiques, nous en avons retenu quatre qui sont aussi les plus couramment observées dans les

ophiolites (Tab. I.1).Les ordres indiqués soulignent la précocité du plagioclase dans les séries proches de celles des rides médio-océaniques (Type I celle du Cpx et plus rarement de l'Opx dans les ophiolites de type II (Tab. I.1).Dans ces dernières, il faut considérer que les limites entre les groupes sont graduelles d'une ceinture à une autre et aussi au sein d'un même assemblage, puisque les magmas ophiolitiques évoluent également avec le temps comme l'indique la stratigraphie des laves (voir aussi la 3ème partie). Des séquences de cumulats types peuvent être déduites d'après les données de terrain, en accord avec l'ordre de cristallisation et la variation dans la proportion des minéraux. Dans le tableau I.1,les principales caractéristiques des liquides supposés être en équilibre avec les cumulats ont été reportées. On soulignera que les magmas riches en éléments incompatibles dont le titane, sont associés à des cumulats à Pl précoce, tandis que les liquides à TiO_2 faible coexistent avec des suites de cumulats riches en Cpx. Lorsque TiO_2 décroît, les séries deviennent de plus en plus sursaturées en SiO_2 ce qui s'exprime par la cristallisation de l'Opx en début d'évolution.

Du point de vue géotectonique, on peut interpréter les séries à Pl précoce (Type I) et à pyroxène précoce, comme respectivement situées loin de la zone d'influence d'un arc ou à proximité d'une zone de subduction (Tab. I.2, fig. 4). Si l'on considère seulement les ophiolites de type II, les différentes séries peuvent se succéder dans une même ceinture. L'évolution du magmatisme serait liée à un appauvrissement progressif du manteau qui aurait donc un rôle passif. Dans les massifs ophiolitiques de Type II, si les suites pyroxénitiques sont les plus développées, il n'est pas rare d'observer dans certaines portions de la chaîne des cumulats riches en Pl à proximité du manteau comme à Chypre (ALLEN, 1975), en Oman (SMEWING, 1980) ou encore dans la suture de Sabzevar, Iran (ALAVI-TEHRANI, 1976).

III - GEOMETRIE INTERNE DES OPHIOLITES - LEUR CLASSIFICATION STRUCTURALE

La nature et la succession des différente termes au sein des ophiolites dépendent de processus liés à la genèse des magmas et des profondeurs de mise en place. Cependant la géométrie interne des termes, leur extension latérale et verticale et leur rapport avec les autres membres de l'ophiolite peuvent refléter des conditions particulières d'ouverture.

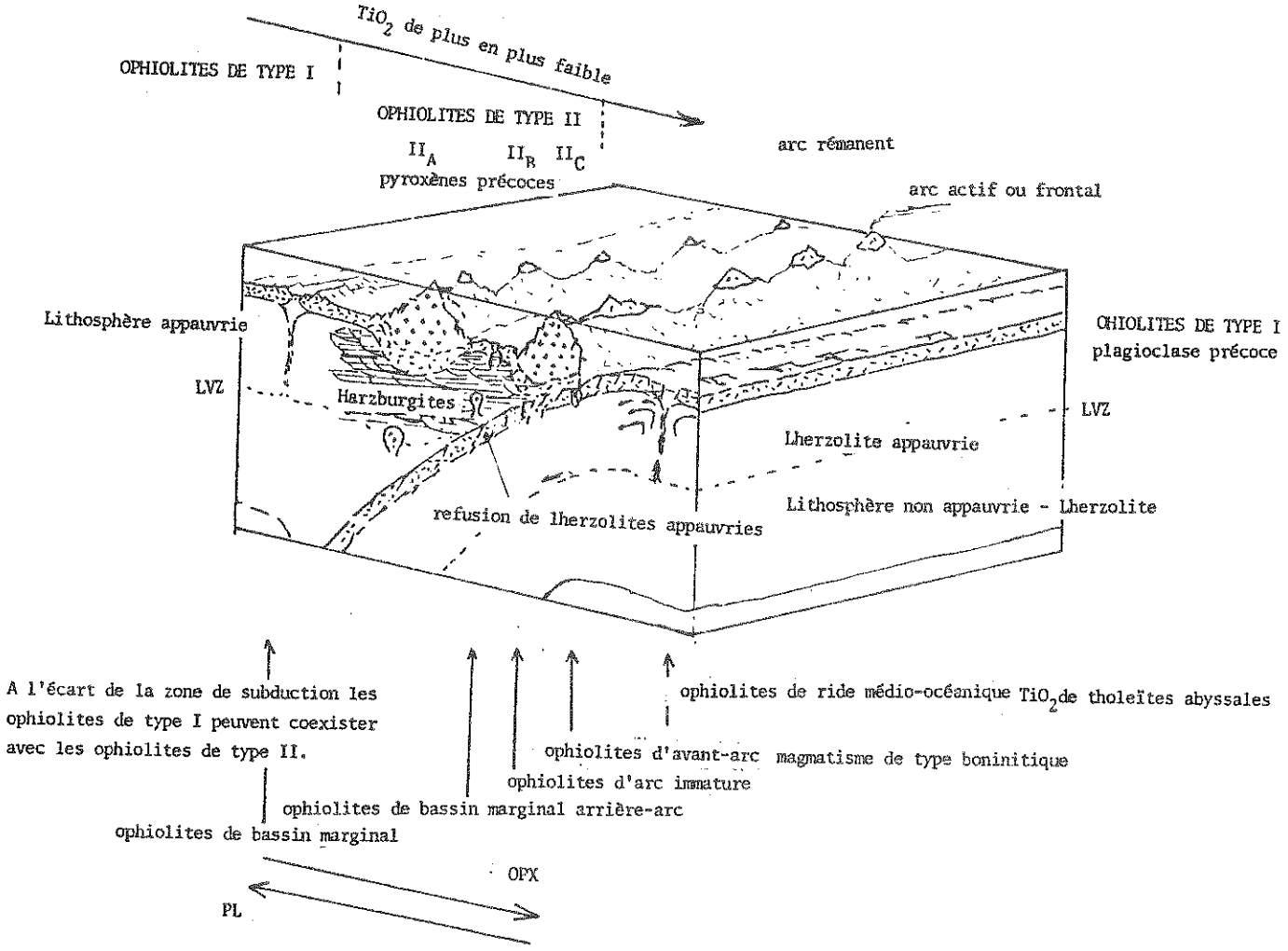

Fig. I.4 - Disposition des différentes séries ophiolitiques au voisinage d'une marge active.

L'organisation interne des ophiolites peut être définie par la position d'un certain nombre de marqueurs d'origine lithologique et/ou structural. Leur nature et signification, permet de décrire les processus tectono-magmatiques intervenant dans le manteau (OHNENSTETTER M., IV.4), dans les chambres magmatiques (OHNENSTETTER M., I.3) ou près de la surface. Leur position théorique a été précisée dans trois systèmes de contraintes (OHNENSTETTER M., 1980, I.2).
Cette deuxième publication (I.2) précise que selon le système envisagé, les diapirs de péridotites seront plus ou moins déformés, les chambres plus ou moins spacieuses et allongées et l'orientation des filons variable. Dans certains

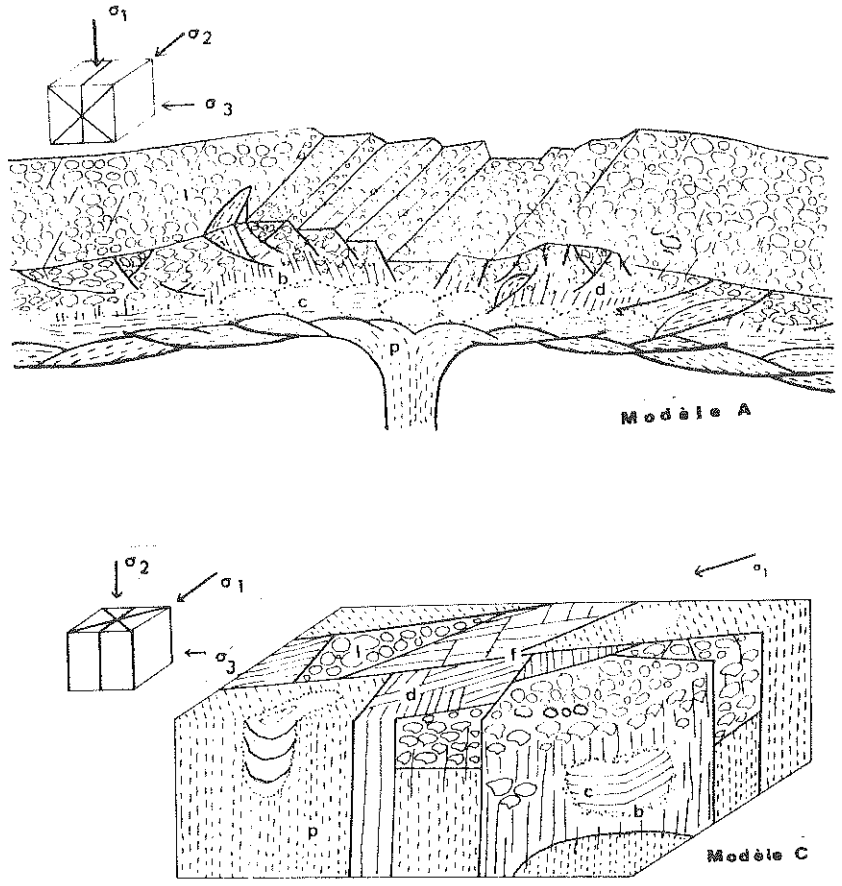

Fig. I.5 - Position des marqueurs lithologiques et structuraux des ophiolites dans les systèmes en extension (A) et en décrochement (C). Tiré de OHNENSTETTER M. (1980). A noter dans le second cas, l'importance des discontinuités latérales, les petites chambres et les textures de diapirs préservées.

systèmes, tels les décrochements,(modèle C), les discontinuités latérales et verticales seraient plus importantes comparées aux systèmes en extension (Fig. 5 modèle A).

Par ailleurs, dans la même note, une compilation bibliographique a permis de préciser que compte tenu de l'état actuel des connaissances sur l'organisation interne des ophiolites, celles-ci pouvaient être classées en deux groupes (fig. I.6).Un premier ensemble formé par les massifs les plus puissants : Nouvelle-Calédonie, Papouasie, Antalya, Oman et Bays of Islands possèdant des sections bien stratifiées où les variations lithologiques dans les péridotites et les cumulats inférieurs sont subhorizontales. D'après les linéations d'étirement dans les péridotites ou la disposition du complexe filonien, on peut aisé-

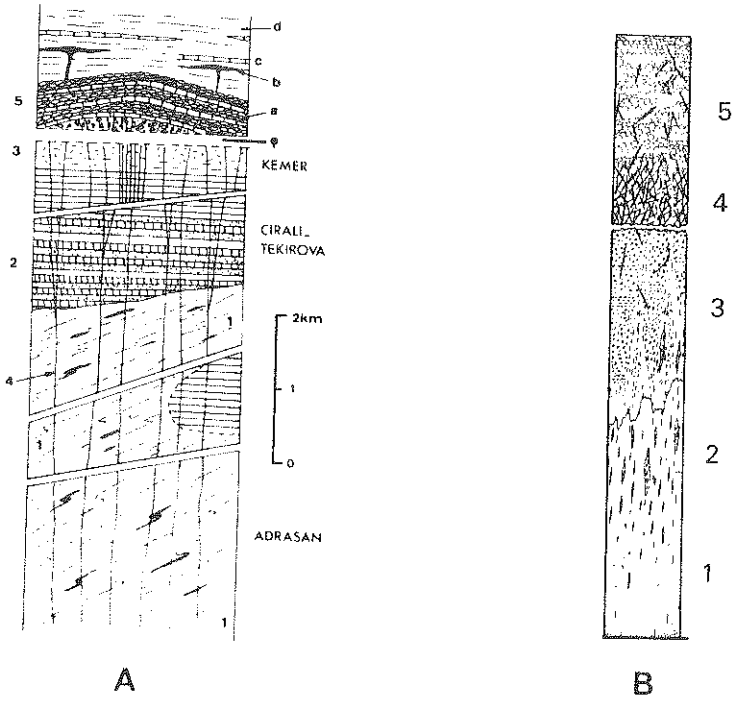

Fig.I.6 - Logs des deux principaux types structuraux d'ophiolites. A: Log d'Antalya, tiré de JUTEAU et al., 1977. B: Log de King River, Cordillère Occidentale de l'Amérique du Nord, tiré de SALLEBY, 1977. On remarquera l'importance des plans marqueurs en position verticale pour la partie basale du log B (péridotites et certains cumulats). A: 1, harzburgite foliée et veines de gabbros et de pyroxénite déformées; 2, séquence de cumulats mafique et ultramafique rubanés (dunite, diallagite, gabbro); 3, Gabbro à amphibole, diorite quartzique; 4, dyke de diabase recoupant 1, 2 et 3 ; 5, série volcano-sédimentaire. B: 1, harzburgite foliée avec inclusions de dunite chromifère, wehrlite et amphibolite ; 2, zone de transition déformée avec dykes de dunite, harzburgite, wehrlite, clinopyroxénite, gabbro, diorite et plagiogranite ; 3, cumulat de gabbro à olivine et dyke de même composition ; les niveaux inférieurs sont riches en clinopyroxénite ; 4, zone de dyke avec écran de gabbro ; 5, pillow lava recoupé de dyke.

ment retrouver l'axe de la ride (JUTEAU et al., 1977). Cette disposition est compatible avec un système en extension (fig.I.7) tel qu'il a été proposé par JUTEAU et al. (1977) ou DEWEY et KIDD (1977). Le deuxième groupe d'ophiolites rassemble de très nombreux massifs dont celui du Troodos, Vourinos, Sabzevar,

Acoje, Canyon Mountain. Dans ces massifs, la foliation des péridotites est verticale (diapirs) tandis que les cumulats peu abondants sont probablement issus de petites chambres. Enfin, les filons ont une orientation variable. L'ensemble de ces faits serait compatible avec un système en décrochement qui interviendrait dans des contextes géotectoniques variés : marges actives et passives, bordure de microcontinent. Il favoriserait, en outre, la mise en place secondaire des ophiolites sans phénomène de subduction ou d'obduction majeur (STONELEY, 1975).

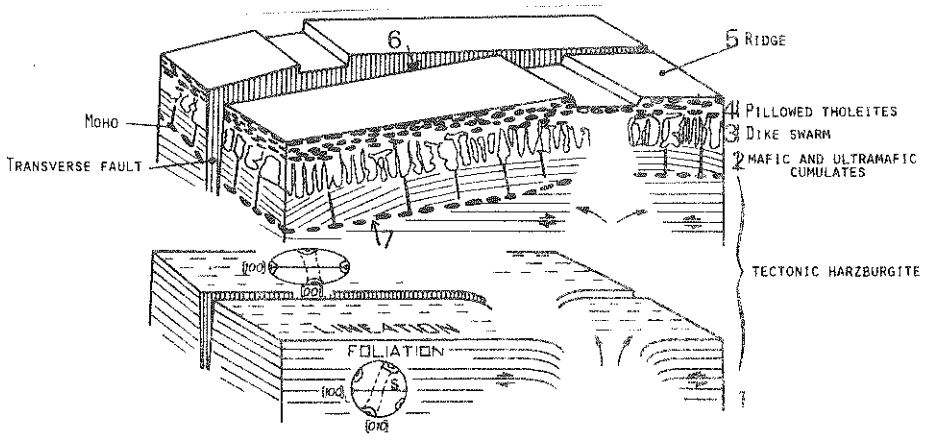

Fig. I.7 - Structure de la croûte et du manteau océanique au voisinage de la ride, tiré de JUTEAU et al., 1977. 1, harzburgite foliée ; 2, cumulat mafique et ultramafique ; 3, complexe filonien ; 4, pillow lava ; 5, ride ; 6, faille transverse ; 7, Moho.

En effet, on verra dans la deuxième partie, que le premier groupe d'ophiolites est largement charrié sur la plate-forme continentale tandis que le second groupe est à proximité de son lieu d'origine. Il y aurait alors variation dans les conditions de mise en place primaire et secondaire qui serait liée au site géotectonique de départ.

Cette classification structurale confrontée à la classification géochimique des ophiolites (conclusion) permet de constater que les ophiolites les plus pauvres en TiO_2 sont aussi celles qui appartiennent au second groupe d'ophiolites à structures verticalisées. Dans ces dernières, quel que soit le système de contrainte envisagé, l'extension fut limitée.

IV - ORGANISATION VARIABLE DU LITAGE DANS LES CUMULATS

Les autres articles rapportés à cette partie mettent en évidence des mécanismes de fonctionnement ou des discontinuités propres à une suture ophiolitique. Dans un premier temps, on envisagera les processus d'accumulation dans une chambre magmatique à partir du gisement du rubanement. Cette approche, importante pour saisir la nature du fractionnement en partie responsable de la composition des liquides observés (3ème partie), l'est aussi pour comprendre les discontinuités structurales et pétrologiques dans les cumulats.

En effet, le plan de variation lithologique dans les cumulats est un marqueur important utilisé lors de l'élaboration des modèles d'accrétion. Il soulignerait la forme de la chambre magmatique (CASEY et KARSON, 1981). La recherche de sa nature et de sa position originelle est donc fondamentale. Dans un premier temps le plan du rubanement a été assimilé à un plan de stratification à partir duquel des croissances de minéraux pouvaient avoir lieu comme dans le cas des complexes stratiformes (WAGER et al., 1960). Plusieurs modèles proposés récemment permettent de diversifier leur origine et surtout d'envisager que le pendage du litage peut être très accentué. Signalons principalement le modèle de MAC BIRNEY et NOYES (1979) qui met l'accent sur les cristallisations aux parois des chambres et celui d'IRVINE (1980) qui fait intervenir la mise en mouvement de tranches de liquide de densité et de viscosité différentes. Dans ce contexte la signification des cumulats sub-horizontaux formés à distance variable de la ride (GREENBAUM, 1972) et issus de remplissages successifs (JACKSON et al., 1975; JUTEAU et WHITECHURCH, 1980 ; PALLISTER et HOPSON, 1981) est à préciser par rapport aux roches grenues litées sub-verticalement (fig.I.8). Par ailleurs, les ophiolites s'étant formées dans des zones tectoniquement actives, la mise en place des magmas peut être syntectonique. On peut alors s'attendre à ce que les processus d'évolution d'une chambre de type complexe stratiforme soit perturbée et que sa morphologie qui dépend des conditions d'ouverture soit variable d'une suture ophiolitique à une autre (OHNENSTETTER M., 1980, I.2).

Dans les ophiolites de Sabzevar qui appartiennent au second groupe structural (OHNENSTETTER M., 1980, I.2 ; OHNENSTETTER, I.3), la morphologie de plusieurs chambres magmatiques, de même que les mécanismes d'alimentation ont été déduits à partir de la position des plans lithologiques dans les cumulats (Fig.I.9).

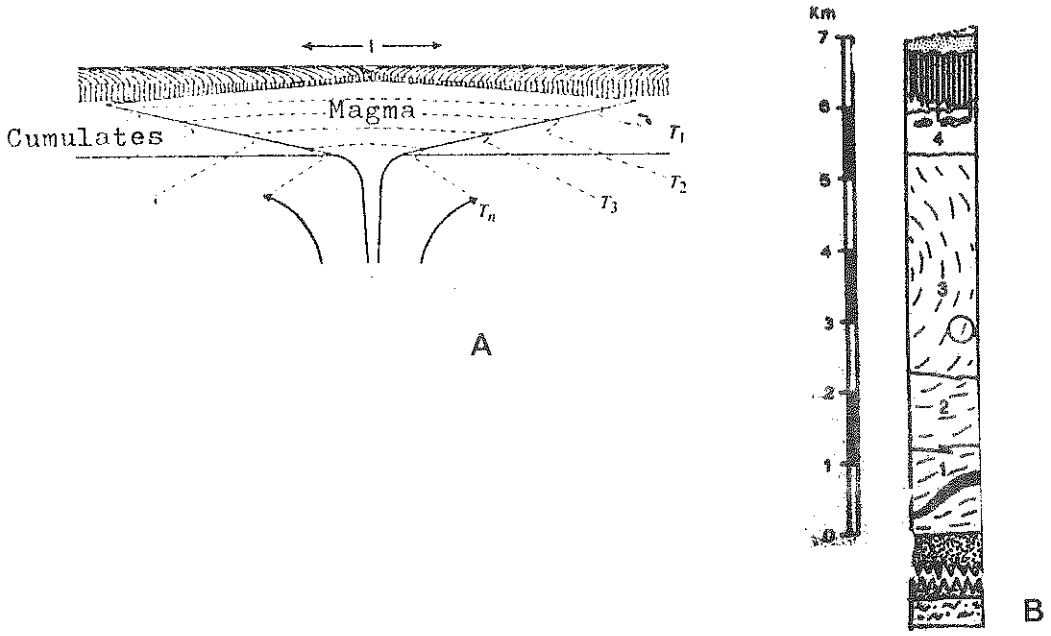

Fig. I.8 - Disposition du rubanement des cumulats à proximité de la ride A : modèle de GREENBAUM (1972) ; les différents isothermes correspondent aussi au dépôt des cumulats divers. B : modèle de CASEY et KARSON (1981) ; on notera la disposition verticale du litage vers le haut de la chambre. Au-dessus des tectonites foliées : 1, dunite, wehrlite, clinopyroxénite ; 2, zone de transition avec des cumulats mafiques et ultramafiques ; 3, anorthosite, troctolite, gabbro à olivine et gabbro dans lesquels le litage devient vertical ; 4, gabbro isotrope surmonté du complexe filonien et des laves.

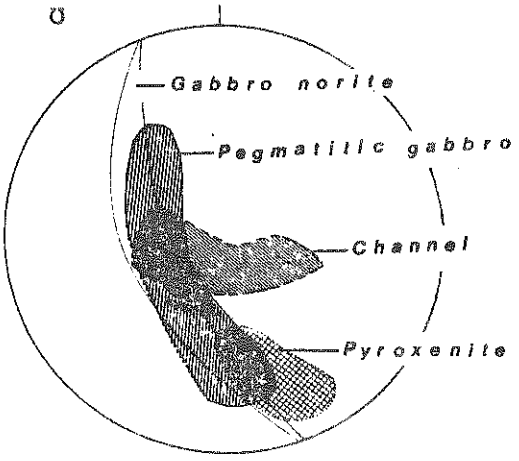

Fig. I.9 - Disposition des chenaux et des roches grenues sécantes dans le haut d'une chambre magmatique de Sabzevar, tiré de OHNENSTETTER M. I.3.

Il est noté que les rubanements verticaux existent principalement vers la base supposée des chambres et vers le sommet. Ils sont subparallèles à des fractures

permettant la montée des magmas et limitant les chambres. Suivant les cas, des cristallisations perpendiculaires aux bordures ou des différenciations par fluage ("flow layering") peuvent être observées. Celles-ci dominent vers le haut des chambres (THAYER, 1963) ce qui correspond à la fin de l'évolution du magmatisme ophiolitique.

En plus de ces cumulats fortement pentés, il existe des cumulats sub-horizontaux ayant cristallisé à proximité du plancher temporaire de la chambre. Le raccord entre certaines de ces roches se fait par l'intermédiaire de cumulats au rubanement incliné qui mime la morphologie de chenaux. Ces derniers apparaissent plus ou moins dissymétriques et profonds par suite du jeu local différentiel des failles sur leur bordure. De plus, la puissance des couches et la taille des grains diminuent régulièrement de bas en haut, traduisant une réduction progressive du volume de liquide résiduel dans la chambre en dépit des arrivées de magmas parentaux qui ont été plus abondants en début d'évolution. Les derniers cumulats et les liquides qui leur correspondent attestent de la diminution du volume des adjonctions de magma et de leur changement de composition avec le temps. Un même schéma pétrogénétique est aussi proposé pour la série magmatique du Vourinos (BECCALUVA et al., II.6).

V - DISCONTINUITE STRATIGRAPHIQUE AU SEIN D'UNE SUTURE

Dans les ophiolites, peu de sédiments sont intercalés dans les laves. En général, le développement d'une série sédimentaire au-dessus des laves donne l'âge minimum d'accrétion, d'ailleurs voisin des âges absolus obtenus sur les roches magmatiques ophiolitiques (conf. 2ème partie). La nature des sédiments peut être compatible avec un environnement géotectonique tel que océan, et/ou bassin marginal plus ou moins proche d'un arc. En cela, elle dépend de l'éloignement de la zone continentale : il apparaît que les sédiments supra-ophiolitiques sont souvent plus siliceux que ceux des fonds océaniques (ROBERTSON, 1981). Dans une même suture, on peut s'attendre à trouver des variations latérales de faciès en fonction de l'âge des ophiolites (fig. I.10A) : les roches vertes les plus anciennes à proximité du continent comme en Corse, étant coiffées d'une sédimentation plus grossière et carbonatée (CARON et DELCEY, 1979 ; CARON et al., 1979). Par ailleurs, des irrégularités au niveau du plancher océanique, par suite de la présence de rides transverses, peuvent conduire à des modifications dans la bathy-

métrie. Celles-ci, de même que l'intervention de mouvements tectoniques seront responsables des discontinuités sédimentaires observées. Par exemple à Chypre, l'importance des brèches dans la zone d'Arakapas (paléofaille transformante) comparée aux séries de ride du Troodos a été soulignée depuis longtemps (WILSON, 1959 ; SIMONIAN et GASS, 1978 ; BECHON, 1982 ; BECHON et ROCCI, 1982).

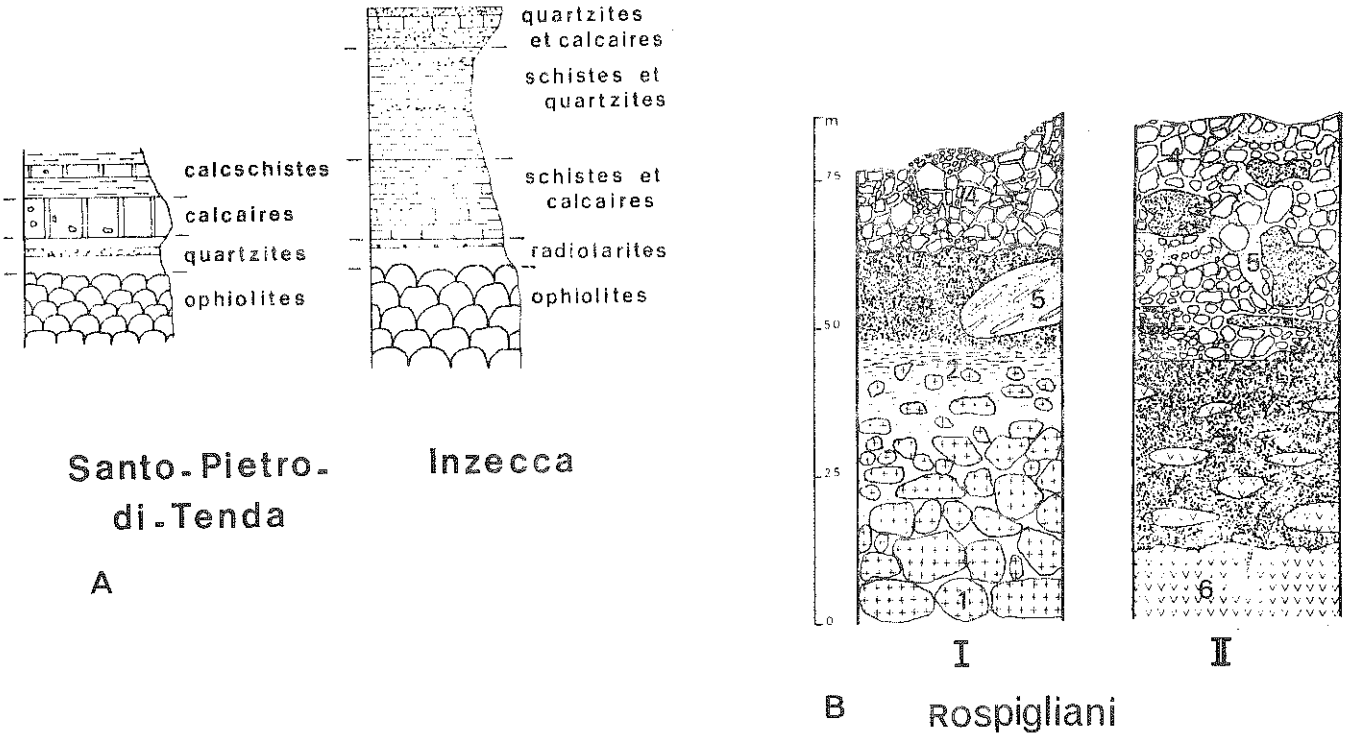

Fig. I.10 - Colonnes stratigraphiques supra-ophiolitique (A) et ophiolitique (B). A : Séries de l'Inzecca et de Santo Pietro di Tenda tiré de CARON et al. (1979) ; elles mettent en évidence des caractères distaux (Inzecca) et proximaux (Santo Pietro di Tenda) par rapport à la marge passive. B : Série de Rospigliani, tiré de OHNENSTETTER M., 1979 ; elle souligne les caractères de la série volcano-sédimentaire au-dessus d'un paléo-plancher de cumulat gabbroïque (I) ou de péridotite serpentinisée (II). 1, cumulat gabbroïque ; 2, sédiment détritique immature à éléments de socle océanique ; 3, Ophicalcite ; 4, agglomérat ; 5, intrusion d'albitite et de basalte contemporains; 6, serpentinite.

Les deux derniers articles (I.4 et 5) qui ont trait aux ophiolites de Corse soulignent que la diversité des séquences est liée d'une part à leur création dans une ride ou dans une zone transformante et d'autre part au stade d'évolution du rift. Dans ce dernier cas, les discontinuités sédimentaires observées traduisent la distance du rift par rapport aux marges (CARON et al, 1979 ; CARON et DELCEY, 1979). A côté de sections ophiolitiques homogènes caractéristiques de zones de ride où une puissante série de laves repose théoriquement sur des cumulats par l'intermédiaire d'une carapace de dolérites, des sections variables latéralement et verticalement leur sont juxtaposées (série de Rospigliani). Dans cette dernière qui se serait développée dans une zone transformante, des sédiments peu remaniés constitués d'éléments arrachés au socle océanique (gabbros, péridotites plus ou moins serpentinisées, dolérites) quelquefois transformés (ophicalcites) alternent avec des agglomérats (fig.I.10 B) dont la composition varie de basalte à basalte andésitique riche en fer. L'évolution des liquides et leur injection vers la surface contemporaine de la mise en place des albitites atteste de l'existence de chambres magmatiques sous-jacentes. Enfin, la remontée de diapirs de serpentinites dans ces zones expliquerait le démantèlement précoce du fond océanique. Les mouvements tectoniques se poursuivraient au Crétacé inférieur et seraient responsables de la création d'olistolithes (LEMOINE et TRICART, 1979). Signalons enfin que les caractères de la série de Rospigliani se retrouvent dans certains lambeaux ophiolitiques tout au long de la chaîne alpine (ABBATE et al., 1972 ; ELTER, 1972 ; GIANELLI et PRINCIPI, 1974 ; LAGABRIELLE, 1982 ; BERTRAND et al., 1981). Les détails des relations entre séries de rides et de transformantes sont maintenant à préciser.

VI - CONCLUSION

Ainsi, il est possible, entre les ceintures ophiolitiques, de mettre en évidence des discontinuités géochimiques et structurales liées à leur environnement géotectonique qui peut être extrêmement varié : océan, bassin marginal, arc et zones transverses à l'intérieur des plaques. Si l'analyse géométrique aide à préciser les mécanismes d'accrétion, en aucun cas, elle ne permet de conclure à un site géotectonique particulier. Celui-ci est à cerner d'abord en considérant la composition géochimique des différents termes de l'ophiolite qui nous apparaît donc comme un des meilleurs moyens d'investigation sur l'origine d'une cein-

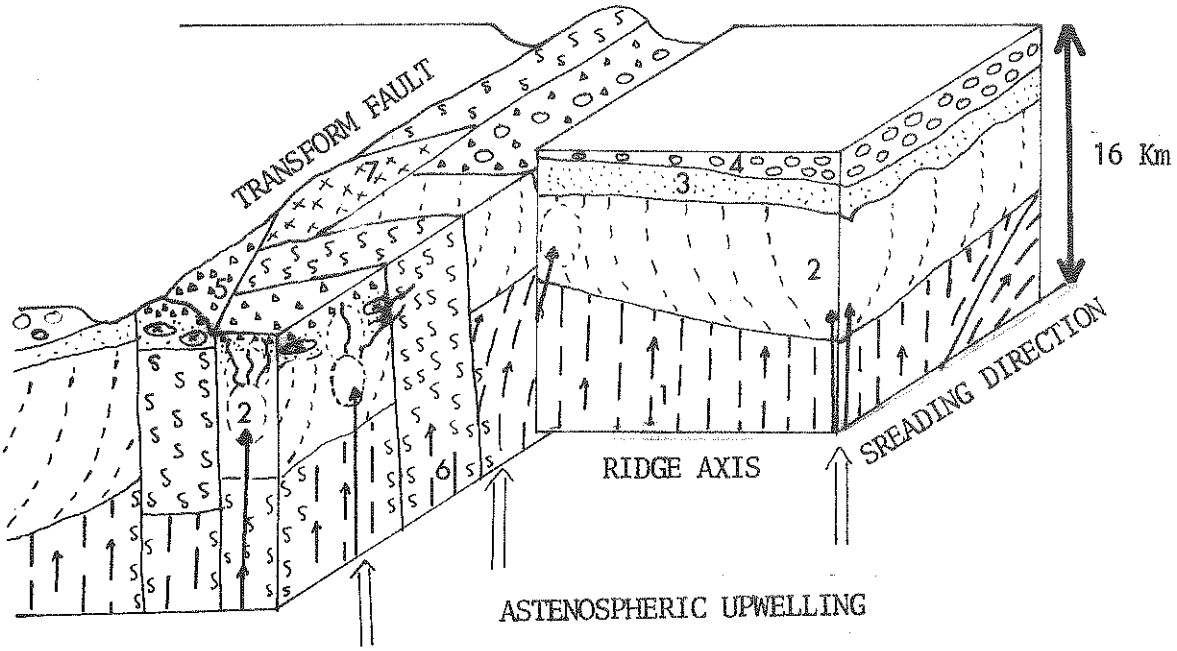

Fig. I.11 - Relation entre ride et transformante d'après l'interprétation des logs ophiolitiques corses. 1, manteau lherzolitique ; les lignes de flux sont indiquées ; 2, chambres magmatiques alimentées par les magmas parentaux qui donnent latéralement les cumulats ; 2' représente des intrusions plus petites à proximité des transformantes ; 3, dolérite massive et cumulats de congélation ; 4, pillow lava dont l'épaisseur se réduit vers les zones transformantes; 5, agglomérats ; 6, diapirs de serpentinites ; 7, ophicalcite dérivant soit de serpentinite, soit de gabbros. 1, 2, 3 et 4 forment la série de ride. 2', 5, 6 et 7 constituent la série type des zones transformantes.

ture. En tenant compte des données pétrographiques et structurales, il est certain qu'un modèle d'accrétion plus affiné peut être élaboré.

Dans le cas de la Corse (Type I), on pourra se reporter au schéma de la Fig. I.11. La foliation subverticale du manteau résulte d'une mise en place diapirique. Un fluage sub-horizontal, s'il n'est pas à exclure étant donné les vitesses sismiques enregistrées perpendiculairement à l'axe des dorsales actuelles et l'interprétation qui peut en être tirée (FRANCIS, 1969), n'est peut être pas systématique en début d'ouverture. Le manteau sert d'encaissant, de même que les laves inférieures et les dolérites, à des chambres de petite dimension emboîtées que l'on peut trouver le long des dorsales et dans les zones transformantes. A l'aplomb des rides, le système de convection dans l'asthénosphère engendre un

renouvellement des zones de fusion ce qui permet aux liquides successivement produits de ne pas trop s'appauvrir avec le temps.

Pour les ophiolites de Type II, un schéma relatif à leur formation doit tenir compte des variations pétrologiques successives et de l'importance et de la durée de l'extension. Suivant la profondeur de mise en place du diapir et des chambres magmatiques associées, les premiers cumulats formés à séries à Pl précoce proche du Type I (tab.I.1) et quelquefois séries riches en Cpx (Type II.A), seront plissées et déformées avec les péridotites. Ces cumulats tectonisés serviront à leur tour d'encaissant à des chambres magmatiques plus superficielles où un rubanement qui mime les parois des chambres pourra se développer. Ce sont ces roches qui donnent les affleurements de cumulats les plus importants dans les massifs ophiolitiques. Enfin, l'évolution de la chambre, de nombreux cumulats verticaux (pyroxénites plagifères, gabbros noritiques), attestent de l'importance des phénomènes d'injection et des départs vers la surface. Ils sont recoupés par des gabbros et diorites à grain variable qui constituent des corps intrusifs à faible profondeur dans la carapace de laves et de roches hypabyssales. Les mélanges magmatiques de diorite, granophyre et basalte témoignent de la persistance des montées de magma dont le volume décroît néanmoins avec le temps. Certaines séries sursaturées en silice proviendraient des derniers épisodes du magmatisme ophiolitique (série II.C).

Le rubanement est fruste au sein des dernières intrusions de faibles dimensions ou dans les filons. Au début de l'évolution de la chambre magmatique principale (série II A), l'alimentation de la chambre a lieu par de nombreuses fissures qui fragmentent le plancher. Progressivement, les injections sont concentrées dans les zones de faiblesse au voisinage des parois de la chambre. Celles-ci seraient aussi dépendantes du réseau de fractures qui délimitent la chambre et la morphologie des bassins. L'épaisseur des cumulats varierait latéralement : elle serait plus importante dans des zones en extension encadrées par des zones mobiles en relief où se concentrent les fractures. Si l'expansion dure assez longtemps, et qu'elle soit favorisée par un système en extension,plutôt qu'en décrochement,un certain état d'équilibre peut être atteint, qui se traduira par un plus grand développement des séries cumulatives. On peut dans ce cas, envisager une participation plus active du manteau dont les lignes de flux peuvent tourner pour obtenir le dispositif classique observé dans les ophiolites à plans structuraux et lithologiques sub-horizontaux (Fig.I.7). D'une manière

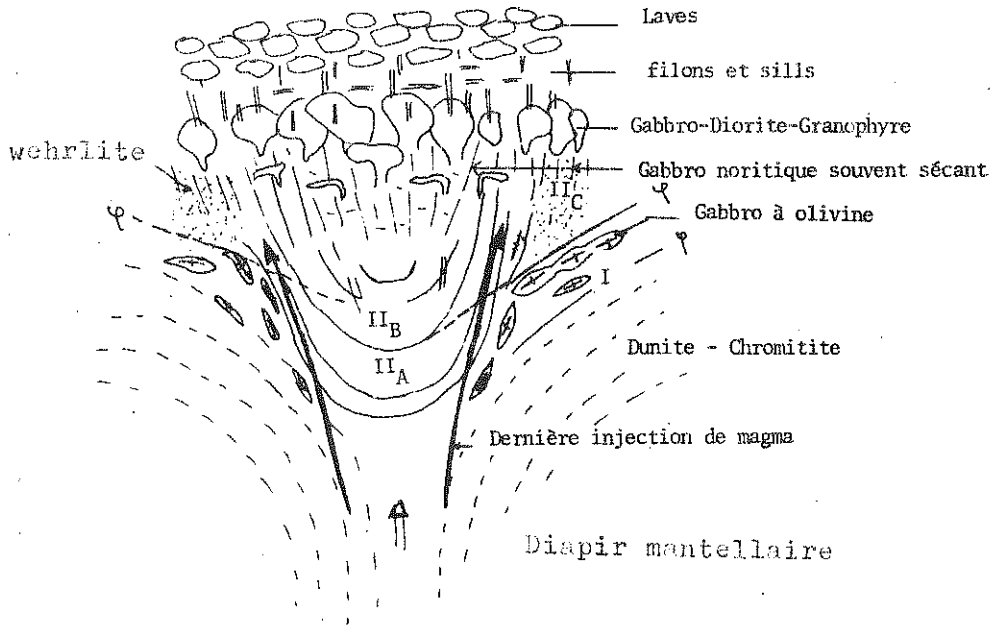

Fig. I.12 - Séquence ophiolitique schématique reconstituée d'après les observations de terrain.

générale, l'évolution couplée des liquides et cumulats des ophiolites de Type II, même si elle est localement retardée, témoigne d'un appauvrissement progressif du manteau et de son rôle passif lors de l'expansion.

Ces considérations sur les variations structurales et pétrologiques intra-ophiolitiques impliquent que le magmatisme puisse avoir lieu dans des environnements plus ou moins extensifs et extrêmement variés. Le contexte géotectonique final proposé pour les principaux types d'ophiolites sera d'autant plus valable que l'on tiendra compte des informations livrées par l'étude des séries volcaniques et/ou sédimentaires associées à la marge (2ème partie). Dans tous les cas, si les divergences entre les sutures semblent nettes, il ne faut pas oublier que chaque ceinture présente aussi des discontinuités longitudinales qui dérivent de processus particuliers jouant en début d'ouverture et de l'importance de l'extension ultérieure.

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Geochemical discrimination between ocean-floor and island-arc tholeiites—application to some ophiolites

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Two discriminative diagrams are proposed to separate island-arc tholeiites (IAT) and ocean-floor tholeiites (OFT). The first diagram, Ti/Cr vs. Ni, has been drawn using 84 island-arc (IAT) and 178 ocean-floor (OFT) samples with silica contents between 40 and 56%. About 97% of OFT and 93% of IAT samples fall, respectively, on opposite sides of the empirical boundary. In the second diagram, where the Ba/Y is less than 4.4 for the OFT and more than 3.9 for the IAT, the overlap between the two groups is about 6%.

Owing to alteration effects, only the discrimination diagram Ti/Cr vs. Ni has been applied to ophiolitic basalts from the Mediterranean belts, Newfoundland, Central and North America, and Mongolia. The effusive and hypabyssal formations plotting either in one group or in the other lead to the suggestion that they have been formed in several possible geotectonical environments. It appears that ophiolites generated in a mid-oceanic ridge are scarce in opposition to those formed in an island-arc setting. In this latter case, ophiolite associations may correspond to the juxtaposition of either island-arc – marginal basin or island-arc – offshore oceanic crust formations.

Deux diagrammes de discrimination sont proposés pour séparer les tholéïtes d'arc (IAT) de celles de la croûte océanique (OFT). Le premier diagramme Ti/Cr vs. Ni a été établi à partir de 84 roches des arcs et de 178 échantillons des rides médio-océaniques dont les teneurs en silice sont comprises entre 40 et 56%. Environ 97% et 93% des laves des OFT et des IAT tombent

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respectivement de part et d'autre d'une limite empirique. Dans le deuxième diagramme où sont reportées les valeurs du rapport Ba/Y: inférieures à 4.4 pour les OFT et supérieures à 3.9 pour les IAT, le recouvrement des deux groupes concerne 6% des échantillons.

A cause de l'altération, seul le diagramme de discrimination Ti/Cr vs. Ni a été appliqué aux basaltes des ophiolites de Méditerranée, de Terre-Neuve, de la cordillère de l'Amérique centrale et du nord, et de Mongolie. Les formations effusives et hypabyssales tombant dans l'un ou l'autre domaine laissent suggérer plusieurs environnements géotectoniques possibles pour leurs formations. Il apparaît que les ophiolites formées dans une dorsale de type médio-océanique sont rares à l'opposé de celles créées dans un environnement d'arc insulaire. Dans ce dernier cas, elles peuvent correspondre à la juxtaposition de formations issues soit de bassins marginaux et d'arcs, soit de croûte océanique "avant-arc" et d'arcs.

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Introduction

During the last few years, many attempts, based on geochemical parameters, have been carried out in attempting to identify the serial character of basic volcanic rocks (Floyd and Winchester 1975; Winchester and Floyd 1976) and (or) their original tectonic setting (Pearce and Cann 1971, 1973; Pearce 1975, 1976). However, as far as tholeiitic basalts are concerned, serious difficulties continue to exist in discriminating between "ocean-floor" and "island-arc" tholeiites (Jakes and White 1972; Hart *et al.* 1972; Pearce and Cann 1973; Gill 1976). Generally, the island-arc tholeiites have (sometimes drastically) lower contents of Ti, Cr, Ni, Zr, Y, Mg, and Fe and higher Ba, K, Rb, Cs, Pb, and Sr (Jakes and Gill 1970; Gorton 1977), but the chemical variations resulting from differentiation processes prevent any simple discrimination based on the absolute values of these elements.

In this paper we propose new discriminant diagrams for distinguishing ocean-floor and island-arc tholeiites based on the distribution of Ti, Ni, Cr, Ba, and Y.

Furthermore, the results obtained are applied to test the original tectonic setting of some well-known ophiolitic complexes. The diversity of ophiolites, according to the geochemistry of their lavas or the petrography of their cumulates (Allègre *et al.* 1973; Mésorian 1973; Pearce 1975; Church 1977; Church and Riccio 1977; Parrot and Ricou 1976) has lead to several classifications (Miyashiro 1975*a*, 1977; Rocci *et al.* 1975). The differences observed may be assigned to geophysical variations in a single setting (Church and Riccio 1977; Parrot and Ricou 1976) or to various tectonic environments as follows: (1) oceanic environment: mid-oceanic ridges, aseismic ridges, etc., (2) marginal basins or inter-arc basins, and (3) immature island arcs.

The first two geotectonical settings are more or less accepted, the last has been much more controversial (cf. Miyashiro (1973, 1975*b*,*c*) and the successive replies (Gass *et al.* 1975; Hynes 1975; Moores 1975)). Nevertheless, it appears that differences of petrography and geochemistry between ophiolites are more and more assigned to various geotectonic settings (Ohnenstetter and Ohnenstetter 1975; Beccaluva *et al.* 1977; Ohnenstetter *et al.*, in press; Sun and Nesbitt 1978).

Discrimination between Ocean-floor and Island-arc Tholeiites

Selection of Data

A total of 262 analyses, taken from the recent literature, were classified into two main a priori groups: (1) ocean-floor tholeiites (OFT, 178 analyses) from present-day mid-oceanic ridge environments and (2) island-arc tholeiites (IAT, 84 analyses) from present-day volcanic island-arcs. The data were selected as follows:

Only *tholeiitic* volcanic rocks with SiO_2 between 40.0 and 56.0% were used. This compositional range includes the picritic and andesitic end members of the basaltic clan. The OFT group includes normal ridge basalts as well as tholeiites probably slightly enriched in incompatible elements (cf. Schilling 1975), but volcanics referred to by authors as plume-ridge segments, and hot spots (i.e., oceanic islands) were excluded.

Analyses of samples with strong evidence of accumulation phenomena as well as of severe alteration (like abnormally high loss on ignition) were disregarded.

Analyses carried out with semiquantitative methods or of doubtful quality were disregarded.

A list of references and localities of the analyses from the literature is reported in Table 1.

Ti/Cr vs. Ni Plot

A discriminant diagram based on Ti and Cr contents has already been shown useful for distinguishing between ocean-floor basalts and low-potassium tholeiites of island arcs (Pearce 1975). A new discrimination diagram is proposed by consid-

TABLE 1. Localities and references of analyses

Location	Reference
Ocean-floor tholeiites	
(1) Mid.-Atlantic ridge, DSDP Leg. 37	Baragar *et al.* (1977)
(2) Mid.-Atlantic ridge, DSDP Leg. 37	Flower *et al.* (1977)
(3) Mozambique ridge, DSDP Leg. 25	Erlank and Reid (1974)
(4) Southeast Indian Ocean, DSDP Leg. 28	Ford (1975)
(5) Pacific Ocean, DSDP Leg. 32	Marshall (1975)
(6) Central Indian ridge	Engel and Fisher (1975)
(7) Indian Ocean, DSDP Leg. 26	Kempe (1974)
(8) Rift valley, Atlantic Ocean, 36°50′ N	Bougault and Hekinian (1974)
(9) Nazca plate, Leg. 34	Rhodes *et al.* (1976)
(10) Mid.-Atlantic ridge, DSDP Leg. 37	Lambert and Holland (1977)
(11) Mid.-Atlantic ridge, DSDP Legs. 2 and 3	Frey *et al.* (1974)
Island-arc tholeiites	
(12) Tonga	Ewart *et al.* (1973)
(13) Viti Levu, Fiji	Gill (1970)
(14) Lau–Colville	Gill (1976)
(15) Lesser Antilles	Brown *et al.* (1977)
(16) New Hebrides	Gorton (1977)
(17) South Sandwich Islands	Hawkesworth *et al.* (1977)
(18) Tonga–Kermadec	Ewart *et al.* (1977)
(19) Mariana trench	Dietrich *et al.* (1978)

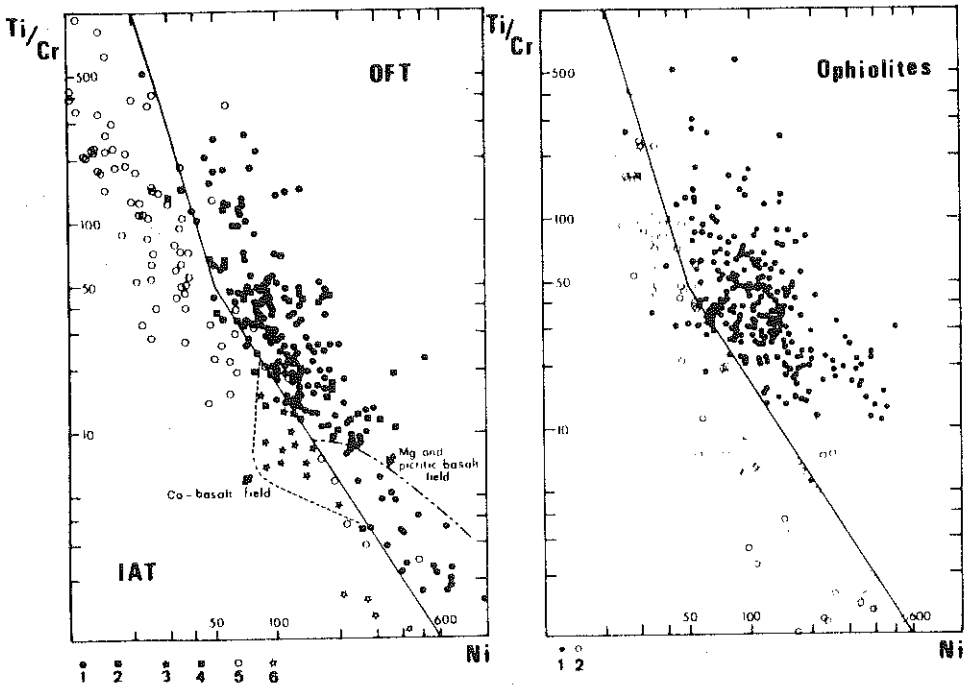


FIG. 1. Ti/Cr versus Ni diagram. On the left, are plotted the present-day basaltic rocks (40% < SiO_2 < 56%) of tholeiitic affinities coming from island-arc and oceanic environments. One hundred and seventy-eight analyses of ocean-floor tholeiites (1) and 84 analyses of island-arc tholeiites (5), have been used to draw the empirical boundary between the two types. Special attention has been paid to the Ca basalts from Leg 37 (3), and to basalts from marginal basins (2), fore-arc lavas (4), and boninites (6). On the right, basaltic and metabasaltic rocks from different ophiolites have been split into two groups. For each ophiolite, a geotectonic environment has been assumed after discrimination (see Fig. 3). Ninety-six percent of the oceanic-type ophiolites (1) are well discriminated, as well as up to 92% of the island-arc ophiolites (2).

ering Ti, Cr, and Ni, the Ti/Cr ratio being lower in IAT at a given Ni content. Ocean-floor tholeiites and IAT in the Ti/Cr vs. Ni diagram of Fig. 1, are clearly separated by an empirical boundary. About 97% of OFT and 93% of IAT samples fall, respectively, in distinct fields. Owing to the scarcity of data, there is a higher degree of uncertainty in the lower part of the boundary (i.e., that regarding the picritic rocks). In such a diagram, the trend of fractional crystallization is parallel to the boundary and the more differentiated rocks are displaced towards the upper left with increase of Ti/Cr and Ni decrease according to tholeiitic fractionation trends. High-calcium oceanic basalts (CaO > 14%) defined in the DSDP 37 (Deep Sea Drilling Project) (especially in holes 332 A and B, and 333 holes; Flower *et al.* 1977; Lambert and Holland 1977) and lavas from marginal basins and from fore-arc are also reported.

All the lime-rich basalts fall in the IAT field. As noted by Lambert and Holland (1977), their composition can not be solely explained by accumulative phenomena (mainly plagioclase), but is also indicative of peculiar magmatic characteristics, different from associated typical OFT. The efficiency of the discrimination is, however, maintained owing to the restricted field of the Ca basalts which cluster around an IAT minimum distribution. Most of the lavas from the marginal basins (Hart *et al.* 1972; Hawkins 1976; Dietrich *et al.* 1978), although scattered, fall in the OFT field. The fore-arc tholeiites, dredged from the nearshore Mariana trench (Dietrich *et al.* 1978), plot along the boundary in the two fields.

In conclusion, the proposed diagram allows one to discriminate between island-arc tholeiites on the one hand and ocean-floor tholeiites from mid-oceanic ridges and marginal basins on the other. Thus, this diagram may be usefully applied to the tholeiitic basalts of ophiolites since the elements involved in this diagram have been postulated as being relatively insensitive to secondary mobilization (Pearce 1975).

Ba/Y Plot

The usefulness of Ba as an indicator of island-arc or ocean-floor tholeiitic affinity has been remarked by Pearce and Cann (1973) and Hart *et al.* (1972), who also state that the K/Ba ratios are invariably less than 30–40 in island-arc tholeiites.

Figure 2 shows the distribution of Ba/Y ratios in OFT and IAT, which generally have values less than 4.4 and more than 3.9 respectively. The overlap between the two groups of rocks is very small

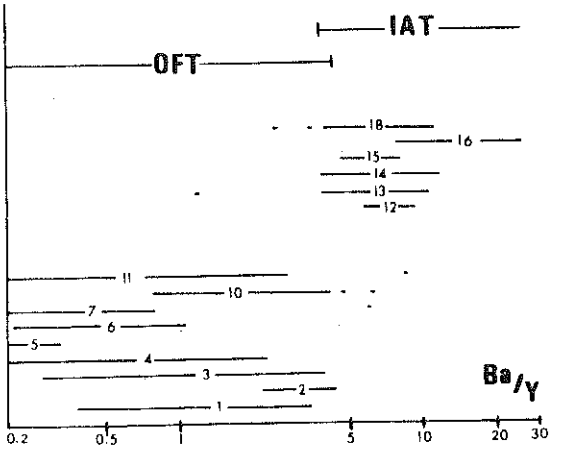


FIG. 2. Ba/Y ratio distribution in ocean-floor tholeiites (OFT) and island-arc tholeiites (IAT). In general, the Ba/Y ratios are less than 4.4 for OFT and more than 3.9 for IAT. Numbers correspond to those listed in Table 1.

and involves only about 6% of the samples. However, it must be noted that in ridge rocks, large-ion-lithophile enriched tholeiites present higher values of barium which would cause an overlap over the IAT range. The mobility of Ba during secondary processes limits the use of this diagram for ophiolites.

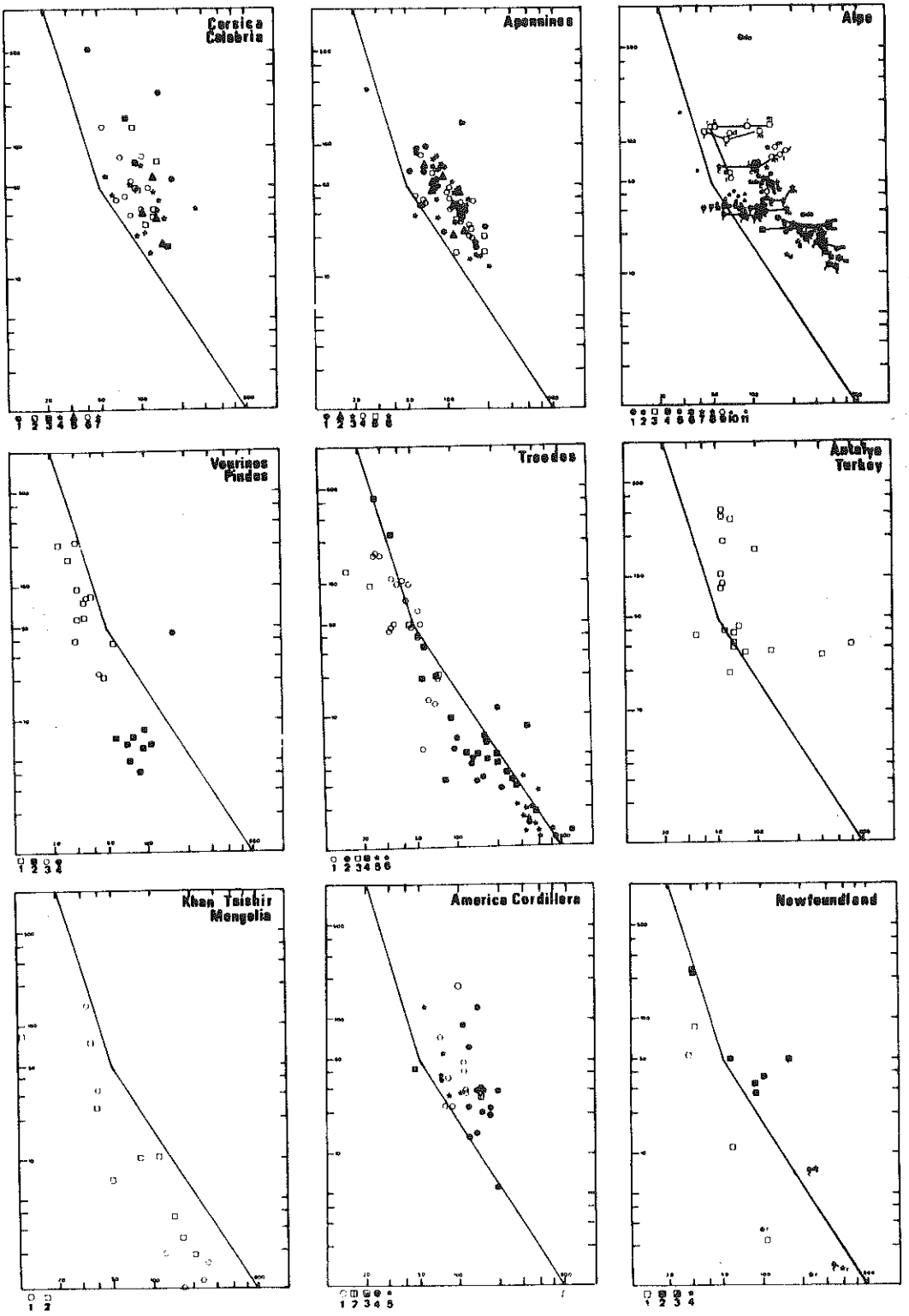
Application of Ti/Cr vs. Ni Diagram to Ophiolites

Analyses of ophiolitic basalts from the Tethyan belt, the Appalachian system, the Central and North America Cordillera, and the western Mongolia complex, are plotted on the Ti/Cr vs. Ni diagram of Fig. 3. The rocks in question occupy the two previously defined fields (Fig. 1). Basalts from each lava group generally plot in a single field, thus emphasizing the validity of the discriminative boundary. Despite the scarcity of analyses in some groups, most of the ophiolite analyses plot in the IAT field, thus confirming the observation that ophiolites with pure oceanic character are scarce, especially in the Tethyan belt (Cann 1976; Rocci *et al.* 1976).

Ophiolites in the Ocean-floor Tholeiite Field

The ophiolites from the western Alpine system (Corsica, Alps, Apennines, Calabria) and from Central and North America occupy the OFT field.

In the western Alpine system, the Ti/Cr and Ni variations may be related either to primary magmatic features and (or) to fractional crystallization. Basalts from Balagne, Corsica (Baud 1975; Gruppo di Lavoro sulle Ofioliti Mediterranee (GLOM) 1977), and the Mt. Maggiorasca, Apennines (Bec-


Corsica Calabria
Apennines
Alps
Vosgines Pindos
Troodos
Antalya Turkey
Khan Tsishir Mongolia
America Cordillera
Newfoundland

caluva *et al.* 1976; Ohnenstetter and Beccaluva, unpublished data), despite fractionation, show the highest initial Ti/Cr content. The hypabyssal rocks from eastern Corsica (GLOM 1977; Beccaluva *et al.* 1977) show the most evolved ferrobasaltic compositions. The variations of Ti/Cr and Ni of the lavas from the French Alps (Bertrand 1970; Mevel 1975) appear to be closely related to their respective geographic locations. Alteration from core to matrix in pillow lavas mainly affects the Ni distribution: the matrix is generally enriched in nickel and the rim is often depleted, resulting in a right to left alteration trend in Figs. 1 and 3 which could shift lavas across the boundaries and introduce errors in characterization. This trend may be clearly observed for the Get lavas whose pillows plot in the IAT field and the cores near the boundary. Four means of metabasalts from the eastern Alps, Austria (Raith *et al.* 1977), plot in the centre of the distribution for the western Alpine system. These results emphasize the lack of geochemical evidence of island-arc magmatism among ophiolitic basalts from the Alps, Apennines, Corsica, and Calabria.

In the Antalya ophiolites, Turkey (Juteau 1975), dikes cutting the harzburgites and gabbros fall in the OFT field.[1]

Tholeiitic lavas from Central and North American ophiolites: Preston Peak, California, western Ecuador, northwestern Colombia, and southeastern Panama (Goosens *et al.* 1977; Snoke *et al.* 1977) plot in the OFT field except for a sample from the upper basic igneous complex of Ecuador. In Ecuador, the development of IAT and calc-alkalic magmatism (Goosens *et al.* 1977) after OFT suggests an evolution of an island arc above oceanic crust. Ocean-floor tholeiite lavas of the Preston Peak ophiolites probably formed in a basin where independent island-arc magmatism developed as for several other ophiolites of the Klamath Mountains (Snoke *et al.* 1977).

In the western America Cordillera, the general association of OFT lavas with IAT and (or) calc-alkalic rocks is in favour of the generation of oceanic crust in marginal basins near the continental margin.

Ophiolites in the Island-arc Tholeiite Field

Most of the eastern Tethyan, the Newfoundland ophiolites, and those from the Khan-Taishir complex of western Mongolia plot in the IAT field.

About 83% of the Troodos basalts (Smewing 1975; Desmet 1977; Simonian and Gass 1978) fall in the IAT field. They show a wide variation in Ti/Cr ratios, which decrease from the so-called axis sequence towards the upper pillow lavas reaching the boninite field. The Arakapas fault basalts (Simonian and Gass 1978) fall in the boninite area. Similarly, the Vourinos lavas from the Krapa and Asprokambo areas (Paupy 1976; Beccaluva *et al.*, in

[1]The data about the dikes of Antalya have been obtained with semiquantitative methods.

FIG. 3. Ti/Cr versus Ni diagrams for different ophiolitic complexes.

Corsica: Dolerites (1) and pillow lavas (4) from Inzecca massif; basaltic dikes from Monte Piano Maggiore (3) (after Beccaluva *et al.* 1977). Prasinites (2), from eastern Corsica, pillow lavas from Balagne (6) (after GLOM 1977); dolerites from Balagne (5) (after Baud 1975).

Calabria: Metabasalts (7) (after Dostal *et al.* 1979).

Apennines: Pillow lavas of Monte Penna (1), Monte Quatese (2), and Monte Maggiorasca (3) (after Beccaluva *et al.* 1976; Ohnenstetter and Beccaluva, unpublished data). Pillow lavas from Levanto, Aulla, Crevole, Vallerano, Roccafederighi, Rio nell'Elba, and Capo Stella (4); basaltic dikes from Impruneta, Rosignano, and Fiume Farma (5); massive lava flows from Val Graveglia, Bargone, Levanto, Fiume Pavone, and Rio nell'Elba (6) (after Ferrara *et al.* 1976).

Alps: Pillow and dolerite from the Deux Pelvats (1) and (8), pillows and a dike from the Gets (2), pillows and a dike from Roche Noire (3), pillows from Ubaye – Roche Noire (4), pillow lavas and a massive flow from Péas (5), pillow lavas from Combe Bremond (6), pillow lavas and dikes from Chenaillet (7), pillow lavas and dikes from Pic Marcel (9) (after Mevel 1975).

Means of metabasites of the lower and upper Schiefferhülle (10) (after Raith *et al.* 1977), and diabases from the Gets (11) (after Bertrand 1970). The symbols are as follows: d = dike, do = dolerite, c = core of pillow, i = intermediate zone of pillow, r = rim of pillow, m = intra-pillow matrix.

Vourinos: Pillow lavas, massive lava flows, sills, and dikes of Krapa (1) and dikes from Asprokambo (2) (after Beccaluva *et al.*, in press).

Pindos: Spilitic lavas (3) and a dolerite (4) (after Montigny *et al.* 1973; Montigny 1975).

Troodos: Pillow lavas, prismatic lavas, and dikes from the basal group and lower pillow lavas (1), and upper pillow lavas (2) (after Desmet 1977); the "axis sequence" (3) and upper pillow lavas (4) (after Smewing 1975); the "axis sequence" (5) and upper pillow lavas (6) from the Arakapas fault (Simonian and Gass 1978).

Turkey: Tholeiitic dikes cutting the Antalya ophiolitic complexes (after Juteau 1975).

Mongolia (Khan-Taishir): Pillow lavas (1) and dikes (2) (after Zonenshain and Kuzmin 1978).

America Cordillera: Lavas from Panama (1), Ecuador lower unit (2) and upper unit (3), Colombia (4) (after Goosens *et al.* 1977); diabases from Preston Peak ophiolite (5) (after Snoke *et al.* 1977).

Newfoundland: Lavas from Stag Island, intrusive member (1), lower basaltic group (2), and upper basaltic group (3) (after Kean and Strong 1975). Betts Cove lower pillow lava member (4) (after Coish 1977). C = core and r = rim of the pillows.

press) plot in the previous two different fields, the Asprokambo rocks corresponding to the Troodos upper series.

In all the above mentioned sequences, a bimodal distribution of Ti/Cr ratios may be observed which suggests at least two different magma pulses.

A dolerite from the Pindos (Montigny *et al.* 1973; Montigny 1975) plots in the OFT field (Bébien *et al.* 1975), whereas two spilites have an IAT character. This distribution improves the existence of two different ophiolite associations in this area (Paupy 1976).

In Mongolia, lavas from dikes cutting the gabbros, those belonging to the sheeted complex, and from the pillows (Zonenshain and Kuzmin 1978), mainly fall in the two previously noted IAT fields of the eastern Tethyan rocks.

Analyses of Newfoundland effusive tholeiitic rocks (Kean and Strong 1975; Coish 1977), of the Stag Island formation and the Betts Cove ophiolite, fall in the IAT field except for the Stag Island lower lavas which plot in the OFT field. The Betts Cove samples fall towards the lower part whereas the other rocks are spread out in the upper part of the diagram. A special occurrence exists for the OFT Stag Island lavas which occur between two IAT formations. The andesites of Long Island correspond to a later development of a mature island arc above an immature arc (Kean and Strong 1975) and perhaps above oceanic crust.

Conclusions

The Ti/Cr vs. Ni diagram and the Ba/Y ratio permit an efficient separation of present-day oceanic and island-arc tholeiites (from picrites up to basaltic andesites and ferrobasalts). The Ca-rich oceanic tholeiites occupy a peculiar position in the IAT field of the Ti/Cr vs. Ni diagram, which corresponds to a minimum in the frequency of the island-arc tholeiites.

Application of the Ti/Cr vs. Ni diagram to ophiolitic complexes shows that their constituent basalts may generally be clearly assigned to one or the other type. Ophiolites composed only of IAT basalts have probably been formed in or near an immature island arc. Among IAT, the effusive rocks which show a picritic character are also sometimes (Troodos), of younger age.

Ophiolitic basalts belonging to the OFT field may have been formed in two main settings: either in an ocean or in a marginal basin.

If these basalts are found with ophiolitic lavas of IAT character, this association would suggest an island-arc – marginal basin or island-arc – offshore oceanic crust environment. It is to be noted that some marginal basin basalts might plot also in the IAT field if they were being generated in a spreading centre above a subduction zone.

Consequently, the geotectonical environment of ophiolites may be proposed after considering the spatial and chronological relationships of the whole association. In this respect, it appeared that on-land ophiolites generated in mid-oceanic ridges are probably scarce in contrast to those developed in an island-arc – marginal basin system.

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GÉOLOGIE. — *Sur la signification des sutures ophiolitiques en fonction de leurs caractères structuraux*. Note (*) de **Maryse Ohnenstetter,** présentée par Georges Millot.

L'organisation interne des ophiolites a été envisagée dans trois systèmes de contraintes : en expansion, en compression et en décrochement. L'analyse des caractères structuraux des grands massifs d'ophiolites a permis de les classer en deux groupes principaux. Le premier qui appartient aux systèmes en extension, est celui des ophiolites issues des océans ou des bassins marginaux. Le second qui évoque les systèmes en décrochement est issu de zones coulissantes à l'intérieur et en bordure de plaques : marges actives et passives.

The internal geometry of ophiolites has been considered for three systems of stress: extension, compression and strike-slip. An analyis of the structural characters of important ophiolitic massifs allows their classification in two main groups. The first one belonging to extensional systems is that with ophiolites formed in oceans or marginal basins. The second one resembling strike-slip systems may come from wrench-faults inside plates and at their boundaries: active and passive margins.

Les ophiolites étant le plus souvent considérées comme des reliques de croûte océanique, il est implicitement admis que leur structure interne est compatible avec un système en expansion ([1], [2]). Cependant, on peut se demander si d'autres régimes tectoniques ont pu régner lors de leur création, compte tenu des données pétrologiques les plus récentes. Si tel est le cas, la position des marqueurs de la déformation, qui varie en fonction de l'orientation des contraintes principales, doit aider à définir la géométrie interne des ophiolites. Trois systèmes de contraintes peuvent être retenus, à partir desquels on peut établir des modèles correspondant respectivement à des zones en extension, compression et décrochement. Ces derniers confrontés aux structures connues de quelques ophiolites permettront de rediscuter la signification de certaines sutures de roches vertes.

I. Le système en extension, modèle A (*fig.* 1). — La contrainte principale σ_1 étant en position verticale, les dykes sont parallèles à l'axe de la ride, soit verticaux dans les plans de distension, soit plus ou moins inclinés dans les plans de cisaillement [1]. Les chambres magmatiques, de préférence remplies par les dykes verticaux, se développent au-dessus du manteau résiduel. Les diapirs de péridotites [3] s'élevant dans l'axe de la ride pourront être entraînés progressivement loin de celle-ci, par l'intermédiaire de failles plates qui résultent de l'ouverture du dièdre des fractures avec la profondeur. A l'écart de la ride, le modèle de Juteau et coll. [4] peut s'appliquer : la foliation subhorizontale des péridotites forme un angle faible avec le rubanement des cumulats sus-jacents et la linéation des minéraux voisine de la contrainte minimale σ_3 indique la direction d'extension, perpendiculaire à l'axe de la ride. Les blocs découpés par les failles normales peuvent tourner selon un axe parallèle à la ride. Ces mêmes failles permettent la remontée facile d'écailles de gabbros et de péridotites. Près de la dorsale, ces dernières pourront encore présenter des structures acquises dans le diapir non oblitérées par les cisaillements profonds, plats. De la superposition des dolérites et laves aux cumulats déposés sur les péridotites, on obtient une section ophiolitique complète. Si les ultramafites affleurent et sont recouvertes par des formations sédimentaires plus ou moins bréchiques liées au jeu des failles et par des laves, un log incomplet en résultera. Le système en expansion peut exister dans les zones d'accrétion des océans et de certains bassins marginaux.

II. Le système en compression, modèle B (*fig.* 2). — Dans ce système, la contrainte intermédiaire σ_2 est horizontale et les dykes sont subparallèles à la ceinture de roches vertes. Les plans de rupture fragile horizontaux et les fractures conjuguées inclinées ne favorisent pas la montée des magmas. Les chambres magmatiques sont recouvertes par des sills plutôt

que par un réseau de dykes. Certains cumulats peuvent être plissés. La déformation dans les diapirs de péridotites est soulignée par une foliation subparallèle à la chaîne et une linéation verticale. Les accidents subplats participent au cisaillement des péridotites en détruisant les structures du diapir, tandis que les failles antithétiques permettent leur remontée. La structure des péridotites cisaillées sera voisine de celle obtenue à l'écart des dorsales dans le manteau résiduel entraîné. Les failles plates, chevauchantes, provoqueront le redoublement des séries ophiolitiques localement recoupées par des filons [5]. Celles-ci seront donc variables latéralement et verticalement. Le système en compression peut opérer à proximité des marges actives et être associé à des mouvements coulissants.

III. Le système en décrochement, modèle C (*fig.* 3). — Dans ce système qui présente, dans un même plan horizontal, des zones, alternativement en compression et en extension, la contrainte maximale σ_1 est légèrement oblique par rapport à la direction générale de la ceinture ophiolitique. σ_2 étant verticale, les fractures le sont aussi, subparallèles à σ_1 ou en position conjuguée, ce qui facilite la montée des magmas. Les chambres magmatiques situées dans les zones en distension sont de taille réduite, car limitées par les fractures abondantes. Dans ces dernières, des gabbros peuvent être formés par fluage. Les péridotites présentent une foliation subverticale, comme on peut s'y attendre dans les diapirs et les accidents coulissants. La linéation est verticale à inclinée, en fonction du mouvement du diapir dans le décrochement. En conséquence, le rubanement des cumulats formés par gravité présente un angle important avec la foliation des péridotites et les épontes des filons. Les cisaillements verticaux jouant pendant et après la formation des ophiolites interrompent ces éléments géométriques repérables et provoquent leur rotation. La ceinture de roches vertes qui en résulte est constituée d'une juxtaposition de sections ophiolitiques lenticulaires, perturbées voire incomplètes : laves reposant sur des péridotites, gabbros intrusifs dans les péridotites, les filons ou les laves. L'allure générale de la chaîne résultera du jeu préférentiel d'un accident cisaillant. Le système en décrochement peut jouer dans de nombreux contextes géotectoniques à l'intérieur des plaques et à leur périphérie : faille transformante des océans (décrochement en distension), zones coulissantes à proximité des arcs insulaires ou limitant des microplaques continentales (décrochement en compression).

IV. Essai d'une classification structurale des ophiolites. — Si on cherche à trouver des exemples parmi les ophiolites, qui soient susceptibles d'illustrer les modèles précédents, il est nécessaire de regrouper les massifs aux caractéristiques structurales voisines. Bien que les données soient encore peu abondantes et fragmentaires, une classification des ophiolites en deux groupes peut être proposée. Un premier ensemble constitué par les massifs de Nouvelle-Calédonie, Papouasie, Antalya, Oman et Bay of Islands présente des sections régulières aux

Explication de la Planche

Légende commune à toutes les figures sur la mise en place primaire des ophiolites : p, péridotites; c, cumulat; b, bordure figée; d, dyke; s, sill; l, lave; f, fracture.

Fig. 1. — Système en extension. Coupe perpendiculaire à la ceinture de roches vertes.

Fig. 2. — Système en compression. Coupe perpendiculaire à la ceinture ophiolitique.

Fig. 3. — Système en décrochement. Coupe parallèle à la ceinture de roches vertes.

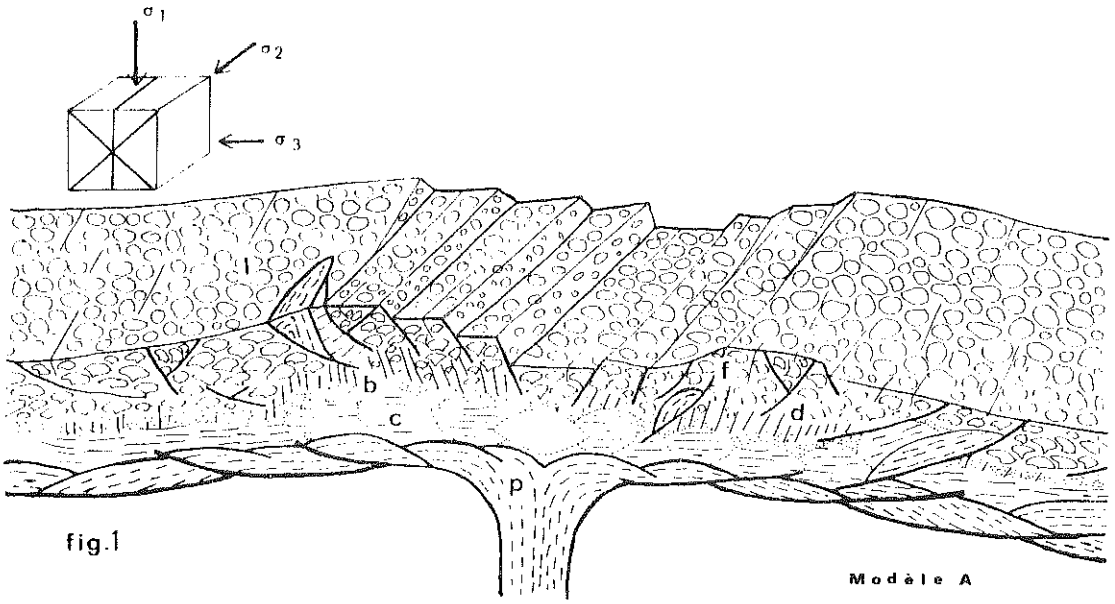

fig.1

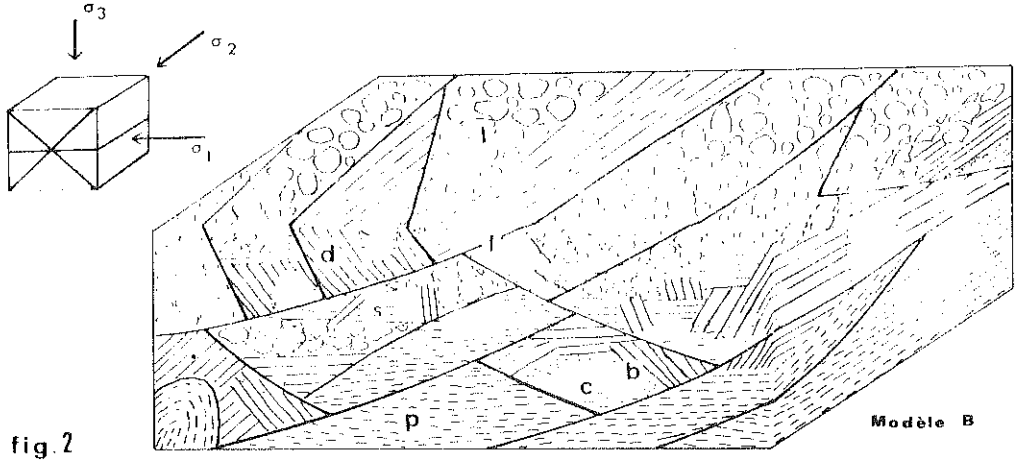

fig.2

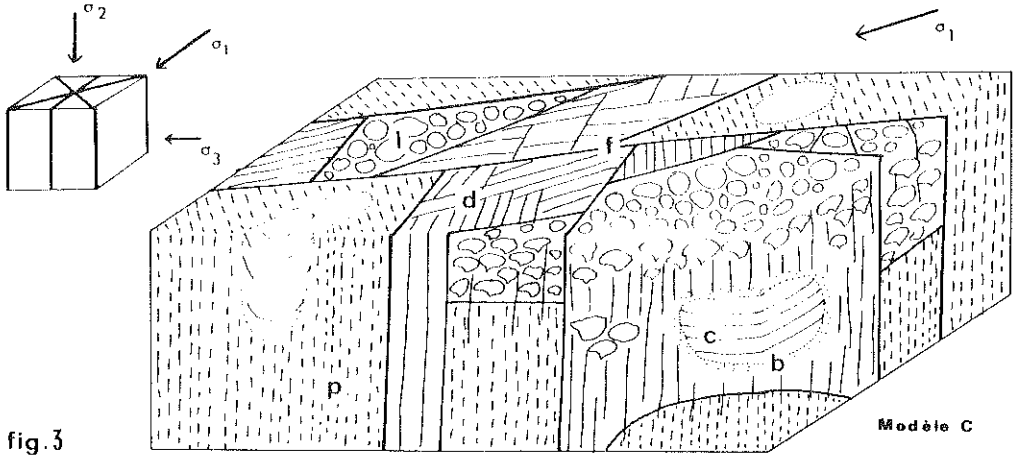

fig.3

variations lithologiques subhorizontales dans les péridotites et les cumulats ([4], [6]). Ce dispositif ainsi que celui des filons sont compatibles avec le modèle A.

De très nombreuses ophiolites, parmi les mieux connues, peuvent être rassemblées en un deuxième groupe. En particulier, la foliation des péridotites y est généralement verticale et variable en direction, comme dans les massifs du Troodos ([1], [7]), du Vourinos [1], de Sabzevar [8], d'Acoje, Philippines [9], de Canyon Mountain [3] et de la plupart des ophiolites de la cordillère occidentale de l'Amérique du Nord [10]. Les cumulats sont peu abondants [8]. Les filons d'orientation variable ([3], [7], [8]) occupent quelquefois des plans conjugués liés à un phénomène de décrochement [11]. La structure de ces ophiolites est compatible avec le modèle C.

V. Signification des sutures ophiolitiques. — En confrontant les trois modèles respectivement A, B et C avec la structure interne des deux groupes d'ophiolites, il est apparu qu'aucun massif, pour l'instant, n'est caractéristique du modèle en compression (B). Le premier groupe d'ophiolites qui comporte les plus grands massifs s'est formé dans les zones en extension, modèle A, des océans ou des bassins marginaux localement interrompues par l'activité des failles transformantes [12]. Au cours du transport des ophiolites, les péridotites basales acquièrent une structure de glissement intracristallin [13].

La plupart des autres massifs, ceux du second groupe, sont issus de systèmes en décrochement (modèle C) situés dans des contextes géodynamiques variés. Il convient alors d'examiner cas par cas leur lieu de formation possible : arc insulaire, intérieur des cratons, marges actives et passives. En domaine continental, les décrochements participent à l'amincissement de la croûte sialique avant l'apparition des roches vertes et peuvent encore jouer un rôle lors de la formation des nappes ophiolitiques [14]. Ces ophiolites ensialiques se situeraient à proximité de leur lieu d'origine : les phénomènes d'obduction ou de subduction ayant été limités pendant leur mise en place secondaire [15]. Dans ce cas, les péridotites peuvent encore présenter des textures de diapirs. Dans l'état actuel de nos connaissances, en Méditerranée orientale, l'opposition des structures du Troodos, Vourinos et Sabzevar d'une part et d'Antalya et Oman d'autre part, conduit à imaginer plusieurs ceintures de roches vertes ce qui réduit d'autant la durée et la portée du chevauchement d'un quelconque paléo-océan téthysien.

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THE ROLE OF POSSIBLE TRANSVERSE FAULTS IN THE DEVELOPMENT OF AN IMMATURE ISLAND-ARC COMPLEX: EVIDENCE FROM THE SABZEVAR OPHIOLITES, NORTH-EAST IRAN.

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RESUME

Dans les ophiolites de Sabzevar, au Nord Est de l'Iran, des études structurales ont montré que l'accrétion avait lieu plutôt le long d'un axe d'extension oblique. Celui-ci a pu se développer soit à l'intersection d'une ride avec une zone transformante qui la décale de peu, soit dans des bassins de distensions losangiques alignés le long d'un décrochement sub-parallèle à la chaîne de montagne. L'attention a été particulièrement portée sur l'origine des cumulats issus probablement de plusieurs chambres magmatiques disloquées. La formation de dikes intrusifs, de cumulats sécants et de chenaux plus ou moins dissymétriques a été controllée par des zones de fractures jouant lors de l'évolution de la chambre magmatique. En conséquence le rubanement résulte, surtout vers la base et le sommet, de cristallisations le long de fractures fortement inclinées et à proximité du plancher temporaire sub-horizontal suite aux effets gravitaire.

ABSTRACT

In the Sabzevar ophiolites, North-East of Iran, structural studies have shown that accretion has rather occured along an oblique spreading axis. A discontinuous ridge may have developed either at a ridge-transform intersection with a small offset or in a pull-apart basin linked to a wrench fault sub-parallel to the mountain belt. Special attention was paid to the origin of the layered cumulates which probably were formed in several disrupted magma chambers. Dike intrusion, cross-cutting cumulates some of them being parallel to the margins of the magma chambers and more or less asymmetric troughs were controlled by fracture zones acting at the time of magma chamber evolution. Therefore the layering results, especially towards the base and summit, from crystallization along steeply inclined fractures and on the sub-horizontal floor due to gravity.

MOTS CLES

ophiolite - cumulat - décrochement - chambre magmatique - Iran

I INTRODUCTION

The formation of ophiolites is thought to result from a process of extrusion-intrusion in a Mid-oceanic ridge. Mechanisms of accretion have been largely modelled from the orientation of lithologic and structural planes in the ophiolitic members and sedimentary covers, that is from mantle to oceanic crust (MOORES and VINE,1971). More refined tectonic analysis has progressively lead to the assumption that structural variations exist inside the oceanic crust, possibly due to ridge-transform intersection (DEWEY and KIDD,1977;PRINZHOFFER et al., 1980; CASEY and KARSON, 1981) or to variable plate velocities (CANN,1974).

The greater importance of extension over strike-slip systems during the opening of an oceanic basin may also be responsible for the variable internal geometry observed in ophiolites. Two main types of structural assemblages have been recognized(OHNENSTETTER,1980). According to this classification, the Sabzevar ophiolitic massif, North-East of Iran, which shows relict diapiric textures in mantle peridotites, rare magma chambers and dikes of variable attitudes, may have originated in a limited basin linked to wrench faults. To restore the initial mechanisms of emplacement, additional information about the accretionary processes may be obtained from cumulates which occur between the mantle rocks and the dike complex. If cumulates crystallized mainly at the margin of a magma reservoir (MCBIRNEY and NOYES,1979) the layering may be used to model the shape and size of the crustal chambers and the path for magma migration. This process may explain the presence of sub-horizontal and sub-vertical layering in undisturbed stratiform complexes (ROOBOL,1972; MC CLAY and CAMPBELL, 1976; GOODE,1977; CAMPBELL,1978). In ophiolites, sub-vertical banding is recognized but its origin is still controversial. It may result from static crystallization at the margin (CASEY and KARSON,1981), flow differentiation (THAYER,1963) in a diapir (AVE LALLEMANT,1976). In some cases, they could be overturned during tectonic movements (DESMET et al., 1980). Whatever their origin they could have to be linked to sub-horizontal cumulates crystallized at variable distances from the

ridge axis (GREENBAUM,1972) from several magma imputs (JACKSON et al., 1975; JUTEAU and WHITECHURCH,1980 ;PALLISTER and HOPSON,1981).

The present study is an attempt to clarify the structural evolution of magma chambers and their feeder zones from several cumulate massifs scattered along the Sabzevar ophiolitic belt. Evidence is shown that the attitude of layering and the formation of troughs depend on the gravitational forces and the position of fracture zones. Flow differentiation occurs mainly at the end of the main magma chamber evolution. The relationships between mantle rocks, cumulates and sheeted dike complexes will be also given. The setting is more compatible with a creation of ophiolites along an oblique spreading axis developed either at a ridge transform intersection with a small offset or in a pull apart basin linked to a strike-slip zone parallel to the mountain belt.

II GEOLOGICAL SETTING.

According to the classification of LENSCH and DAVOUDZADEH (1981), the ophiolites from the East-West trending Sabzevar range belong to a Mesozoic belt bordering the northern zone of the Central-East Iran microplate (fig.1), where they are associated with mélanges (TAKIN,1972). On the basis of paleomagnetic data which indicate a similar present-day position of the northern part of Iran as in Cretaceous time (WENSINK and VARENKAMP,1980), the existence of a large Tethys in this area has been excluded (BAROZ et al., in press). The opening and closure of the limited ophiolitic basins were possibly triggered during lateral movements(STAMPFLI,1978; OHNENSTETTER,1980) which persist since a long time in the North Eastern Iran(MAJIDI,1980). The ophiolites of Coniacian age were later covered by a volcano-sedimentary unit during Campanian, Maestrichtian and Paleocene (LENSCH et al., 1977; BAROZ et al., in press) which belongs to the lower part of an arc sequence. From Maestrichtian time to Miocene various magmatisms succeed,some of them testifying of the mature stage of the arc building up (LENSCH et al., 1977). Tectonic and metamorphic events occur at the same time, the more important being after the deposition of the Upper Cretaceous volcano-detritic series (BAROZ et al.,in press). The ophiolitic complex exposed in the Southern part of the Sabzevar range (fig.1) is relatively complete. It is constitued of nearly 80 % mantle peridotite and about 20 % cumulates, hypabyssal rocks and rare lavas. The successive tectonics lead to highly scattered ophiolitic blocs, also found in the mélange zone and bounded with faults.

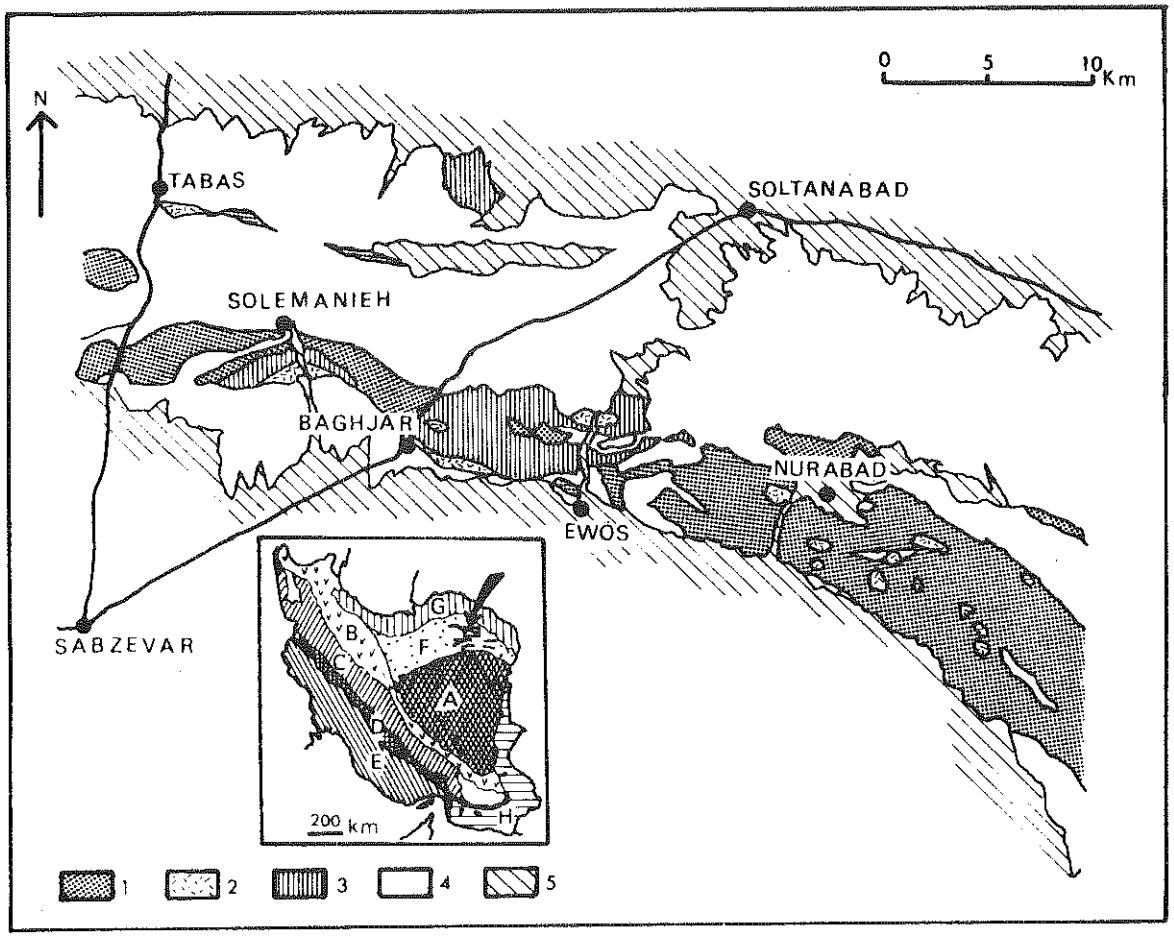

fig. 1 Geological map of the Central part of the Sabzevar belt modified after LENSCH et al. (1977) and BAROZ et al. (in press).
1) harzburgite, 2) intrusive rocks, 3) hypabyssal formation, 4) arc sequence, 5) quaternary
Sketch map: A) Central Iranian mass,B) Urmian Dokhtar zone, C) Sanandaj-Sirjan zone, D) Zagros thrust zone, E) Zagros folded belt, F) Kavir zone, G) Alborz mountains, H) Makran-East Iran, in black ophiolites, arrow indicates the Sabzevar area.

Geochemical data indicate that liquids exhibit close ressemblance with some immature island arc series (LENSCH et al.,1979). Later studies from the sheeted dike complex reveal that the composition of magma migrating in dike evolves with time (BAROZ et al., in press).

To reconstruct the ophiolitic sequence from puzzling blocs, general observations have been noted along the belt together with local finer studies in the Solemanieh area, and from cumulate massifs outcropping near Tabas and Nurabad (fig.1).

A- PERIDOTITES

Mantle peridotites are well exposed along the Sabzevar range, forming the highest summit. The attitude of the structural planes may reveal the direction of flowage in the mantle at least at the time of magma injection in the upper crust. In the Nurabad massif (see later) , the layering of the folded enclosing peridotites was subvertical at the time of the gabbroic injection (fig.8). This fold axis is variably inclined. Similarly in the huge peridotite massif composed of harzburgite and clinopyroxene (Cpx) -poor lherzolite from the Solemanieh area, layering and foliation are more often sub-vertical (NOGHREYAN,1982) except towards the edge of the massif where they are slightly inclined. Late open folding and thrusting led to a rotation of the formerly sub-vertical planes. As a rule, the highly fractured peridotites are locally fresh. The banding, one meter thick, is formed by repeated sequences of pyroxenite and dunite separated from harzburgite by sharp contact. Segregation of sub-vertical pyroxenites bordered by dunites are folded with the enclosing peridotites. In some places, pockets of dunites and chromitites occur. In thin sections, granoblastic textures are abundant with cuspate boundaries between minerals which indicate syntectonic recrystallization rather than intra-crystalline gliding (AVE LALLEMAND, 1976). The attitude of layered plane and foliation from peridotites which may testify to the persistance of a diapiric structure is characteristic of some ophiolitic massifs (OHNENSTETTER,1980; NICOLAS and VIOLETTE,1982).

B- THE SHEETED DIKE COMPEX

The upper part of magma chamber in ophiolites are often in stratigraphic contacts with sheeted dike complex. Informations about the chronology of magma emplacement, the nature of successive liquids and the direction of expansion may be deduced from the petrography and structural studies of dikes. In the Sabzevar range, the chronology of magma imputs has been mainly deduced from dike relations in the Baghjar and Solemanieh areas (BAROZ et al.,in press). Two maxima exist from the sheeted complex of Baghdjar: one at a small angle to the layering of the cumulate massif may be related to sill formation, the other is sub-perpendicular to the layering, like dikes cutting the upper cumulates (fig.2). Other dikes more often steeply inclined vary in strike.

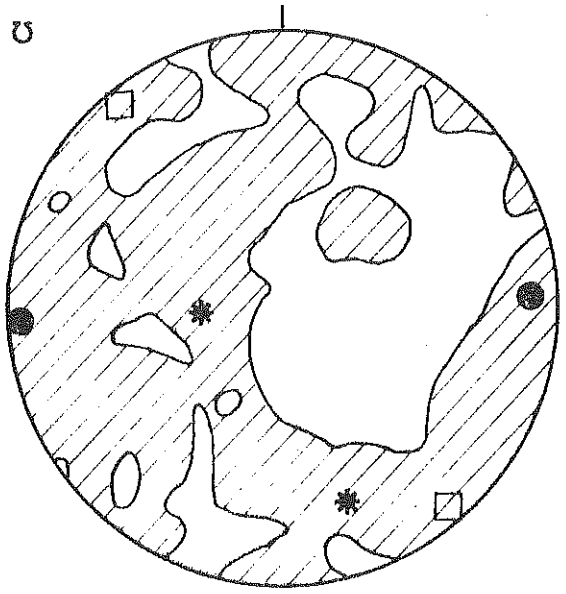

fig. 2 Attitude of the hypabyssal formations. Lower hemisphere. Equal area projection. Circle, dikes cutting the upper cumulates; square, dike from the Solemanieh hypabyssal formations; stars, dikes from the Baghdjar sheeted complex. The 0.6% contour interval is drawn after BAROZ et al.(in press).

The successive liquids may be linked to the evolution of the Solemanieh chamber which represents the main ophiolitic magmatic event. The least evolved liquids are SiO_2 saturated (Quartz tholeiites) like most of the Eastern Tethyan ophiolites (ROCCI et al., 1975). Their composition is compatible with an island arc environment (OHNENSTETTER and OHNENSTETTER,1976; BECCALUVA et al.,1979). In the Baghdjar area the late intrusion of very altered porphyritic andesitic basalt which bears some ressemblance with North-South dikes crossing the Solemanieh upper gabbros may testify that new magma batches poorer in Al and incompatible elements still arrive to the surface after the main ophiolitic intrusion (BAROZ et al.,in press). As in other ophiolitic complex, they indicate a progressive depletion of the mantle source (BECCALUVA et al., manuscript).

C- THE CUMULATE MASSIFS

Several small massifs of cumulates which correspond to a few percent of the ophiolitic association occur in the Sabzevar range. They are tectonically included in peridotites or sheeted dike complexes except the Tabas massif which is against Eocene rocks. The constitutional banding and structural features have been outlined in three massifs.

The first one, in the Solemanieh area is the more complete despite some disruption at the base and near the summit. The evolution of the layering, the crystallization order and the importance of the successive inputs may be estimated from its study. The second massif near Tabas belongs to an upper zone of a magma chamber. The nature of the layering would permit to point out the relative importance of static conditions of crystallization and gravitative features. The importance of fracture zones to control the layering attitude and the shape of the magma chamber will be also shown. Similarly, in the Nurabad massif, tectonic studies permit to follow magma migration but towards the lower edge of a chamber partly enclosed in peridotites. The comparison between all of these massifs favorizes the modelling of the evolution of chambers and the flowage of magma at some accretionary boundaries. Cumulate terminology (WAGER et al.,1960;WAGER and BROWN,1968) will be used to describe rocks without any implication on their origin.

III THE SOLEMANIEH MASSIF

This massif forms an elongate body with its southern border limited straighly by an important tectonized mélange zone. Between the "mélange" and the layered massif, from South to North, tectonic sheets of harzburgite, cumulate wherlite and pegmatitic gabbros, belong probably to the lower part of the magma chamber. Similarly towards North, the upper cumulitic sequence has been strongly faulted against the harzburgitic body (NOGHREYAN,1982). Based on textural and petrographical evidences (SADREDINI,1974; NOGHREYAN,1982), the cumulate massifs may be divided in three zones (fig. 3A).

Cumulates from the lower zone occur in tectonic sheets forming well bounded outcrops. There plagioclase (Pl) -poor rocks like wehrlite alternate with olivine (Ol) clinopyroxenite locally cut by dikes of pegmatitic gabbros. Sheets of coarse-grained melagabbros (Ol-rich gabbros and gabbronorites) occur between wehrlitic bodies and the middle zone of the cumulitic pile. Most of the lower rocks are coarse-grained adcumulates with grain size sometimes reaching the centimeter. Cumulus Ol and Cpx are partly enclosed in orthopyroxene (Opx) and Pl. No gravitational feature has been recorded in layers.

The transition between the lower cumulate and the middle zone is emphasized by the disappearance of Ol, the strong Pl and Opx increase and by a better layering in finer-grained cumulates. The layering

striking East-West and dipping towards North is marked by phase contact and mainly ratio and form contacts. Layers are wider towards the South reaching about 0.5 m and more reduced at the top giving a fine-scale layering (1-10 cm wide). Each layer is composed of melanocratic material at the bottom and leucocratic gabbro towards the top. From South to North , the decrease in thickness is also accompagnied by a decrease in pyroxene rich bands. Cumulus Pl and Opx define an igneous lamination. Their presence in postcumulus phase increases towards the top. Cumulus and postcumulus Cpx may occur in the same rocks. Therefore most of the noritic gabbros are mesocumulates and sometimes adcumulates especially at the base and orthocumulate at the top. Besides internal petrographic variations, pure gravitational features, well developed in troughs may also indicate the polarity of the magma chamber like in the Tabas massif (see later). The channels which modify the plane of mineral deposition may explain the repartition of the projecting lithological planes (fig.3B). The origin of troughs is partly related to tectonic movements (see later). Similarly the presence of near vertical faults later filled by gabbro is in agreement with the existence of tectonic movements during the formation of the cumulate pile. In one place, a reverse platy fault is marked by foliated gabbro injected along the fracturated planes probably as a crystal mush.

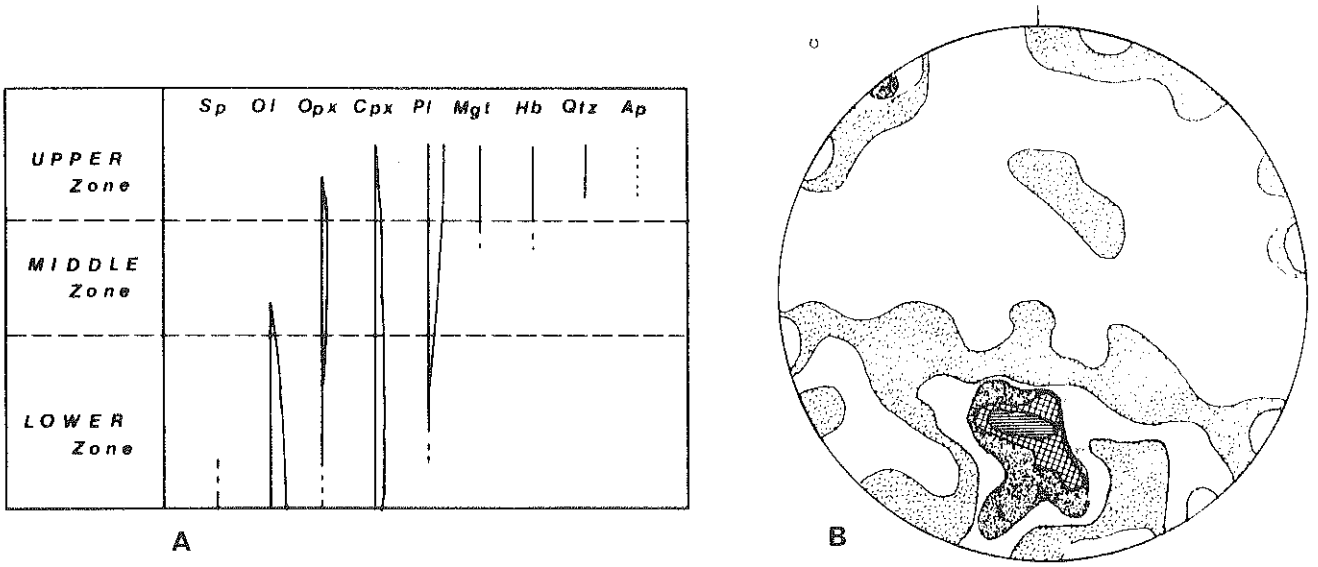

fig. 3 Cumulates from Solemanieh. A) crystallization order: Sp=spinel, Ol=olivine, Opx=orthopyroxene, Cpx=clinopyroxene, Pl=plagioclase, Mgt=magnetite, Am=amphibole,Qtz=quartz
B) Contoured diagram for the layering. Lower hemisphere.Equal area projection, contour intervals: 2.5%, 5%, 7%, 7.5%,12.5%.

Further to the North, in upper cumulates, the layering progressively vanishes as dioritic rocks occur. The last noritic gabbros (orthocumulates) and quartz (Qtz) - amphibole (Am) bearing gabbros have the smallest grain size. Some of them possess an intersertal to ophitic texture. Large euhedral branching hornblendes sometimes sub-perpendicular to the layering belong to a leucocratic gabbro which may invade the fine-grained gabbro-norite. These pegmatitic patches either with Cpx or Am may form a low angle with the layering or may cut abruptly the cumulates. Moreover, near vertical diabasic dykes trending North-South which cut the upper cumulates, increase towards the North and may reach locally high proportion (Fig.2).

Thus the cumulate massif of Solemanieh presents rather complete sequence despite strong faulting both at the bottom and top. The crystallization order is Sp-Ol-Cpx-Pl±Opx-Mgt-Am±Qtz (fig. 3A). The lower part of ultramafic composition with coarse grained adcumulates and locally heteradcumulates indicate rather equilibrium conditions of crystallization and injection of some magma pulses. In the middle part olivine-gabbro and gabbro-norite are often well layered. The absence of repetitive sequence indicate limited replenishment of the chamber. Such facts as well as the gradual decrease of the layers and the size of minerals may be explained by a reduction of the magmatic cells with time. An increase in the cooling rate is also suspected from the textural evolution of cumulates (ad- to orthocumulates). In the upper part of the chamber, some quartz-diorite and granophyre testify to local intrusion in the superficial effusive formations. The persistance of magmatic activity is indicated by dykes of basalt and basaltic andesite.

IV THE TABAS MASSIF

A - GEOLOGICAL SETTING

The Tabas gabbroic massif is an elongated body (1.5 Km long - 0.5 Km wide) tectonically included in post-ophiolitic andesitic series with slices of Nummulitic calcareous sediments, sometimes bounding the gabbros (fig.1). It is bordered by major sub-vertical faults, striking N 90 to N 140, which indicate important late horizontal displacements. Towards the contact, the dip of the layering in the gabbros gets nearly vertical. These faults are probably conjugated with minor N 40 to N 60 trending faults, better developped in the Western part of the massif. Serpentinites in the tectonic contact are transformed products after plagiferous pyroxenites. They contain numerous amygdules of gabbroic heteradcumulates.

The massif itself is highly fracturated with valley and crest parallel to the major faults. It is build up mainly of gabbro-norite like towards the summit of the Solemanieh massif. Locally they are associated with more or less Pl-rich pyroxenites, especially in the Northern part of the massif.

B - THE CONSTITUTIONAL BANDING

1. Evidence for static conditions of crystallization.

In this massif, especially towards its Southern part, uniform layering is much more developped than rhythmic layering. Irregular variations in grain size and proportion of minerals are responsible for the diffused boundaries of the tortuous, laterally limited, uniform layering. Some pegmatitic patches alter the general isotropic medium-grained rocks. Towards North, the increase of banded zones, one to three meters thick which succeed to uniform gabbros permits to precise the main characters of the mineral accumulation. When rhythmic layering exists, form contacts are numerous besides better developed grain size and proportion contacts. In the following described section, the order of the successive layers does not constitute the rule. In fig. 4 A, from West to East, medium-grained gabbro and Pl-rich mesocumulate (1) alternate.

Contacts are clear between layers of variable width in which minerals have no preferred orientation. A zone of gabbroic crescumulates with stout, branching Cpx getting wider and wider towards East indicate the top of the magma chamber towards North -East (2). They are followed by Pl-rich orthocumulates containing xenocrysts from below (3). Grain size increases regularly from base to top. A coarse-grained noritic gabbro with an ortho to mesocumulate texture succeeds to the Pl-rich cumulates (4). If rythms are to be defined in this banded rocks, they would begin with medium to coarse grained Pl-orthocumulate or isotropic gabbro and end with mesocumulates or crescumulates. Everywhere the branching Cpx indicates a similar growth direction except in one case where it is opposite (5). Therefore it may be considered that growth is generally initiated from the temporary floor but in some case branching crystallization may start from a chain of minerals suspended in the magma near the top of cumulitic pile.

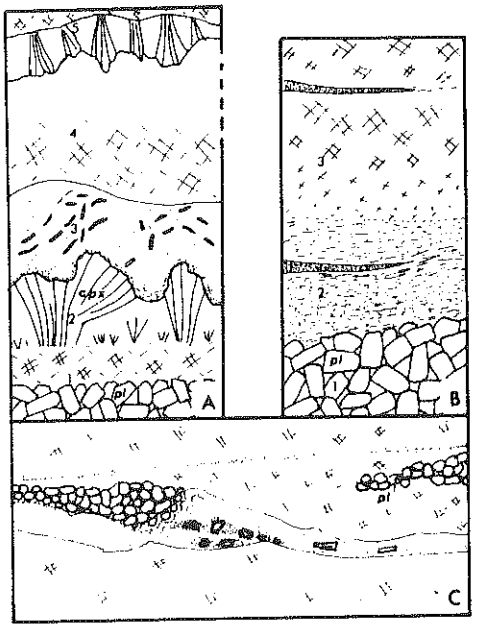

fig. 4 Aspects of the layering in the Tabas massif. A) contact of form, proportion and size from uniform to rhythmic layering (see text); B) interruption of layers by current action (see text); C) sliding and warping of layers. Small points, fine grain cumulates often laminated; crosses, isotropic gabbro of variable size. Pl=plagioclas Cpx=clinopyroxene.

Another section (fig. 4 B) shows from base to top the succession of Pl-rich orthocumulate (1), oriented fine grained gabbro (2) cut by a more laminated one and gabbroic mesocumulate with an isotropic texture marked (3) by Cpx, the size of which increases towards the top. Elsewhere, in the decussate isotropic gabbro a stage of progressive reducing grain size may be defined after an increasing one. Finer-grain Opx-rich gabbro-norite with acicular Opx interrupts this gabbro. They are later followed by noritic gabbro of variable grain size and Pl content.

In pyroxenites from the North Western massif, proportional phase contact is mainly responsible for the layering in heteradcumulates. Pl-rich pyroxenites, clinopyroxenites, gabbro-norite and gabbro result from variable proportion of cumulus Cpx and heteradcumulus Pl.

Flank of channel

2. Trough structures

The previous descriptions have showned that the appearance of banded cumulates towards the top of isotropic gabbros are linked to better grain size and ratio contacts and to the development of form contacts. Inside the banded zones, static conditions of crystallization as evidenced by the wispy layering in crescumulates occur after the formation of mesocumulates. These conditions are sharply interrupted by current action and gravitative forces which lead to orthocumulates and laminated rocks. All of these processes are better observed than in troughs.

Trough structures with different width/depth ratio may be classified in two groups, those developed on ondulatory layers and those corresponding to erosional channels reaching several meters wide. Troughs formed by wispy layering have been locally favored by different growth advancements of minerals in crescumulates. Ondulatory layers marked by a melanocratic band are sometimes slightly dissymetric with gentled slopes and filled with layers wider towards the center. When the depth of troughs increases, the layers inside get more irregular. Towards the top grain size increases and the trough progressively vanishs. The banding becomes parallel to the general layering of the massif. Some Pl aligned in sub-vertical axial plane of symetric to slightly dissymetric troughs may indicate the relictual direction of flowage towards North-East. In an another place, Pl-rich orthocumulate succeeding to Cpx bearing crescumulate contains numerous xenocrysts having an orientation suggesting current flowing towards South from the Western border to the Eastern one (fig. 4A) Banding in well developped channel forms a high angle with the previous layered rocks which have more often a coarser grain size and wider layers

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fig. 5 Point diagram of some selected cumulates after rotation. Wulff net, lower hemisphere. A) Opx-rich gabbro-norite (full triangle) showing igneous lamination (open triangle), B) pegmatites of gabbro, C) pyroxenite (full triangle) and related gabbroic pegmatites (open triangle) , D) gabbro-norite flanking channel, each symbol indicates a different trough; LM=magmatic lineation ; poles from Western flank (full star) and Eastern flank (open star). In each projection girdle and its pole are drawn except for D) where poles of girdles are only reported for clarity. E) envelops of laminated gabbro-norite, pegmatitic gabbro, pyroxenite and gabbro flanking channel. Wulff net, lower hemisphere.

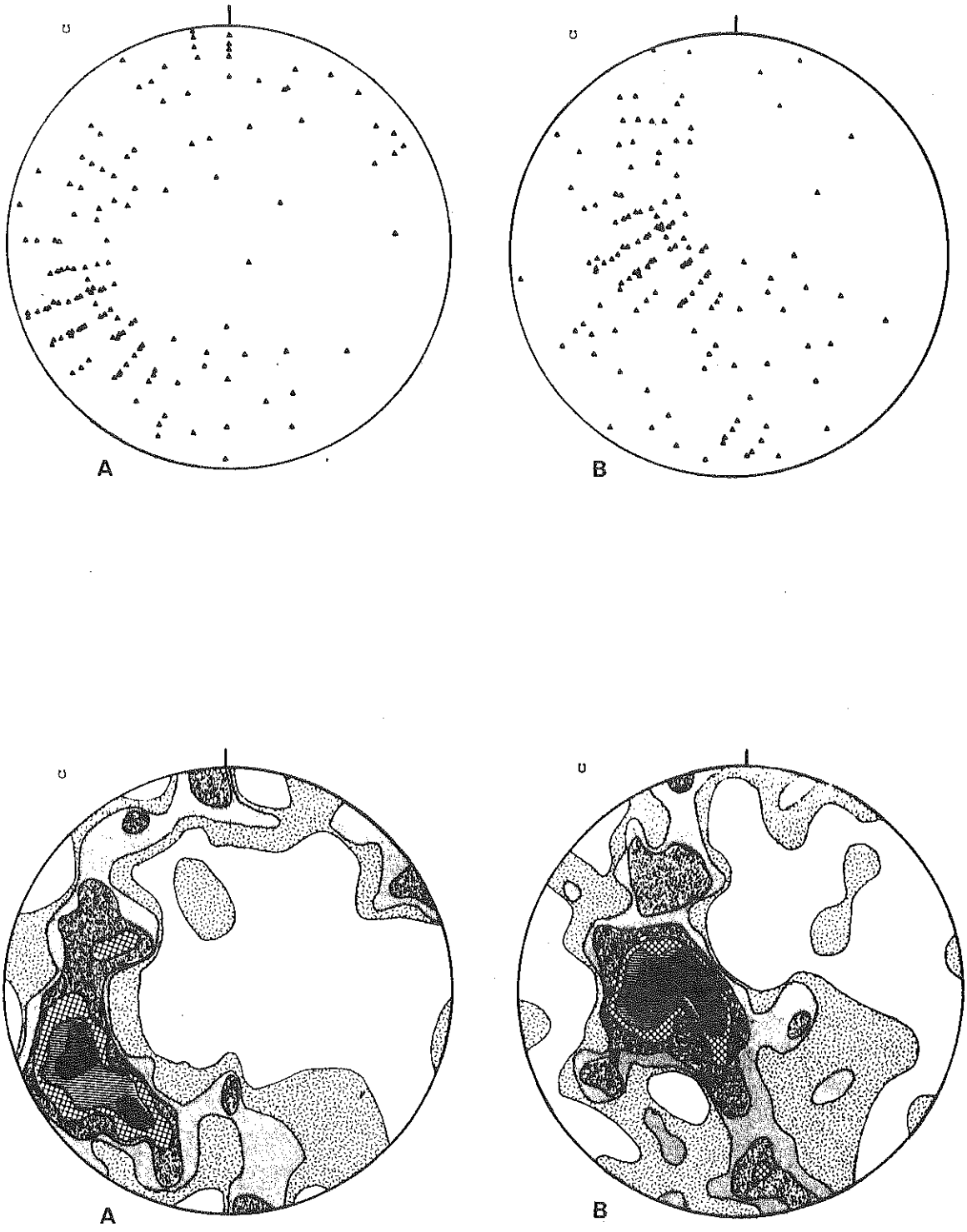


fig. 6 Point and contoured diagrams from the layering attitude of the Tabas massif. Lower hemisphere. Equal area projection. A) before rotation; B) after a rotation of 40° around an horizontal axis trending N 130. Contour interval: 0.7%, 1.5%, 3%, 4.5%, 6%,9%.

than those in the troughs. The direction and dips of the layering from the enclosing cumulates are sometimes slightly modified towards the channel margins. Generally, erosional features are observed at the base of the channel before mineral deposition. For example, current action has isolated knobs of coarse Cpx belonging to crescumulate and created gulfs like in fig. 4A. These cavities and generally the lowermost part of the channel have been filled with strongly laminated fine grained gabbro containing coarse broken Cpx. The bordered melanocratic gabbro may extend even outside the channel. The layering inside may be laterally irregular with some pinched or disrupted melanocratic layers and contorted bands linked to slump structures. Some margins are disturbed by small vertical penecontemporaneous normal faults. Sliding of the broken coarse grained layers has led to the warping of the preceeding bands (Fig. 4C). Settled blocks may penetrate the underlaying crystal mush. As previous, criss-crossed banded gabbros filled the troughs with layers thicker towards the center.

Elsewhere in the massif, uniform poorly layered rocks are sometimes cut by criss-crossed layers composed mainly of noritic gabbros and norites with acicular Opx. For exemple, in the North of the massif, a fine-grained noritic gabbro (So: N170-60E) cut pegmatitic gabbro associated with pyroxenites (So: N60-80S) more or less concordant with the general direction of the layered gabbros.

C - STRUCTURAL ANALYSIS OF THE LAYERING

The attitudes of the lithological planes vary greatly in strike and often strong dipping (fig. 6 A). From gravitational features, it may be considered that most of the bottom of the troughs have been nearly sub-horizontal or slightly inclined at the time of the deposition. Therefore, a rotation of about 40° about an horizontal axis striking N 130 permits to restore the channel in the requisite position (fig. 6 B). As previous, points and contoured diagrams reveal several maxima but less strong dip. In order to discuss the structural origin of the layering, the poles of some remarkable cumulates have been reported: 1) pyroxenite and associate pegmatite , 2) pegmatitic gabbro, 3) laminated Opx-rich gabbro-norite and 4) gabbros bordering the various channels. The superposition of the different envelopes drawn from the four groups permits to get the point maximum: N 174- 35 E of fig 5 . Two types of envelopes may be defined in the vicinity of which poles from the layering fall. The first one results from the channel borders whereas the second forms a girdle from all the other cumulates.

The poles from each studied channel have been reported on fig. 5 D as well as the axis of the resulting girdle from flanks. Depending upon the outcrops, one or two flanks have been measured. In the most favorable case, two girdles are obtained from both flanks revealing dissymetry of the channel. The Western flank presents a stronger dip than the Eastern one. The resulting Western axis striking towards North-North East has a stronger dip (15°N) than the subhorizontal Eastern one. In one case, the axis from a Western flank is directed towards East. When a solely flank exists, it corresponds to a Western flank of a channel. If magma flux is canalized by the channel, it is directed mainly towards North-North East and locally towards East, as indicated by the axes of the girdles.

The second type of girdle is obtained from the juxtaposition and locally the superposition of secant Opx-rich gabbro-norite, pyroxenite and gabbroic pegmatite. The resulting axis strikes N 70 with a dip towards East. Pyroxenite and secant gabbro-norite striking about East-West have the strongest one. Most of the other layered rocks direct from North West-South East to North East- South West. Point maxima is got from the superposition of the Western flank of channel with the secant laminated gabbro-norite and some gabbro pegmatite.

Therefore, the variable attitude of the Western flank of channels may result from a similar mecanism as for the emplacement of some gabbro-norite and pegmatite. A general wraping may be excluded because of the structure of the channel or the criss-cross relations of the cumulates. If secant cumulates fill open cracks linked to fractures, the development of some channel would be also triggered and controlled by faults. The attitude of layering from the Western flank would be directly linked to the intensity of tectonics as the local current action in response to the moving floors. All of these features occur towards the upper part of a magmatic chamber where uniform layering and static conditions of crystallization prevail.

V THE NURABAD MASSIF

This massif (1.5 long- 1 km wide) is tectonically included in serpentinites. Innner structures and numerous petrographic types indicate at first a complex mecanism of intrusion (fig. 7). From base to top, two zones may be distinguished. At the bottom, numerous cumulates are in dykes cutting peridotites. Above when the proportion of surrounding rocks decreases, the layered massif occurs, characterized by lateral variation of petrographic rock type.

A - GABBROIC DYKES OF THE SOUTERN BORDER

Towards the Southern border of the massif, numerous serpentinized peridotites, sometimes layered, are cut by strikes filled with leucocratic pegmatitic material. In peridotites, the fine pyroxenitic layers are sometimes folded. Axial planes are sub-vertical and the fold axes dip 25 to 45° towards the North. In the S diagram, the two evident girdles drawn by the S_o plane indicate no rotational movement or transport of the broken peridotites (fig. 10 A) before the fracturation in the brittle field and the related injection of magma. Peridotite fragments appear as subhorizontal slices which decrease in size and abundance towards the summit. At the bottom, rhombic fragments of peridotite are limited by numerous, wide, subhorizontal gabbroic dykes. Other dykes with higher dips and more variable in direction limit laterally the extent of peridotitic blocks (fig. 7 A). Towards the summit all the dykes getting wider and abundant separate small peridotite blocks (10 cm) with triangular sections.

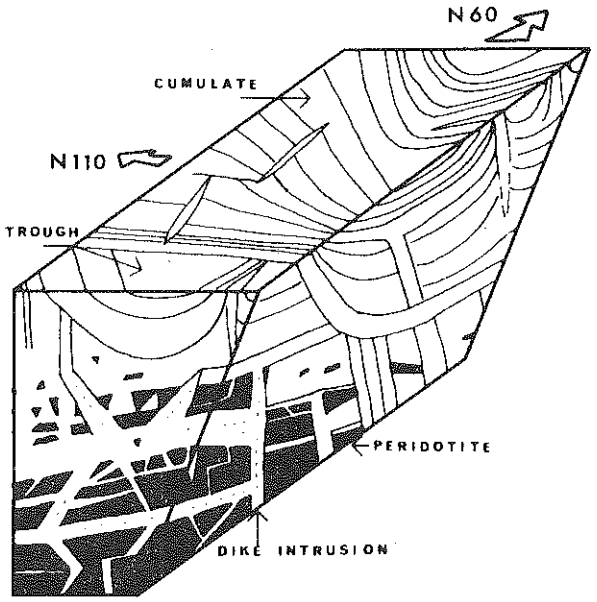


fig.7 Block -diagram from the Nurabad massif. Lower zone: peridotitic block (black) with a sub-vertical layering isoclinally folded. They are cut by dikes of variable size and attitude. Upper zone: layered massif. Note the accordance of layering from subvertical dikes to horizontal layers in troughs.

The nature and morphology of the gabbroic rocks are broadly similar in every dike. They present unchilled margin and a layering sub-parallel to the border. In most of the case, pegmatitic facies are developped at both margins with clinopyroxenes perpendicular to the layering and growing towards the center. Coarse grained leucocratic gabbro separated from the pegmatitic border by a rather sinuous plane occupies the central part. Magmatic flowage is observed in fine and medium grained noritic-gabbros with Opx and Pl aligned, which succeed towards the center to the previous pegmatitic gabbro and leucogabbro. Moreover, patches of coarse-grained acid rocks with prismatic or acicular amphiboles present in the center of some dikes print out differentiation phenomenon with collection and trapping of residual melts in the axial zone. The acid melt alters marginally the pegmatitic clinopyroxenes.

B - THE LAYERED MASSIF

Elsewhere in the massif, where no dike exists, the measured planes correspond to lithological and form contacts. The stereogram reveals the extreme diversity of the layering direction which is similar to the gabbroic dike projection (fig.8). The different attitudes of layered planes may be observed in a single place.

The Northern and Southern slopes of the massif are constitued mainly by noritic-gabbros trending East-West. Pegmatites of noritic gabbros with perpendicular pyroxene, leucocratic finer grained rocks sometimes laminated and medium grained noritic-gabbros give a wavy layering. In several areas of the Southern slopes, sub-verticallayered rocks in narrow bands cut the preceeding layering. Their borders are curved and dyssymetric. Pure gravitative features are dominant but some crystallization under static conditions also occurs. The finer banded rocks are constitued of laminated pyroxenites, medium grained norites, Pl-rich norites sometimes with broken inclusions and gabbroic pegmatites. In the Northern slope the layering, highly variable, is developed in numerous small trough structures.

The summit of the massif is made up of leucocratic coarse grained gabbro sometimes interrupted by fine grained more melanocratic rocks and by pegmatites of acid differentiates. Fine-grained noritic gabbros may be included also in medium grained leucogabbro.

In the Eastern part of the massif, petrographic rock types are more abundant and besides form contact, phase and proportional contacts are

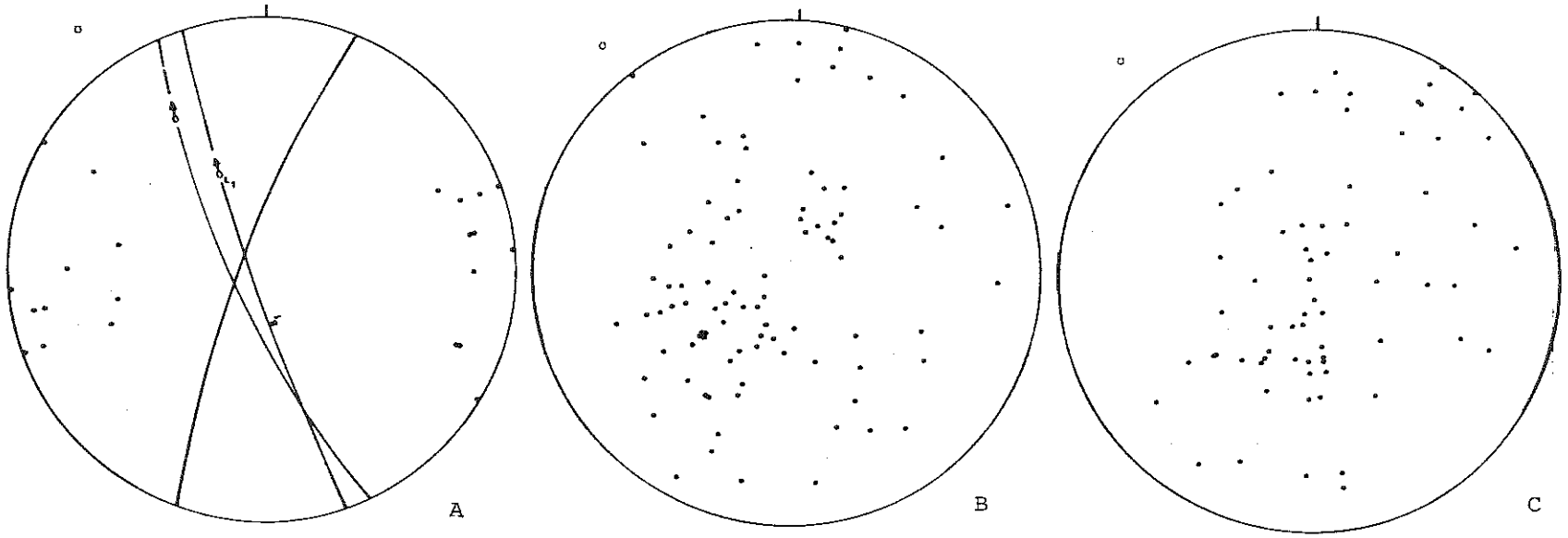


fig. 8 Point diagram from the Nurabad massif. Lower hemisphere, Wulff net
A) pole of the banding in peridotites, stereographic plot of the axial plane of fold is drawn together with fold axis, B) pole of pegmatitic dike cutting the peridotites, C) poles of the layered cumulates in the upper massif.

responsible for the layering. The banding mainly constitued of coarse grained pyroxenites, bronzitites, norites with euhedral cumulus Opx, gabbros, noritic-gabbro and diorite trending East-West to N 130, gently dips towards North. Leucocratic differentiates are concordant to the layering or cut it with a low angle. Some narrow fine-banded rocks secant to the layering with a high dip show evidence of magmatic flowage. They may be composed of more often laminated and fine-grained pyroxenites, gabbros, leucogabbros, norites with oriented acicular Opx.

The Nurabad massif is mainly constitued of noritic gabbros with subordinate pyroxenite, norite and gabbro in the Eastern part and leucogabbro towards the summit. Coarse grained rocks, pegmatitic facies with perpendicular minerals giving a wavy layering dipping gently towards the North are more abundant than laminated fine grained rocks better developed in secant narrow bands trending in two main directions about N 160 and East-West. In the South, the presence of undisturbed layered mantle peridotite cutted by pegmatitic gabbros suggests the late development of magmatic chamber in low pressure areas where brittle failure occurs. Magmatic differentiation may underwent in the dikes which center contains locally acid residues. The similarities in dike orientation and layering in the massif and the characteristics of magmatic features suggest that all the gabbros have been emplaced under similar tectonic conditions. The Nurabad massif may represent a margin of a magmatic chamber emplaced in a relatively cold mantle peridotite in which shear fracture regime controls the incoming of magma pulses and the crystallization history: magmatic differentiation and layering. This fact may explain the abrupt petrological variations observed troughout the massif.

VI THE DEVELOPMENT OF MAGMA CHAMBER. EVIDENCE FROM THE COMPARISON OF THE THREE MASSIFS.

Although some discrepancies may be pointed out between the massifs, most of the structural and petrological facts obtained from their study may lead to some general considerations about the formation of magma chamber in ophiolites. All the massifs are tectonic slices from cumulitic pile. The most complete section which corresponds to a vertical zoned magma chamber is given by the Solemanieh massif although tectonized at the base and summit. The crystallization order Sp-Ol-Cpx-(Pl±Opx) is similar to the more usual series in type II ophiolites (ROCCI et al., 1975).

The Tabas massif may be the upper part of a chamber represented by the Solemanieh cumulitic series. Both have at the same level common crystallization sequence and structural features. Nevertheless in some places secant cumulates (pyroxenite-norite) with early Opx testify to other late magmatic event with a SiO_2 sursaturated character. Such cumulates are also well exposed in the Nurabad massif. As it possesses peculiar petrographic relationships it does not represent a piece of the Solemanieh type magma chamber. In any case, its origin is probably linked to a zone of magma migration where liquid inputs have been imprinted. The layering from these massifs may result from the superposition or juxtaposition of common processes.

Some accumulative features may be related to the stratigraphic position in the chamber. For example, in the Solemanieh massif, the size of the minerals and the layers decrease and adcumulate are progressively replaced by meso, ortho and locally heterad and crescumulates. Pure gravitational features on a small scale are well developed in the upper part and are often associated with current action. In every massif, the importance of intrusion and layering related to dyke injection have been revealed. The base and edge of the chamber are generally cut by wide dike showing comb structure, pegmatite and in situ differentiation like in the Solemanieh and Nurabad massifs. In this last one, all the lithological planes cover the field of pegmatitic dikes. Moreover, in this massif and in the Tabas one, the presence of secant cumulates, sometimes with diffused contact and linked to flow differentiation, untill a high level, testifies to the importance of dike intrusion during all the evolution of the magma chamber. They generally occupy a meridian position.

It may be emphasized that flow differentiation is especially related to the latest magmatic intrusion having SiO_2 sursaturated character (pyroxenite, norite with acicular Opx) compared to the main magmatic event. Similarly, layering features may be highly modified by the development of troughs of variable size and morphology .They are more or less dyssymetric and imply the presence of ridges and cavities on the floor which force the magma to migrate in some directions.

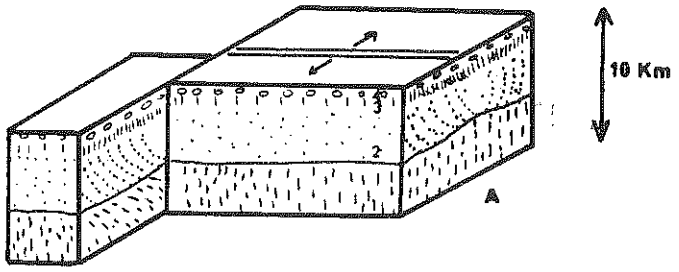
In the three massifs, all the lithological planes have been shown to correspond to troughs, dike and planes of deposition. East-West to N 130° are the more obvious strikes of the layering but with very different dips

. Its variable inclination towards North may be explained by 1) late rotation around an East-West subhorizontal axis. 2) the progressive accordance between vertical cracks and a slightly inclined plane of deposition sub-perpendicular to the Earth gravitational forces. This latter hypothesis testifies to major injection slightly oblique to the mountain belt. Other sub-vertical planes with variable attitude correspond to dike intrusion related to the opening of fractures. Similarly, it has been pointed out that the creation of troughs results from moving zones. Especially in the Tabas massif, the dip of the Western flanks varies in response to displacement along fractures. Consequently, faults which control dike intrusion and trough are also responsible for the attitude of layering which mimics their strikes (fig.7). Mineral deposition due to Earth gravitational forces is favored mainly between fracture zones. In a chamber, trough and dike will be especially developed towards all the borders whereas sub-horizontal rhythmic layering will occur towards the core of the chamber during the lower to middle stage of its evolution. Dike intrusion leading to replenishment of the chamber intervenes mainly at the beginning. Layering in the upper part of the chamber is often modified by several processes due to injection and departure of acid plugs towards the surface. Inside the plugs, of gabbro-norite to dioritic composition, sub-horizontal discontinuous layers may contrast with the subvertical flowage of the secant cumulates.

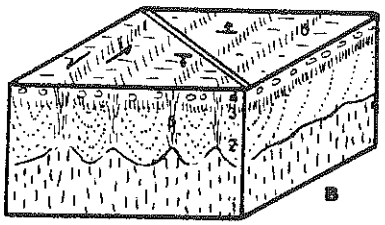
At every level in the chamber, the layering results from the crystallization along steeply inclined fracture zones and sub-horizontal floor created by gravitational forces. If the attitude of cumulates mimics the border of the chamber, its morphology and size which depend on the depth of emplacement will result also from the tectonic regime of the expanding crust. Especially, numerous sub-vertical fracture zones will favor the juxtaposition of small magma chambers.

VIII PRIMARY EMPLACEMENT OF OPHIOLITES IN THE SABZEVAR BELT.

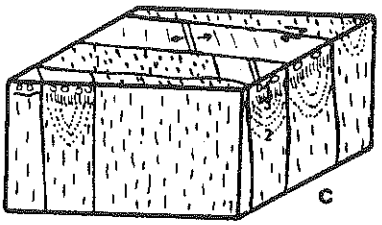
The preceeding chapter has emphasized the role of fracture zone to control the size, morphology and layering of the magma chamber. The attitude of the petrological planes may also be compared with the strike of the sheeted complex and the flow line in peridotites to precise the tectonic activity of the accretionary zone. Important intrusion of basic magma



MID-OCEAN RIDGE + TRANSFORM FAULT



OBLIQUE SPREADING:
SUPERPOSITION OF RIDGE AND TRANSFORM FAULT TECTONICS



OBLIQUE SPREADING IN PULL APART BASINS LINKED TO STRIKE-SLIP SYSTEM

fig. 9 Accretion in extensive to strike-slip system. The spreading axis is always indicated. A) ridge interrupted at transform-fault boundaries, B) superposition of ridge and transverse faults, oblique spreading, C) creation of pull-apart basins linked to strike-slip system. 1) peridotites, 2) cumulate from magma chamber, 3) sheeted dike, 4) lavas, 5) fracture zone.

and associated rocks may occur from extensional to strike-slip systems. It may be expected that in ophiolites, some geological and structural features would be compatible either with one system or the other. In extension zone, an active spreading ridge leads to an oceanic crust later obducted on a passive margin (JUTEAU et al., 1977). The fossil spreading axis, nearly parallel to the present-day ophiolitic belt may be indicated on the field by the direction of mineral lineation in peridotites and cumulates and the dyke complex. As peridotites show relictual diapiric structure: sub-vertical foliation at a high angle to the layered cumulate formed by gravity, the lineation is of little use to determine the direction of the ridge axis. Most of the layered planes in cumulates are parallel to slightly oblique to the spreading axis whereas magma migration in the channel structure follows transverse zones interrupting the ridges (fig. 9A).

Some hypabyssal rocks are at a small angle to the major cumulate and the other sub-vertical dykes are perpendicular or at a high angle to the ophiolitic belt. Their variable attitude as well as the other structural elements deduced from peridotite or cumulate seem to be insufficiant to assume an East-West trending ridge created in an extensional system. Most of the observed features may be better explained if strike-slip movements play an important role associated or not with extensional tectonics. In the first case, small ridge segments are interrupted by numerous transverse fracture zones (fig. 9 B). If the offset is lower than 20 km, we may expect the superposition of ridge and transverse tectonics leading to complex fracture zone and scarps like in ocean (WHITMARSH and LAUGHTON,1975 SEARLE, 1979). For example, all the radiating dikes in the sheeted complex or in cumulate and the channel structure may be explained by transverse ridge and "en échelon" pattern. Flowage in the peridotite may occur in multiple directions along limited zones which favor the preservation of the relictual diapir. This model which leads to oblique spreading and to existence of leaky-transform faults ; THOMPSON and MELSON, 1972 ; VOGT and JOHNSON,1975) is more compatible with the observed facts.

The shearing of continental crust and creation of ophiolites may be also only related to strike-slip movements. Accretion along a linear belt parallel to vertical faults may occur inside pull-apart basins (fig. 9 C). Few, discontinuous extensive zones are in agreement with the high proportion of peridotites (80%) against cumulates and lavas, if ultramafic emplacement does not result from secondary tectonic emplacement. There the lineation in diapir may be variably inclined depending upon the superposition of vertical and horizontal movements in the fracture (OHNENSTETTER, 1980). The sub-vertical foliation plane in the peridotite, trending about North-South is also in agreement with a maximum stress axis at a small angle to the belt. The attitude of the layering in cumulates which mimics the border of the rhombic graben and dike intrusion related to the spreading axis is highly variable. Dikes sub-perpendicular or at a high angle to the strike of the belt are compatible with this oblique spreading.

VIII CONCLUSION

The attitude of the main structural planes have been examined in mantle rocks, dike swarm and mainly in cumulates from a portion of the Sabzevar ophiolitic belt to define an emplacement model during accretion.

The nearly vertical foliation plane indicates that peridotites rise in several points of the chain like diapirs in which syntectonic recrystallizations prevail. Numerous vertical shear fractures prevent from the emplacement of wide magmatic chambers. Magma pulses of variable compositions crystallize at different depths injected either in peridotite or in the effusive formations: sheeted diabase complexes, lavas and sills. Layering depends about the gravity and the magmatic flowage in tensile and shear fracture zones. Faults which control the magmatic injection and the shape of the magma chamber as well as the dike swarm are thought to be linked to the tectonics of opening. Several accretionary processes have been reviewed from pure extensive systems to opening related to strike-slip system. It seems that this later one is important either to create limited pull-apart basins or to modify the ridge tectonics if numerous transverse zones exist.

Wrench faults may occur in numerous geotectonic environments from continental to oceanic areas and island arc (KOIDE and BHATTACHARJI, 1977; EGUSHI et al., 1979; GARFUNKEL et al., 1981; PAGE et al., 1981). In the Sabzevar range additional informations from the composition of ophiolitic magma and the time of accretion (BAROZ et al.,in press) permit to put some other constraints on the geotectonic environments. On the basis of liquid composition, a genesis in a Mid-ocean ridge or from a marginal basin far from a subduction zone is excluded. As the accretion was of short duration and later followed by a mature island arc (LENSCH et al., 1977) ophiolites may have been created near a continental border which has later evolved towards an active margin. The existence of strike-slip system is also compatible with short-lived basins later emplaced on continental margins without large thrusting (BROOKFIELD,1977; NOGOLE-SADATE, 1978). These tectonics may be linked to the East-West sliding of continental blocks during a general convergence of major plates.

ACKNOWLEDGEMENTS

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OPHIOLITIC SEQUENCES IN CORSICA

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ABSTRACT

In eastern Corsica, the ophiolite sequences related to mid-oceanic ridge or to transform fault zone are presented. In this latter case, oceanic accretion may occur by contemporaneous basalts and more acidic residue cutting ophicalcites. Some basalts related to ridge segments may present a transitional character with a LIL-enrichment. These variations are related to successive normal and anormal oceanic ridges. The most important type of sedimentary cover are evocated. The lherzolites and cpx-poor lherzolites are considered to be residual products left after partial fusion giving the oceanic tholeiites. In particular, the Monte Maggiore Massif possesses preserved protogranular and porphyroblastic texture related to partial fusion events during diapiric uprise of the mantle. Contamination of the more or less depleted mantle by the veining liquids are thought to occur at lower pressure in oceanic environment.

In Eastern Corsica, Mesozoic sediments and ophiolites of piemontaise and Ligurian origin have been disrupted, folded and metamorphosed during the Alpine orogen (Durand Delga, 1978). In the Schistes lustrés four lithostratigraphic series, two of which are associated with ophiolites, have been recently defined (Caron et al., 1979). The Inzecca series of Malm to Upper Cretaceous age by analogy with the Apennines, comprising from bottom to top radiolarites, limestones, interbedded shales and limestones and finally shales and quartzites, has been long known as the most important cover above pillow lavas associated with gabbros and serpentinites (Amaudric du Chaffaut et al., 1972). The Santo Pietro di Tenda series defined by Delcey (1974) is now considered to be an other postophiolitic sedimentary series more epicontinental: abundant limestone and detrital material especially rhyolitic pebbles (Caron and Delcey, 1979). This series presents some analogies with the sediments of the upper ophiolitic series in Calabria (Lanzafame et al., 1979).

The ophiolites themselves are dismembered. A reconstruction of a complete se-

quence has been proposed on petrological considerations (Ohnenstetter and Ohnenstetter, 1975). From bottom to top ophiolites are built up of a lherzolitic ultramafic unit, a mafic unit with troctolite, euphotide, ferrogabbros and acidic residue, massive dolerite, few dikes and pillow lavas. Another ophiolitic series, defined in two places is composed of gabbro breccias, ophicalcites, agglomerates, massive diabases and albitites resting on a serpentinite and/or gabbro floor. This series may have been formed in a transform fault zone in which peculiar oceanic accretion has occurred (Ohnenstetter M., 1979). These ophiolitic slices have been metamorphosed during three main events (Ohnenstetter et al., 1976): a sub-sea floor metamorphism (greenschist to amphibolite facies), a high pressure - low temperature polyphased event related to a possible subduction zone, perhaps intra-oceanic (Mattauer and Proust, 1976) of eo-Alpine age (Maluski and Mattauer, pers. comm.) and then a last greenschist metamorphism during Eocene-Oligocene age. Important tectonism occurred after these different metamorphic events (Amaudric du Chaffaut and Saliot, 1979).

In the volcanic pile formed by abyssal tholeiites, the pillow lavas are often the less differentiated, whereas massive dolerites and dikes show a marked Fe enrichment as well as Ti, V, Zr, Y, P and Zn, and a Mg, Al, Ni and Cr decrease. Similar trends occur in present day oceanic tholeiites (Beccaluva et al., 1977, Gruppo di Lavoro sulle Ofioliti Mediterranee, 1977). The mineral phases controlling the fractional crystallization are found in the gabbros: olivine ± Cr-spinel, plagioclase, clinopyroxene, iron ores and apatite. Immiscible processes may occur in the latest stage permitting the acidic residue to escape (Ohnenstetter et al., 1979, Ohnenstetter and Ohnenstetter, 1980). This is in agreement with experimental data (Dixon and Rutheford, 1979). Moreover, in the Rospigliani series, besides basalts and ferrobasalts, scarce andesites and dacites, comparable to those of the Galapagos ridge, have been recognized (Ohnenstetter M., 1979, 1980). Basalts and more acidic liquids have been emplaced simultaneously. In the tholeiitic basalts LIL and incompatible elements vary from place to place giving a transitional character to some lava pile: in Balagne, basalts are slightly light REE enriched (Venturelli et al., 1979) similarly as the Montgenèvre, Alps, lavas (Lewis and Smewing, in press) and Calabria basalts (Dostal et al., in press). In opposite, the Inzecca basalts possess a depleted LREE pattern (Allègre et al., 1976, Noiret and Allegre, 1976). The pyroxenes from the Inzecca lavas have Na and Ti contents similar to those formed in normal mid-oceanic ridge segments whereas the Balagne py-

roxenes are Na and Ti richer (Capedri and Venturelli, 1979) in agreement with the transitional character of its lavas. In the Rospigliani series ferrobasalts possess augite accompanied by subcalcic augite and pigeonite. Similar geochemical and petrographical variations may be found in oceanic basalts along mid oceanic ridge when normal ridge segments succeed to anormal ones (Bryan et al., 1976, White and Shilling, 1978). In this latter case, alkalinity may increase with variable LIL concentrations.

In Corsica few peridotite massifs are fresh. The partly serpentinized Monte Maggiore body possesses a poor layering, protogranular texture and a foliation plane in which the spinel lineation is slightly inclined. A weak olivine and orthopyroxene fabric is compatible with a sinistral shear movement (Jackson and Ohnenstetter M., in press). Lherzolite and clinopyroxene-poor lherzolite are associated with harzburgite and dunite which mineral compositions are similar to the oceanic ones. Further detailed structural and petrographical studies indicate that a major partial fusion event occurs in the Seiland facies (O'Hara, 1967). The percolating liquids issued from different zones migrate through a more or less depleted peridotite, well layered, folded and equilibrated in the spinel facies. The veining liquids modify locally the peridotite assemblage with the crystallization of apatite, and Ti spinel, pyroxene and amphibole. These minerals are important to control the leaching of incompatible elements (Beswick and Carmichael, 1978) and the composition of the eutectic liquid. The peculiar Na-rich clinopyroxene in numerous Alpine bodies may be explained also by a metasomatic phenomenon (Kornprobst et al., submitted). When the peridotite was in the brittle field, it was cut by numerous dikes which testify of the magma ascent.

Most of the magmatic features observed in ophiolitic rocks especially lavas and peridotites may be explained in an oceanic environment. The lherzolite and cpx-poor lherzolite in the spinel facies represent residues left after partial fusion giving the oceanic tholeiites. The veining liquids in the Seiland or in plagioclase facies may play a role for the creation of anomalous ridge segments.

In the Ligure-Piemontais ocean, small ridges are interrupted by abundant transform fault (Abbate et al., 1972, 1980,Gianelli and Principi, 1974, Gianelli, 1977, Lemoine and Tricart, 1979, Ohnenstetter M., 1979, Bertrand et al., 1980, Lemoine, 1980) characterized by peculiar lithostratigraphic series and sometimes magmatism. The sedimentary cover of the ophiolitic lavas present lateral facies variation according to distance from its passive margin (Bourbon et al., 1977). The closure of

this ocean is still an unsolved problem.

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GÉOLOGIE. — *La série ophiolitifère de Rospigliani (Corse) est-elle un témoin des phénomènes tectoniques, sédimentaires et magmatiques liés au fonctionnement des zones transformantes* ? Note (*) de **Maryse Ohnenstetter,** présentée par Georges Millot.

Une série ophiolitifère volcano-sédimentaire, appelée série de Rospigliani, et constituée de brèches à éléments de gabbros, d'ophicalcites, d'agglomérats, de diabases massives et d'albitites, se serait formée dans une zone transformante, sur un substratum de serpentinites et/ou de gabbros. Les événements tectoniques, sédimentaires et magmatiques, en particulier la contemporanéité de certaines émissions acides et basiques, prouvent qu'un certain type d'accrétion océanique peut avoir lieu dans les failles transformantes.

A volcano-sedimentary ophiolitic series, the Rospigliani series, consisting of gabbro breccias, ophicalcites, agglomerates, massive diabases and albitites may have formed on a serpentinite and/or gabbro floor possibly in a transform-fault zone. The tectonic, sedimentary and magmatic events, especially the contemporaneity of some acid and basic magmas intruded into the sediments, indicate that a particular type of oceanic accretion may occur in transform zones.

Les travaux récents sur la stratigraphie des schistes lustrés de la Corse alpine [1] ont abouti à la distinction de deux séries postophiolitiques : celle de l'Inzecca [2] et celle de Santo-Pietro-di-Tenda qui aurait une origine plus externe que la précédente [3]. Leurs sédiments reposent sur des laves en coussins, elles-mêmes superposées, au moins dans la série de l'Inzecca, aux gabbros et aux serpentinites [2]. L'étude des fragments ophiolitiques en position « infrajaspe » nous a conduit à constater que l'association de certaines laves, gabbros et serpentinites avec des ophicalcites et des brèches sédimentaires n'était pas fortuite. Les analogies relevées dans les massifs de la Casaluna et au nord de l'Inzecca permettent de définir un type de série ophiolitifère : la série de Rospigliani. Bien que les recherches soient encore préliminaires, la genèse de cet assemblage qui met en œuvre plus ou moins simultanément des facteurs tectoniques, sédimentaires et magmatiques permet de compléter nos connaissances sur la paléogéographie alpine et sur la mise en place des magmas en milieu océanique.

1. COUPES DANS LE NORD DU MASSIF DE L'INZECCA. — 1. *Situation géologique.* — Le massif de l'Inzecca (*fig.* 1), bien connu pour sa trilogie ophiolitique [4] et sa formation sédimentaire postophiolitique [2], présente à sa partie NE, de part et d'autre du Tagnone, une zone riche en « écailles ophiolitiques » de petites dimensions ([5], [6]). Celles-ci sont alignées à N 140 et insérées dans la formation d'Erbajolo, terme supérieur de la série de l'Inzecca, considérée comme étant du Crétacé inférieur ([2], [7], [8]). C'est dans ces écailles que les relations entre laves, albitites, dolérites, gabbros, serpentinites, ophicalcites et brèches de natures diverses vont être examinées à l'aide de quelques coupes.

2. *Punta di Castelnuovo.* — Au nord-ouest de la Punta di Castelnuovo (*fig.* 2 *a*), des ophicalcites sont surmontées par des agglomérats et des brèches à matrice de hyaloclastites, qui enferment des inclusions de roches carbonatées à proximité du contact. En s'éloignant de celui-ci, les ophicalcites deviennent rapidement schisteuses; une matrice chloriteuse enserre alors des noyaux de calcite, puis des éléments de ferrodiorite. Progressivement, on passe à une brèche de ferrodiorite qui contient çà et là des éléments acides, le tout envahi par des filonnets de calcite. Vers l'Est, un filon de 3 m de large, constitué d'albitite et de basalte mis en place simultanément, recoupe les ophicalcites. La déformation plastique observée dans les roches acides de certains filons est contemporaine de leur mise en place.

3. *Rospigliani.* — Au sud de Rospigliani, des brèches volcaniques sont recouvertes par des jaspes. Des roches acides, de grain et couleur variables, recoupent des laves et/ou parti-

cipent à la formation de brèches magmatiques à matrice leucocrate, dont les textures indiquent l'affrontement des magmas acide et basique (*fig.* 2 *b*). Des ophicalcites sont localement associées à ces brèches.

D'autres écailles montrent que l'injection de ces magmas en filon et en sill, dans un matériel gabbroïque plus ou moins plastique (flaser gabbro), a abouti à la formation de brèches magmatiques polygéniques.

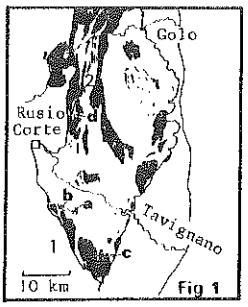


Fig. 1. — Disposition des ophiolites dans les Schistes Lustrés indifférenciés des massifs étudiés en Corse orientale d'après Caron et coll. (1). 1, massif de l'Inzecca ; 2, massif de la Casaluna ; *a*, Punta di Castelnuovo ; *b*, écailles de Rospigliani ; *c*, Punta Corbara ; *d*, Punta Oscaglia et Punta Muzzone ;

4. *Punta Corbara.* — Au sud-est du massif de l'Inzecca, on peut retrouver la plupart de ces roches. Il existe en plus, vers la Punta Corbara, des brèches sédimentaires remaniant des fragments de ferrogabbros, ferrodiorites, albitites, basaltes et dolérites qui sont situées sur des ophicalcites (*fig.* 2 *c*). Ces brèches à stratification grossière, de granulométrie très variée, peuvent s'interrompre latéralement. Elles sont soit monogéniques, soit polygéniques, et possèdent dans ce dernier cas des éléments plus émoussés. La matrice généralement chloriteuse peut devenir plus acide, spécialement dans les brèches polygéniques.

Ailleurs, des serpentinites vertes à noires, localement transformées en taleschistes et veinées de calcite blanche, sont associées aux ophicalcites (*fig.* 2 *d*).

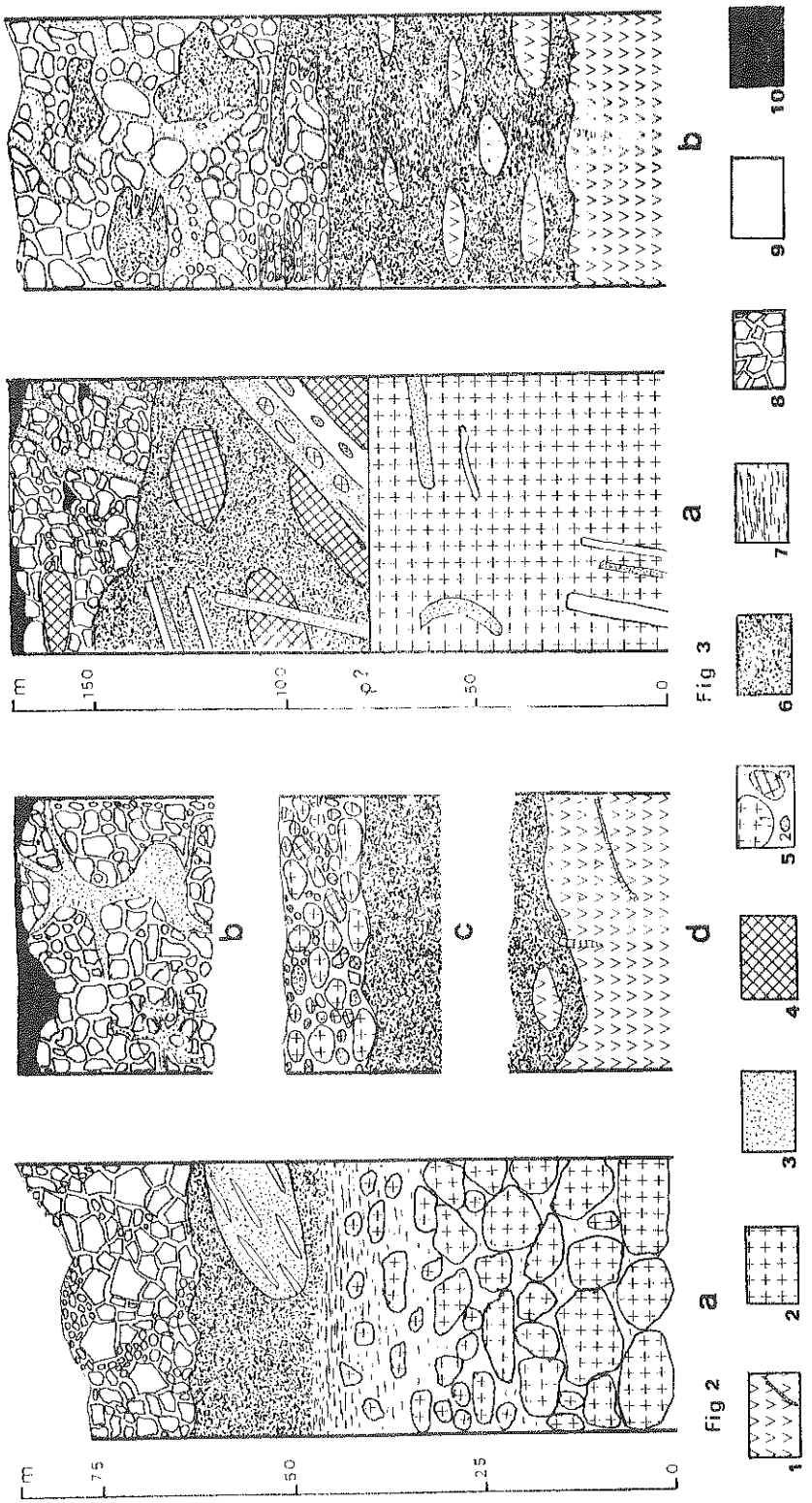
II. Coupes dans la Casaluna. — 1. *Situation géologique.* — Dans la région de la Casaluna, à l'est du village de Rusio, des formations volcano-sédimentaires apparaissent entre les gabbros différenciés du Cervone et les schistes lustrés couvrant la partie Nord du Monte Maggiore ([6], [9]). Deux coupes E-W ont été relevées de part et d'autre d'un thalweg (*fig.* 1).

2. *Oscaglia.* — La coupe de la Punta Oscaglia située en rive droite est la plus orientale (*fig.* 3 *a*). Les termes sommitaux de la série sont constitués par des jaspes de couleur rouge

Explication de la Planche

Fig. 2. — Coupes schématiques dans le massif de l'Inzecca ; *a*, Punta di Castelnuovo ; *b*, Rospigliani ; *c* et *d*, Punta Corbara. 1, serpentinites avec filonnets de calcite ; 2, gabbros plus ou moins déformés et ferrodiorites ; 3, albitites ; 4, dolérites et basaltes phyriques ; 5, brèches à éléments : 1, gabbros et ferrodiorites, 2, albitites, 3, dolérites ; 6, ophicalcites avec filonnets de calcite ; 7, formations schisteuses ; 8, agglomérats ; 9, basalte aphyrique ; 10, jaspes.

Fig. 3. — Coupes schématiques dans le massif de la Casaluna ; *a*, Punta Oscaglia ; *b*, Punta Muzzone. Même légende que la figure 1 pour les sigles (1 à 10).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vif qui reposent de façon plus ou moins discontinue sur des diabases aphyriques à légèrement phyriques. Ils sont insérés dans ces laves massives ou dans des agglomérats à matrice locale de hyaloclastites et de brèches. Du magma acide a été injecté dans toutes ces roches. Les ophicalcites sous-jacentes apparaissent juxtaposées à des diabases légèrement phyriques. Les deux formations sont recoupées par des filons, aux directions voisines, constitués de diabases aphyriques et de roches acides renfermant des xénolithes de gabbros, amphibolites et ophicalcites. Il arrive que leurs mises en place simultanées s'accompagnent d'importants phénomènes de bréchification. La base supposée de la série s'appuie sur des gabbros, par l'intermédiaire d'un contact vertical dont la nature est délicate à préciser.

3. *Punta Muzzone.* — Sur la rive gauche du thalweg, les affleurements de la Punta Muzzone présentent des serpentinites accolées aux ophicalcites ou incluses dedans, le tout veiné de calcite blanche (*fig.* 3 *b*). Des brèches volcaniques mal classées, recoupées par d'abondants filons acides, forment des promontoires dispersés dans les roches carbonatées. Localement, des successions d'ophicalcites et de brèches volcaniques se répètent plusieurs fois sur 2 m.

III. Caractéristiques de la série de Rospigliani. — La synthèse des observations précédentes conduit à reconstituer provisoirement la succession des phénomènes volcano-sédimentaires de la série de Rospigliani.

Les jaspes de faible épaisseur, qui peuvent être interstratifiés dans les laves, représentent probablement le sommet de la série et un équivalent possible des radiolarites du Jurassique supérieur situées à la base des séries postophiolitiques [8]. Dans les émissions volcaniques dont la puissance est limitée, les laves en coussins sont peu fréquentes à l'opposé des agglomérats, brèches et coulées massives plus ou moins phyriques. Ces laves sont superposées aux ophicalcites ou alternent avec elles.

Un magmatisme acide est souvent contemporain des émissions basiques. La persistance des injections est attestée par la présence des filons de nature différente dans les ophicalcites.

Les ophicalcites sont associées soit avec des serpentinites, soit avec des brèches sédimentaires à éléments de gabbros. Ces fragments proviendraient de l'altération et de la destruction de la croûte océanique, constituée de serpentinites et de gabbros qui joueraient le rôle de plancher pour la série de Rospigliani.

IV. Pétrogenèse des zones transformantes. — On sait que, dans les zones transformantes, les différents termes de la croûte océanique apparaissent, de même que de nombreuses brèches, en particulier celles dont le ciment est carbonaté, liées à la surrection des panneaux de serpentinites [10]. On est donc tenté de placer la genèse de la série de Rospigliani dans une zone transformante, rejoignant dans cette démarche les auteurs ayant travaillé sur des séries voisines dans les Apennins [11] et sur des formations des Alpes [12].

Admettons que cette hypothèse soit la plus plausible, d'après nos connaissances sur l'expansion océanique. Nous devons alors tenir compte des observations effectuées sur la série de Rospigliani, en particulier celles concernant les phénomènes magmatiques. Elles peuvent en effet contribuer à l'élaboration de modèles sur le fonctionnement des zones transformantes actuelles.

Dans ce contexte, la découverte d'injections simultanées de magmas acide et basique souligne la diversité du magmatisme dans ces zones. De plus, la présence de roches acides cristallisées à partir de liquides immiscibles [13] laissent supposer l'existence de chambres magmatiques, où se forment des cumulats plus ou moins évolués [14] et d'où s'échappent des liquides à divers degrés de fractionnement.

Ainsi, il y aurait formation de croûte océanique dans ces zones transformantes [15], qui ne seraient donc pas de simples fenêtres donnant sur les couches océaniques profondes [16].

Les brèches sédimentaires remanieraient des fragments de croûte océanique créée au niveau des dorsales et dans les failles transverses, remontés grâce aux diapirs de serpentinites ([15], [16], [17]). Enfin, s'il existe des séries ophiolitifères caractéristiques des zones transformantes, on doit pouvoir les distinguer de celles que produisent les rides médio-océaniques normales et anormales [18].

(*) Remise le 12 novembre 1979.

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DEUXIÈME PARTIE

LA PLACE DES OPHIOLITES

DANS LES RECONSTITUTIONS TECTONIQUES

"Le fonctionnement de ces sutures ne peut faire l'objet que d'hypothèses. Celle de phases de distension précédant les phases de compression tectogéniques, est la plus satisfaisante pour l'esprit, mais il est clair que les premières, contrairement aux secondes, ne laissent pas de traces, hormis, justement les montées magmatiques".

J.H. BRUNN (1961)

DEUXIÈME PARTIE

LA PLACE DES OPHIOLITES DANS LES RECONSTITUTIONS TECTONIQUES

I - LES TYPES D'OUVERTURES POSSIBLES

1 - Les zones en extension localement interrompues par des transformants

2 - Ouvertures obliques par rapport à la chaîne

3 - Ouvertures liées à des décrochements

4 - Conclusion

II - LA MISE EN PLACE SECONDAIRE DES OPHIOLITES

1 - Collisions frontales

2 - Collisions obliques

III - LES OPHIOLITES ET TERRAINS ASSOCIES

1 - Les terrains antérieurs

2 - Les ophiolites et sédiments associés

a) La composition des ophiolites
b) La durée de l'accrétion

3 - Les formations post-ophiolitiques

IV - L'ACCRETION OPHIOLITIQUE : SON ROLE DANS L'OROGENE

1 - Rappel de quelques données fondamentales

2 - Développement de l'orogène

DEUXIÈME PARTIE

LA PLACE DES OPHIOLITES DANS LES RECONSTITUTIONS TECTONIQUES

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Vouloir interpréter le rôle des ophiolites dans les reconstructions paléogéographiques revient à situer leurs mises en place primaire et secondaire dans le temps et l'espace par rapport à l'orogène. A priori, à l'échelle de la tectonique des plaques, la montée des magmas basiques et la formation du cortège des roches grenues traduisent une certaine distension compatible avec des systèmes en extension ou en décrochement tandis que l'empilement de nappes ou d'écailles implique un serrage lié à des systèmes en compression ou en décrochement. Dans ce cadre, il est généralement admis que la position des ceintures ophiolitiques dérive de la succession de systèmesen extension puis en compression. Cependant, BROOKFIELD en 1977 a insisté sur les processus de mise en place secondaire des ophiolites facilitée par de grands décrochements. Le rôle de ces derniers a également été souligné par moi-même en 1980 pour l'ouverture de certains bassins ophiolitiques. Les décrochements expliqueraient certaines discontinuités structurales entre les sutures au niveau des péridotites, de la chambre magmatique et des roches hypabyssales (OHNENSTETTER M., 1980 ; I.2).

Si la géométrie interne des ophiolites peut donner quelques indications sur leurs conditions de mise en place primaire et secondaire, en revanche elle est insuffisante pour estimer le lieu d'accrétion dans un hypothétique bassin et la durée des différents processus envisagés. Ces informations peuvent être recherchées dans les terrains associés, antérieurs, contemporains ou postérieurs aux ophiolites et qui sont formés à des distances variables de la marge. L'écart de temps entre mise en place primaire et secondaire peut être aussi apprécié par l'âge des séries stratigraphiques de part et d'autre des discordances angulaires. Par ailleurs, la nature du site géotectonique de même que l'évolution du manteau au voisinage des zones d'ouverture peuvent être estimées d'après la composition chimique des laves ophiolitiques : des concentrations élevées en éléments incompatibles traduiraient une origine loin d'une zone de subduction (océan ou mer marginale) tandis que les teneurs plus faibles, indiqueraient une origine à proximité d'un arc insulaire.

En conséquence, la mise en évidence de discontinuités géochimique , stratigraphique et structurale, réalisée principalement au cours de la première partie, nous a incité à développer une conception non unitaire des ophiolites. C'est pourquoi cette partie dresse un bilan des mécanismes possibles d'ouverture et de fermeture, de la durée et du lieu d'accrétion avant de juger des conditions de mise en place primaire et secondaire des ophiolites. D'une manière

générale, on notera la brièveté de l'accrétion, l'importance des décrochements dont les rejeux tectoniques peuvent se substituer le plus souvent à ceux du couple extension-collision frontale lors de la genèse des sutures. Ces considérations nous ont guidé pour les reconstitutions paléogéographiques proposées pour la Grèce (BEBIEN et al., 1980 ; II.2) et la chaîne de Sabzevar, Iran (BAROZ et al., à paraître ; II.3) qui s'intègrent dans l'évolution des régions méditérranéennes et du Moyen-Orient (ROCCI et al., 1980 ; II.1).

I - LES TYPES D'ACCRETION POSSIBLES

La simple considération de la position des contraintes dans l'espace incite à penser que seules les fractures profondes liées aux décrochements ou à des systèmes en extension peuvent faciliter la montée des magmas (DIETZ, 1966) et l'établissement de chambres magmatiques sous-crustales (OHNENSTETTER M., 1980). Bien évidemment, on peut imaginer toutes sortes d'ouvertures possibles suite à une extension plus ou moins oblique par rapport à la marge ou à la future chaîne de montagne. Pour simplifier on schématisera trois types de rupture de croûte :

1 - Les zones en extension localement interrompues par des transformantes

Les modèles d'accrétion des rides Médio-océaniques en état d'équilibre sont compatibles avec des systèmes en extension (fig. II.1). Un même contexte est largement admis pour la création des ophiolites (CANN, 1974 ; NICOLAS et VIOLETTE, 1982). Des modifications sur la morphologie des chambres, la présence ou non de complexe filonien et la position de ce dernier, sont expliquées par des vitesses d'accrétion différentes, les chambres les plus larges s'établissant dans les rides les plus rapides (SLEEP et WOLERY, 1975 ; ROSENDAHL, 1976). Enfin, une inconnue persiste sur l'état d'écoulement du manteau à l'aplomb d'une dorsale selon que l'on considère ou non la rotation des lignes de flux (fig. II.1, voir aussi la 4ème partie).

De plus, les structures obtenues au niveau de la ride peuvent être localement interrompues par le jeu de failles transformantes (VOGT et JOHNSON, 1975), qui dans ce cas sont relativement passives vis-à-vis du magmatisme.

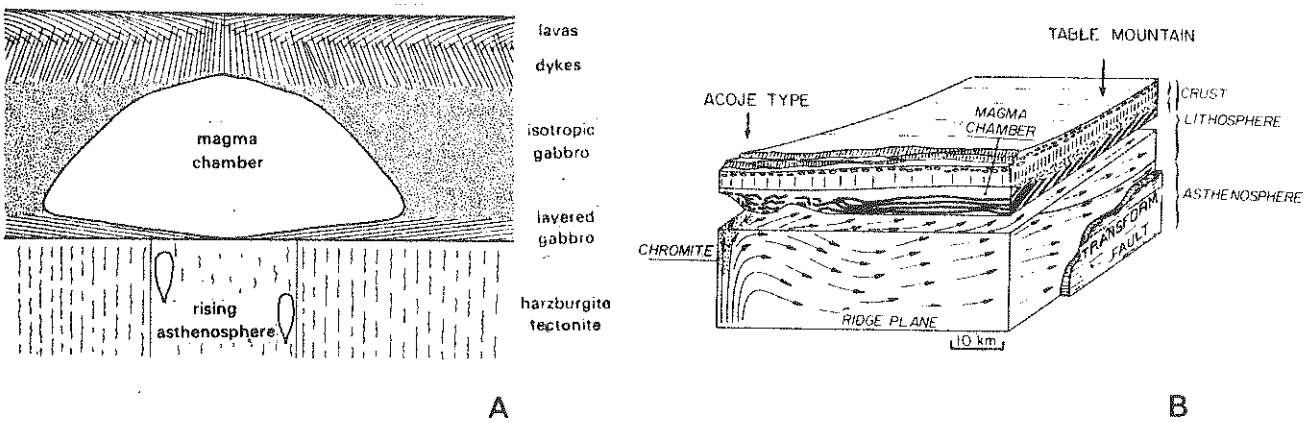


Fig. II.1 - Modèle d'accrétion océanique. A - Zone étroite d'injection, tiré de CANN (1974) ; noter la trace verticale du plan de fluage à l'écart de la ride. B - Divergence horizontale des lignes de flux à l'écart de la ride qui viennent buter sur les zones transformantes, tiré de NICOLAS et VIOLETTE (1982).

2 - Ouvertures obliques par rapport à la chaîne

Si l'on poursuit l'analogie avec l'océan, il existe pour des décalages de la ride de part et d'autre de la transformante, sur une distance inférieure à 25 km, une interpénétration des structures. A côté de fractures parallèles à la ride et à la transformante, il existe de très nombreuses failles obliques (fig. II.2A) qui occuperaient les plans de Riedel (WHITMARSH et LAUGHTON, 1975) soulignés par des escarpements (SEARLE, 1979). Dès lors, il est possible de concevoir qu'en profondeur l'écoulement du manteau subisse des modifications (VOGT et JOHNSON, 1975) se traduisant en surface par un axe d'extension oblique par rapport au mouvement des plaques (fig. II.2B, VOGT et JOHNSON, 1975) qui doit aussi imprimer une morphologie particulière aux chambres.

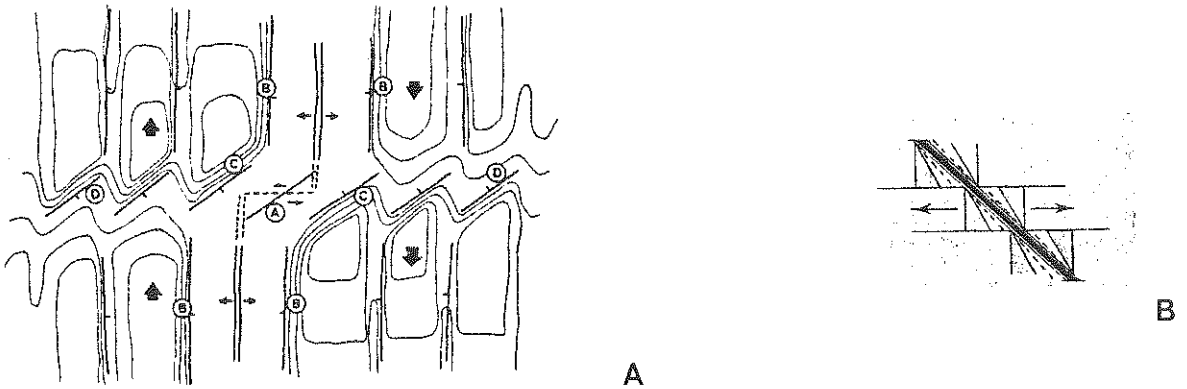


Fig. II.2 - A, Formation d'escarpements obliques, tiré de SEARLE (1979). Ⓐ, fractures de tension et failles normales à l'intersection de la ride et de la transformante ; Ⓑ, surrection parallèle à la ride ; Ⓒ, escarpement. En pointillé, une véritable transformante pour des rejets supérieur à 20 km, fig. II.2B, extension oblique par rapport au mouvement des plaques, tiré de VOGT et JOHNSON (1975).

Les accrétions obliques aboutissent à la formation de zones transformantes où le magma peut percoler ("leaky transform zone") au sein desquelles (fig. II.3) l'ouverture de fractures transverses est accompagnée d'intrusions de péridotites (THOMPSON et MELSON, 1972). Au Nord de l'Oman, la structure interne des ophiolites est compatible avec une succession de rides et de failles transformantes où le magmatisme est développé (SWEWING, 1980).

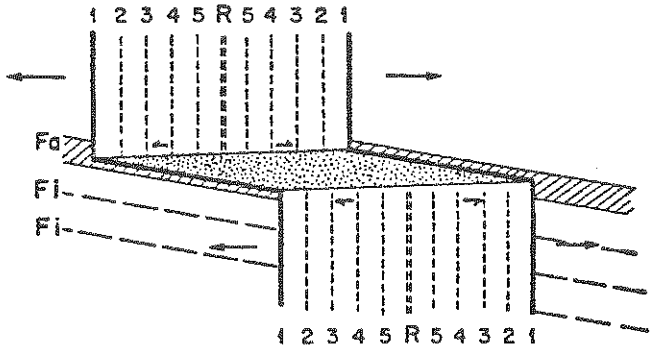


Fig. II.3 - Magmatisme dans les zones transformantes (en pointillé), tiré de THOMPSON et MELSON (1972). Les chiffres indiquent les stades de développement successifs. Fa est la direction d'une ancienne faille oblique sur la direction d'extension proposée ; Fi est inactive.

3 - Ouverture liée à des décrochements

Ce modèle d'organisation interne des ophiolites a été décrit précédemment (OHNENSTETTER M., 1980). Soulignons qu'il ne peut conduire qu'à des extensions limitées, favorisant la préservation des structures de diapirs dans les péridotites. En précisant ce type d'ouverture, deux cas peuvent être envisagés pour la position des zones distensives. Le jeu relatif des failles décrochantes (fig. II.4A) créerait des bassins par étirement (pull-apart basins), de sections rhombiques (EGUCHI et al., 1979 ; READING, 1980 ; GARFUNKEL et al., 1981) caractérisés par un axe d'extension sub-perpendiculaire au décrochement c'est-à-dire à la ceinture ophiolitique (fig. II.4B). Une extension locale plus prononcée permettrait de réunir plusieurs bassins juxtaposés dont l'aspect final s'approcherait d'une ouverture oblique (fig. II.4B).

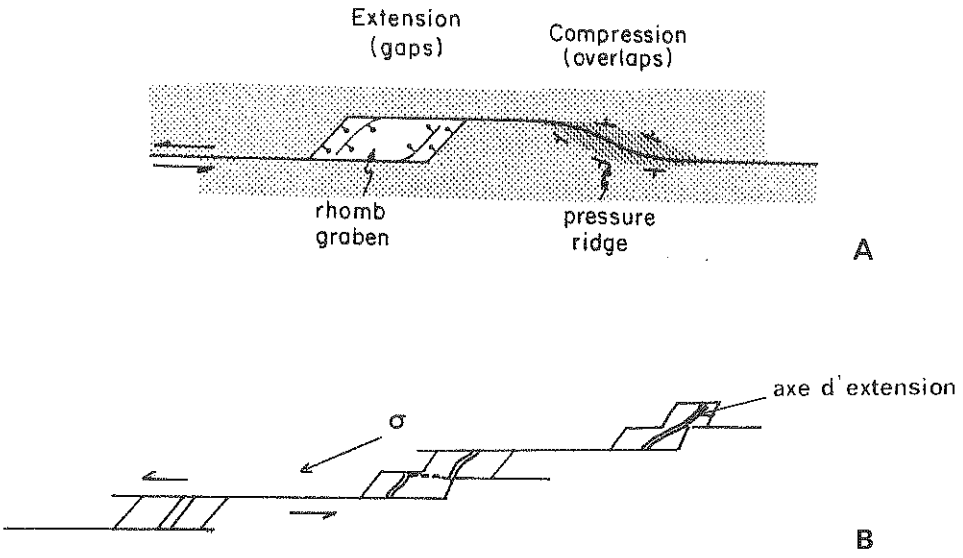


Fig. II.4 - Ouverture des bassins rhombiques liés au décrochement. A, schéma tiré de GARFUNKEL et al. (1981) montrant des zones de transpression et transtension. B, Evolution possible des bassins rhombiques ; noter l'axe d'extension fortement oblique par rapport à la chaîne.

Enfin, on peut concevoir que les ouvertures principales se fassent parallèlement à la contrainte maximum et aux failles normales, c'est-à-dire obliques (fig. II.5) sur la chaîne de montagne (OHNENSTETTER M., 1980). L'axe d'extension serait alors plus proche de la direction de la suture que dans le cas précédent.

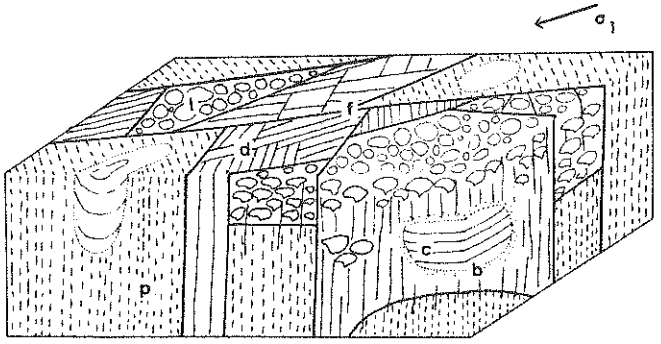


Fig. II.5 - Ouverture parallèle à la direction de la contrainte maximum dans un décrochement, tiré de OHNENSTETTER M. (1980) ; l, lave ; d, dyke ; f, faille ; c, chambre magmatique ; b, bordure de la chambre ; p, péridotite.

Ces décrochements peuvent jouer dans des sites géotectoniques variés (KOIDE et BHATTACHARJI, 1977 ; OHNENSTETTER M., 1980). Par exemple des décrochements sont invoqués dans une marge passive pour expliquer la présence de laves

et la remontée du manteau (TYSDAL et al., 1977). Ils seraient aussi à l'origine de la fragmentation des continents conduisant au rift et à l'océan (MOORE, 1979 ; GARFUNKEL et al., 1981). Dans les arcs (fig. II.6) ils préfigureraient la future zone de subduction (HILDE et al., 1977 ; SPÖRLI, 1980 ; PAGE et al., 1979) , tandis qu'à l'arrière de celui-ci, ils permettraient l'ouverture de bassins marginaux (EGUSHI et al., 1979).

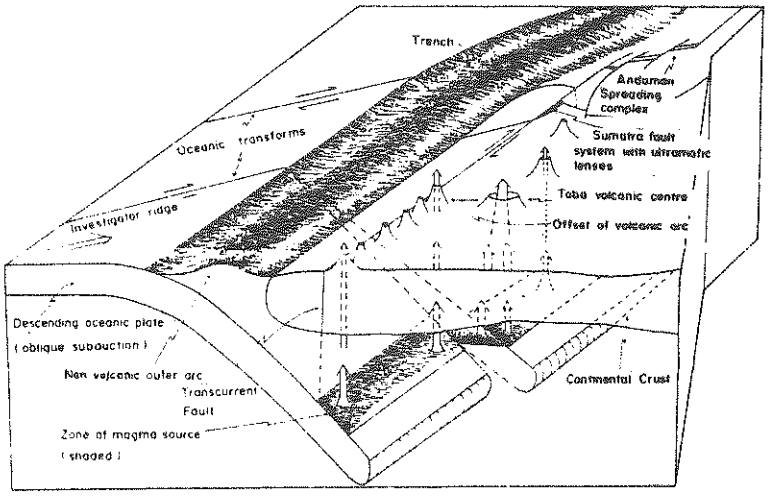


Fig. II.6 - Jeux de décrochement dans un système d'arc insulaire, tiré de PAGE et al., 1979. Noter le magmatisme le long des décrochements et la montée diapirique des péridotites (faille de Sumatra). Le bassin d'Andaman peut aussi résulter d'ouvertures rhombiques (EGUSHI et al. (1979)). On peut aussi constater la fermeture oblique à gauche et l'ouverture oblique à droite.

4 - Conclusion

De ce qui précède, on peut envisager que le système en extension favorise le développement d'un réseau de chambres magmatiques larges, régulières, associées à un complexe filonien aux directions bien réglées, parallèles à la ride ou à la suture. Au contraire, dès que l'extension devient oblique par suite de la superposition du jeu de failles parallèles aux rides et zones transverses, de la création de bassins de "pull-apart" ou encore du développement de zones distensives de même direction que la contrainte maximum, la position des repères structuraux devient plus compliquée. En surface, la direction des dykes est variable. En profondeur, cela peut se traduire, entre autre, par des contacts irréguliers entre cumulats et manteau, par la limitation du nombre et de la taille

des chambres magmatiques dont la morphologie est contrôlée par le jeu des failles synthétiques et antithétiques. Ces considérations ont conduit à proposer que chacun des deux groupes structuraux d'ophiolites soit respectivement caractéristique de systèmes se rapprochant du modèle en extension ou en décrochement (OHNENSTETTER M., 1980) : celui aux variations lithologiques sub-horizontales (Sud de l'Oman, Antalya) serait créé dans une ride en extension. Pour le second groupe à structure verticalisée (Chypre, Vourinos, Sabzevar) une origine dans une ceinture en décrochement conviendrait mieux.

Cependant, à partir des différentes conditions d'ouverture envisagées, on peut aussi concevoir que les ophiolites bien stratifiées doivent aussi leur existence à des jeux décrochants qui conduisent à la création de bassins rhombiques plus ou moins expansifs. Leur formation dans un bassin ou un océan est aussi en accord avec le fait qu'elles sont largement charriées sur la plate-forme continentale. Ainsi, leur structure interne, et leur caractère allochtone évident, les opposent aux ophiolites à structure verticalisée dont le vecteur de charriage serait plus court.

II - LA MISE EN PLACE SECONDAIRE DES OPHIOLITES

Dans une zone de suture, on est obligé de constater la réalité de l'empilement des nappes ophiolitiques au-dessus d'une croûte continentale par l'intermédiaire d'amphibolites et de zones de mélanges. Ce dispositif résulte du rapprochement des ophiolites par rapport à la marge, suivie de leur incorporation au continent et aux formations associées. Selon l'hypothèse de la tectonique des plaques, un grand nombre de modèles a pu être élaboré en fonction de la position du segment océanique par rapport au continent, à l'arc et aux marges. Comme précédemment, pour simplifier, on retiendra deux types d'approche : l'un implique des collisions frontales successives, l'autre une collision plus oblique.

1 - Collisions frontales

Dans le premier cas, le plus couramment décrit, de nombreuses variantes existent. D'une manière générale (fig. II.7) un processus de subduction entraînant la disparition progressive de la croûte océanique et le métamorphisme de celle-ci et des séries de marge associées est antérieur à l'obduction ou char-

riage de la croûte océanique sur l'avant-pays, contemporain de la mise en place des flyschs (HALL, 1976 ; WELLAND et MITCHELL, 1977 ; GEALEY, 1980). La subduction pourrait être intra-océanique (MATTAUER et PROUST, 1976) ou à la limite d'une croûte continentale et océanique. Elle pose le problème de la persistance de l'assemblage ophiolitique, de sa remontée, et de l'absence de magmatisme calco-alcalin correspondant. Quant à l'obduction, mécanisme imaginé par R. COLEMAN en 1971, des questions relatives au contraste de densité, nécessitent pour qu'elle ait lieu une surrection de la croûte océanique avant son charriage (DEWEY, 1976 ; COLEMAN, 1977). Pour annihiler cette difficulté, il est fréquemment proposé qu'un sous-charriage du matériel continental sous la croûte océanique est intervenu à la suite d'une subduction, ce qui conduirait au même empilement de nappes (MOORES, 1970 ; DAVIS et SMITH, 1971 ; MATTAUER et PROUST, 1976).

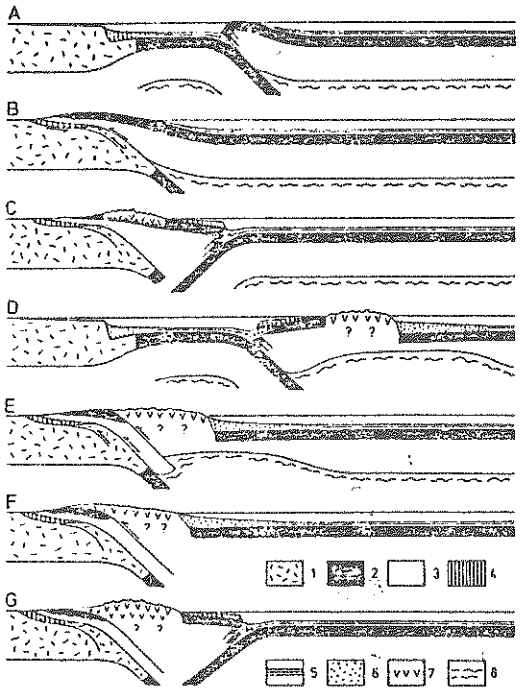


Fig. II.7 - Subduction et obduction tiré de DEWEY (1976). A,B,C, subduction partielle puis obduction sur la marge. D, E, F, G impliquent une collision avec un arc. 1, croûte continentale ; 2, croûte océanique ; 3, manteau ; 4, sédiments continentaux ; 5, sédiments océaniques ; 6, sédiments d'origine volcanique ; 7, volcanisme siliceux intermédiaire ; 8, zone à faible vitesse.

L'équilibre isostatique serait rétabli par un soulèvement général. Dans cette optique (fig. II.8), il est tentant d'assimiler les ophiolites à des re-

liques de bassins marginaux avant-arcs, charriées sur le continent par suite d'une collision entre marge passive et arc (GEALEY, 1977, 1980).

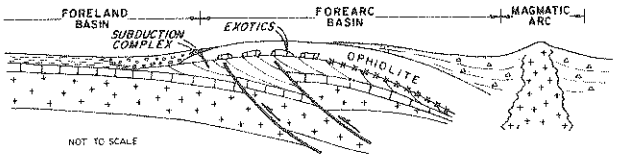


Fig. II.8 - Relation entre plan de subduction et d'obduction lors d'une collision entre arc et continent, tiré de GEALEY (1980). +, croûte continentale ; X, auréole métamorphique qui souligne le plan de subduction.

2 - Collisions obliques

Le deuxième modèle revient à considérer les secteurs en compression associés aux décrochements (BROOKFIELD, 1977) qui conduisent à des collisions obliques (BADHAM et HALLS, 1975) c'est-à-dire à des zones ni purement convergentes ni purement transformables (ANGELIER et al., 1982). L'existence de grands accidents longitudinaux qui favorisent l'écaillage des séries et les mélanges tectoniques est souvent reconnue à la fin de la collision (fig. II.9), postérieurement à la mise en place des nappes (BRUNN, 1961 ; PARROT et DUGAS, 1980 ; ROBERTSON et WOODCOCK, 1980). Ils favoriseraient la formation d'arcs induits

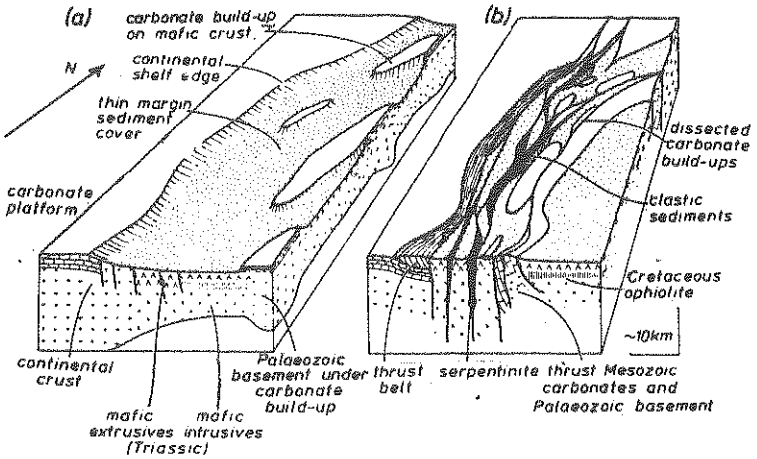


Fig. II.9 - Sites tectoniques dans le complexe d'Antalya, tiré de ROBERTSON et WOODCOCK (1980). a, développement de la marge passive au Mésozoïque ; b, décrochement provoquant la dissection parallèlement à la marge à la fin du Crétacé ou au début du Tertiaire.

(BRUNN, 1976 ; RICOU, 1978). Localement, ils peuvent être plus précoces : c'est le cas en Macédoine grecque où ils ont joué entre les phases tectoniques principales (MERCIER et VERGELEY, 1972). Enfin, on évoque souvent le rejeu d'accidents anciens intra-océaniques liés aux zones transformantes pour amorcer le décollement des futures nappes (KARSON et DEWEY, 1978 ; SALEEBY, 1979 ; WOODCOCK et ROBERTSON, 1982). Ces dernières hypothèses impliquent une forte obliquité de la ride par rapport à la marge continentale (fig. II.10) : voir à ce propos les schémas de KARSON et DEWEY (1978) ou ceux de SPRAY et RODDICK (1981).

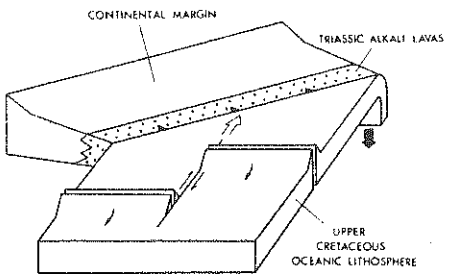


Fig. II.10 - Obliquité de la ride par rapport à la zone de subduction, tiré de SPRAY et RODDICK (1981). Ce modèle a été appliqué à Chypre pour une ride Est-Ouest.

Dès lors le problème de la précocité des nappes par rapport aux décrochements peut être soulevé. En effet, en fonction des irrégularités de la bordure des plaques, les accidents longitudinaux dans le cas de collision oblique, sont à l'origine du départ des nappes dont le vecteur de charriage sera généralement assez faible (fig. II.11). MOORES (1970) à propos de la Cordillère Ouest-américaine indique que raccourcissement et décrochement agissent plus ou moins simultanément. Dans certains arcs (N. Sumatra) la présence d'ophiolites est expliquée par un processus de "transduction" (fig. II.6) qui implique des décrochements importants (PAGE et al., 1979). Par ailleurs, de nombreuses études sur la collision soulignent qu'en approchant de cette zone, la déformation horizontale est absorbée par un réseau de failles longitudinales (NUR et BEN AVRAHAM, 1978). Dès lors, la reconnaissance tout au long d'un accident de plans de décollement à pendage variable pourrait traduire l'existence d'une collision oblique. En un même endroit, l'inclinaison du plan considéré pourrait aussi changer avec le temps.

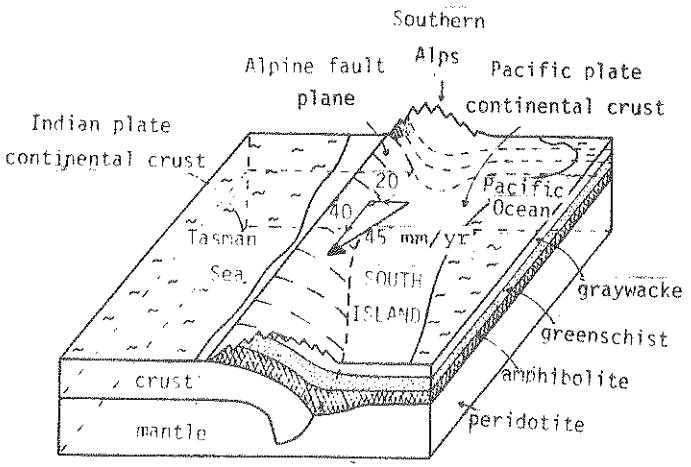


Fig. II.11 - Changement de raccourcissement crustal le long d'une suture en fonction du pendage du plan de chevauchement en Nouvelle-Zélande, tiré de ALLIS (1981). Le charriage est plus important au sud qu'au nord où l'érosion prédomine.

Ce système, certainement moins efficace quant à la disparition de la croûte océanique, permet, par le jeu d'irrégularités au niveau de la bordure, de favoriser le chevauchement des écailles denses de la croûte océanique. Par ailleurs, ce type de collision qui ne requiert ni subduction, ni obduction majeure (BROOKFIELD, 1977 ; HALL, 1980 ; OHNENSTETTER M., 1980) convient aussi à une taille limitée des bassins ophiolitiques.

3 - Conclusion

On peut dégager de ce qui précède que le jeu des décrochements peut être fondamental pour expliquer la mise en place secondaire des ophiolites et en particulier le départ des nappes. Les coulissages peuvent intervenir lors d'une modification du mouvement des plaques après une ouverture en extension ou décrochement. Dans ce dernier cas, les décrochements agiraient depuis l'ouverture des bassins ophiolitiques jusqu'à leur tectonisation avec des fragments continentaux. La position de la suture ophiolitique résultante correspondrait aussi à la patrie d'origine, une hypothèse qui semble se vérifier pour la plupart des sutures ophiolitiques grecques et pour certaines d'Iran (BEBIEN et al., 1980 ; II.2 ; BAROZ et al., à paraître, II.3). En considérant les larges massifs à plans structuraux et lithologiques sub-horizontaux, la suture observée peut être supposée plus ou

moins proche du bassin d'origine. Si les mécanismes d'extension puis de collision frontale peuvent aisément rendre compte de la disposition des nappes et de la géométrie interne des ophiolites, il n'est cependant pas exclu, même dans ce cas, que les coulissages aient, localement, favorisé la création de bassins et le départ des nappes. D'autres arguments indirects (chap. suivant) permettront de juger de l'importance et de la durée de l'extension avant la mise en place secondaire des témoins océaniques.

III - LES OPHIOLITES ET LES TERRAINS ASSOCIES

Les discussions à propos des mises en place primaires et secondaires mettent en évidence qu'un grand nombre de dispositifs structuraux dans des sites géotectoniques variés peuvent expliquer la présence d'ophiolites sur un continent. Cependant, à l'échelle d'une ceinture, pour juger de la validité d'un mécanisme par rapport à un autre, il semble utile de connaître la durée de l'expansion ainsi que le lieu de l'émission. Ces informations peuvent être obtenues indirectement par l'étude des terrains associés à la marge et à l'ophiolite. Comme la nature des formations antérieures, contemporaines et postérieures à l'ophiolite a été largement exposée à propos de la reconstitution des évènements géologiques dans les trois articles présentés (II.1. 2 et 3) on se contentera de dégager les points saillants propres à illustrer les fondements de notre conception sur la place des ophiolites dans une suture.

1 - Les terrains antérieurs

Les considérations précédentes ont souligné que les sutures ophiolitiques ne se tenaient pas toujours à proximité de leur lieu de création. En particulier, ceci conduit à considérer que plusieurs ceintures de roches vertes juxtaposées peuvent provenir d'une même patrie qu'il convient de rechercher. Ce problème peut être résolu par la reconnaissance des terrains antérieurs à l'ophiolite, c'est-à-dire ceux de la marge. Ils permettent de déceler le nombre de bassins subsidents qui ont été ensuite le siège du magmatisme ophiolitique (SMITH, 1973). D'après la paléogéographie des chaînes alpines, certains sillons auraient été le siège d'un magmatisme ophiolitique, isolant ainsi des môles à caractère plus continental tandis que d'autres bassins ensialiques seraient sans ophiolites (SENGÖR et MONOD, 1980).

La Téthys avait donc,à partir du Trias, plusieurs centres d'extension (DEWEY, 1977 ; OHNENSTETTER, 1980 ; BOULIN, 1981 ; SENGÖR et YILMAZ, 1981). La nature du volcanisme et des sédiments des marges est détaillée dans la première publication (ROCCI et al., 1980 ; II.1). On peut simplement préciser que les écarts de temps peuvent être très variables entre le début de l'extension et la venue des ophiolites (Permo-Trias-Jurassique, Trias-Jurassique supérieur, Trias-Crétacé moyen à supérieur).

Dans les cordillères qui bordent certains continents, les sutures juxtaposées sont de plus en plus jeunes vers leur périphérie. Ceci est particulièrement net dans le cas de la cordillère Nord américaine où des ophiolites du Permien, Trias, Jurassique et Crétacé se succèdent vers l'Océan Pacifique (MOORES, 1970 ; MAXWELL, 1974 ; IRWIN, 1977). Ce processus peut se retrouver, avec cependant moins de netteté, en Méditérranée orientale où les ophiolites du Sud de la chaîne sont les plus jeunes, de même que dans les sutures péri-indiennes (fig. II.12, BOULIN, 1981).

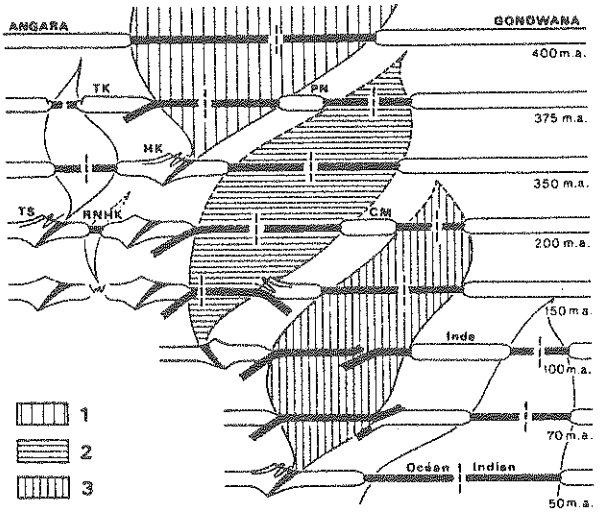


Fig. II.12 - Esquisse paléogéographique depuis le Paléozoïque, tiré de BOULIN (1981). évolution des espaces océaniques péri-indiens d'Afghanistan

1. Océan préhercynien. 2. Océan précimmérien. 3. Océan préalpin. TK, Turkestan ; PN, Monts de Paghman-Nurestan ; HK, Hindou Kouch ; TS, Tien Shan ; CM. Montagnes centrales d'Afghanistan; RNHK, Rift nord Hindou Kouch. N.B. l'espace pré-Tien Shan est interprété comme bassin marginal de l'océan préhercynien.

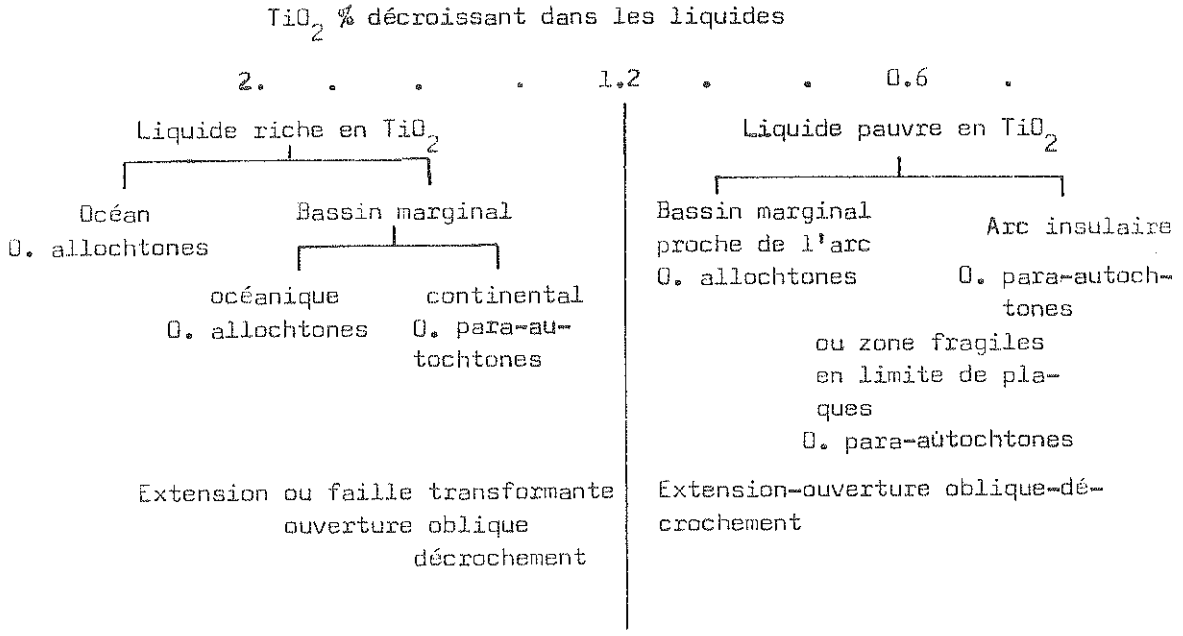
2 - Les ophiolites et sédiments associés

Lorsque les ophiolites se forment, on peut considérer, en plus de leur nature et de leur structure, la durée de l'accrétion, leur position dans le bassin et la nature des sédiments qui les recouvrent.

a) La composition des ophiolites

Dans la partie précédente, les discontinuités pétrologiques et structurales au sein d'une suture ou entre les sutures ont été soulignées. En particulier s'il est possible de séparer rapidement les complexes ophiolitiques dont le chimisme est caractéristique soit d'une zone d'accrétion médio-océanique, soit d'un environnement d'arc immature, il est plus difficile de préciser le lieu exact de formation : océan, bassin marginal en domaine océanique ou continental dans le premier cas, arc immature, bassin marginal à proximité de l'arc, "rifting" dans une zone mobile à l'intérieur ou en bordure des plaques dans le second cas. Le tableau II.1 résume les environnements géotectoniques possibles compte tenu des concentrations de TiO_2 dans les liquides : on peut envisager une multitude de sites ce qui est en accord avec la possibilité de remontée de l'asthénosphère dans les zones fragiles distensives indifféremment liées à des systèmes en extension ou en décrochement.

Tab. II.1 - Environnement géotectonique possible des ophiolites.
O : ophiolites.



b) La durée de l'accrétion

Celle-ci peut être estimée d'après les mesures d'âge absolu réalisées sur les ophiolites, les informations livrées par les sédiments supra-ophiolitiques donnant un âge postérieur à la fin du magmatisme. La différence calculée à partir des âges obtenus par les deux méthodes est cependant toujours faible. Par exemple, en Oman, les zircons extraits d'une coupe perpendiculaire à la chaîne donnent un âge compris entre 93.5 et 97.9 M.A. (TILTON et al., 1981) qui est proche de celui des données stratigraphiques (TIPPIT et al., 1981). Des remarques semblables peuvent être faites à partir de l'étude des ophiolites et de leur couverture à Chypre (VINE et al., 1973 ; DELALOYE et al., 1980) ou en Grèce (BEBIEN et al., II.2). Ces différentes données impliquent soit que le magmatisme des ophiolites de Méditérranée orientale fut de courte durée soit que la zone d'accrétion la plus jeune a été seulement préservée.

Cependant, en Corse, les radiolaires au contact des laves donnent un âge minimum du Malm (voir la bibliographie dans OHNENSTETTER et al., 1981 ; II.3) identique à celui connu dans les Apennins et plus récemment dans les Alpes (DE WEVER et CABY, 1981) tandis que les datations sur roches ou minéraux indiquent un âge légèrement plus ancien compris entre 185 et 142 M.A. Ceci a permis de proposer que l'expansion a pu localement durer 40 M.A. au maximum (OHNENSTETTER et al., 1981 ; III.3).

En Méditerranée orientale, l'âge des amphibolites, en écaille infra-péridotitique permet d'estimer indirectement la durée de l'accrétion si l'on considère que le métamorphisme principal qui les affecte résulte d'un dynamométamorphisme (WILLIAMS et SMITH, 1973 ; MALPAS et al., 1973) dû à l'emplacement du massif de péridotite encore chaud (PAMIC et al., 1973 ; KARAMATA, 1974, 1979 ; WOODCOCK et ROBERTSON, 1977 ; OKRUSCH et al., 1978 ; MALPAS, 1979 ; SPRAY et RODDICK, 1980). Les hypothèses quant à la nature de ces roches ont été discutées à propos de la Grèce (BEBIEN et al., 1980, II.2). Si leur nature, en particulier le caractère transitionnel à alcalin des métabasites les rapproche de séries volcano-sédimentaires de marge , d'autres hypothèses envisagent leur formation dans une zone de subduction intra-océanique (PARROT et WHITECHURCH, 1978 ; NICOLAS et LE PICHON, 1980). Enfin, un métamorphisme dans une zone transformante à proximité de la marge n'est pas exclu (SPRAY et RODDICK, 1981).

Tab.II.2

			ROCHE	MINERAL	METHODE	RESULTAT	AUTEUR
OMAN		écaille basale	amphibolite	hb phyllite	$^{40}Ar/^{39}Ar$ "	$90^{+}_{-}3$ $79.5^{+}_{-}3$	Lamphere (1981)
		ophiolite	plagiogranite	zircon	U/Pb	93.5-97.9	Tilton et al.(1981)
CHYPRE		écaille basale	amphibolite	hb	$^{40}Ar/^{39}Ar$	$83-90^{+}_{-}3$	Spray et Roddick (1981)
		ophiolite	lave filons	roche "	K/Ar "	$75.3^{+}_{-}5.4$ 83 ± 3	Delaloye et Desmet (1979) Delaloye et al. (1980)
TURQUIE		écaille basale	amphibolite	hb,pl,mus-covite	K/Ar	$102^{+}_{-}7$ -$83^{+}_{-}3$	Thuizat et al. (1981)
		ophiolite	gabbro diorite	pl	"	$\sim$ 85	Thuizat et al. (1979)
		"	dolérites en filons		"	70-75	" " "
TERRE NEUVE	Bay of Islands	écaille	amphibolite	amph.	$^{40}Ar/^{39}Ar$	$458^{+}_{-}14$-$548^{+}_{-}16$	Archibald et Farrar (1976)
		ophiolite	Q-diorite	amph.	"	452 ± 12	" " "
	Little Port	écaille	amphibolite	"	"	$477^{+}_{-}13$-$457^{+}_{-}13$	" " "
		ophiolite	Q-diorite	"	"	$469^{+}_{-}13$-$465^{+}_{-}15$	" " "

En confrontant l'âge du métamorphisme des amphibolites à celui des ophiolites (Tab. II.2), on constate que ces âges sont voisins pour les massifs de Méditérranée Orientale comme en Oman, à Chypre et dans les Taurides ou pour les ophiolites paléozoïques de Terre-Neuve. 20 à 30 MA suivant les cas séparent l'âge de formation des amphibolites de la mise en place secondaire des ophiolites sur le continent (SPRAY et RODDICK, 1980 ; BEBIEN et al., 1980, II.2 ; THUIZAT et al., 1981).

En conséquence, si le métamorphisme des amphibolites est lié à la mise en place primaire des ophiolites, une hypothèse confortée par le métamorphisme de contact observé au Zagros entre péridotites et sediments (RICOU, 1971 ; HALL, 1981) (les paragenèses de skarns à fassaïte, wollastonite et calcite (HALL,1981), indiquent des températures de 800ºC pour une pression de 2 Kbar) la montée du diapir mantellique a eu lieu à proximité de la marge continentale, ce qui est aussi en accord avec la création de bassins étroits.

3 - Les formations post-ophiolitiques

Leur intérêt réside dans le fait qu'elles donnent des indications sur la façon dont la zone à ophiolites a évolué avec le temps. En considérant leur âge et leur nature de part et d'autre des discontinuités angulaires, ces séries per-

mettent aussi de suivre la mise en place secondaire des ophiolites.

Les sédiments qui recouvrent les laves ophiolitiques ne sont pas homogènes que ce soit entre les sutures ou au sein de celles-ci. La présence de radiolarites, non systématique, semble indépendante de la présence d'ophiolites (MARCOUX et RICOU, 1979 ; RICOU et MARCOUX, 1980). Au sein d'une suture comme en Corse, leur absence peut traduire le caractère proximal des séries par rapport à la marge (CARON et DELCEY, 1979). L'évolution d'une série sédimentaire avec le temps montre que très rapidement la bathymétrie change en raison du caractère instable de la zone. Dans le massif du Vourinos, séries pélagiques et néritiques alternent (PICHON et LYS, 1976) tandis qu'à Terre-Neuve, des grauwackes recouvrent les ophiolites (CHURCH, 1972). L'importance du caractère néritique des séries supra-ophiolitiques peu après leur création a été soulignée par BRUNN (1980). Comparés aux séries sédimentaires des océans actuels,les sédiments des ophiolites sont généralement plus siliceux en raison de l'étroitesse des bassins (ROBERTSON, 1981). Par ailleurs selon ce même auteur les séries de Méditerranée Occidentale seraient plus riches en Mn et plus pauvres en sulfures comparées à celles de Méditerranée Orientale.

La surrection rapide des zones à ophiolites annonce l'arrivée des flyschs à olistolithes comblant le bassin et souvent liés à l'avancée des nappes. La tectonisation de ces différentes séries conduit à la formation de mélanges.

Quelquefois un arc insulaire s'établit sur le substratum ophiolitique (CHURCH, 1973 ; BAROZ et al., II.3). Dans le cas de l'Iran, les séries volcano-sédimentaires calco-alcalines ne sont pas séparées de plus de 10 MA de la formation de l'ophiolite compte tenu des erreurs analytiques (BAROZ et al., II.3). D'une manière plus générale, les séquences ophiolitiques sont souvent recoupées par des intrusions granitiques plus récentes liées à un magmatisme calco-alcalin d'un arc insulaire développé en milieu océanique (RODGERS, 1973 ; JACQUES et CHAPPEL, 1980 ; DUPUY et al., 1981) ou plus continental (ABOTTS, 1978).

Ce décalage dans le temps entre intrusion acide et formation de l'ophiolite n'est pas toujours très important. C'est le cas en Grèce Orientale et en Yougoslavie où les deux activités magmatiques sont contemporaines (ROLLET, 1969 ; BEBIEN et MERCIER, 1977 ; BEBIEN, 1982). Par ailleurs dans ces ophiolites de Grèce ou au Chili, l'absence de péridotites en surface, l'intrusion de laves basiques dans des séries rhyolitiques (SUAREZ, 1979 ; BEBIEN, 1981) sont en accord

avec une création d'ophiolites ensialiques où la participation continentale serait plus importante (DE WIT et STERN, 1981). Leur extension limitée dans le temps conduirait à des massifs para-autochtones.

Bien que n'ayant pas personnellement travaillé sur les formations associées aux ophiolites, j'ai néanmoins étudié avec attention les travaux se rapportant à leur origine. En effet, elle permet de discuter indirectement l'environnement géotectonique des ophiolites. Une compilation bibliographique a permis d'apporter quelques renseignements sur le nombre de sillons possibles dans une chaîne, la durée et le lieu d'accrétion par rapport au continent, la mobilité tectonique de la zone d'émission et son évolution une fois le magmatisme ophiolitique achevé. Certaines des conclusions seront rappelées brièvement au début du chapitre suivant afin de les intégrer aux propos sur l'orogène.

IV - VUE SCHEMATIQUE DE L'OROGENE A TRAVERS L'HISTOIRE DE L'ACCRETION OPHIOLITIQUE

1 - Rappel de quelques données fondamentales

Les considérations sur la mise en place primaire des ophiolites indiquent que d'un point de vue structural, un grand nombre de sites géotectoniques sont possibles pour la création d'ophiolites : ride médio-océanique, arc, bassin marginal plus ou moins proche d'une zone de subduction peuvent être retenus suivant les cas, après examen de la composition des liquides. Cette apparente diversité dans la constitution des ophiolites et des environnements géotectoniques qui en résultent ne doit pas masquer l'ubiquité de certains de leurs caractères de même que celui des formations associées. Par exemple, la durée de l'accrétion, et l'écart de temps entre la mise en place primaire et secondaire (maximum de 30 MA) sont particulièrement brefs pour les ophiolites de Méditerranée orientale. En Méditerranée occidentale l'accrétion aurait pu se poursuivre pendant 40 MA. Cette durée est particulièrement critique lors de l'expansion d'une ride de type Atlantique. En effet, d'après l'épaisseur de la croûte océanique révélée par des études de géophysique, ROOTS et al. (1979) ont déduit que le système de ride atteint son équilibre après 40 MA environ, une fois que la croûte s'est amincit exponentiellement depuis le début de l'accrétion. En conséquence, seuls quelques fragments des ophiolites de Méditerranée occidentale pourraient avoir les caractères d'une ride normale. Pour les fragments les plus précoces de cette suture

et pour la plupart des autres ophiolites, une certaine prudence doit s'imposer lors de l'établissement d'hypothétiques analogies entre ride médio-océanique et ophiolites. Dans les ophiolites du type de Méditerranée orientale, une expansion de courte durée est aussi en accord avec le dynamométamorphisme des séries volcano-sédimentaires des marges lors de l'intrusion des péridotites à haute température. Celle-ci aurait lieu dans une marge mobile, devenant de plus en plus instable et superficielle avec le temps comme l'indique la nature des sédiments supra-ophiolitiques.

Ainsi l'apparition des ophiolites interviendrait dans un bassin lié à l'histoire d'une marge continentale formée lors de l'amincissement, puis de la rupture d'une plaque continentale. La répétition dans le temps et dans l'espace de ce processus aboutit à l'individualisation de plusieurs bassins ophiolitiques devenant plus tard les sutures ophiolitiques

2 - Développement de l'orogène

Si l'évolution de nombreux bassins de courte durée conduit à l'apparition des ophiolites puis à leur mise en place secondaire sur la marge continentale, il apparaît exceptionnel que le jeu des plaques se résume à la succession extension-compression. Un mode de représentation (fig. II.13) permet de schématiser de nombreux mécanismes tectoniques successifs qui doivent pouvoir convenir à la plupart des orogènes. Tel que le schéma se présente, les autres alternatives au chemin 1 impliquent l'intervention de coulissage avant, pendant ou après la mise en place primaire et secondaire des ophiolites. Quelques possibilités sont indiquées dans la légende de la fi. II.13. Le cas le plus général correspond à un coulissage qui dure plus ou moins longtemps avant que les plaques entrent en collision. En modulant la durée de l'extension on peut créer des bassins de taille variable qui viendront progressivement chevaucher la marge lors d'une collision frontale ou oblique.

Pour illustrer ce propos, on peut concevoir, pour les ophiolites dont l'expansion a duré assez longtemps comme en Corse, le schéma suivant : coulissage pour créer le rift, au moins à partir de la fin du Lias, extension avec rides et zones transformantes (180-140 MA), cessation de l'activité magmatique au Malm par suite de l'ouverture de l'Atlantique Nord, compression conduisant au rapprochement des plaques océaniques et continentales, au moins à partir du

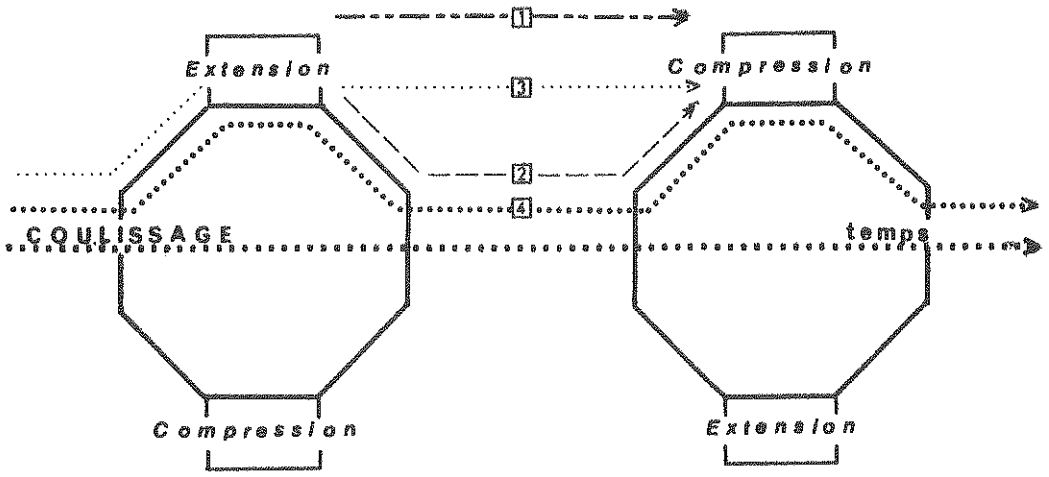


Fig. II.13 - Esquisse de la succession des évènements tectoniques conduisant à la mise en place primaire et secondaire lors d'ouverture et de fermeture plus ou moins obliques par rapport à la bordure de la marge. 1, modèle généralement admis : extension-collision frontale ; 2, extension puis collision oblique ; 3, ouverture oblique avec collision frontale ; 4, modèle général pour le développement complet d'un orogène aboutissant à l'accrétion succesive de fragments continentaux et océaniques à la marge d'un continent : ouverture oblique suivie d'une extension plus ou moins longue, puis fermeture oblique.

Crétacé moyen (CARON, 1977 ; COHEN et al., 1981) et à l'incorporation tectonique des fragments ophiolitiques sur le continent au Crétacé supérieur et au Tertiaire. Dans ce cas, il est possible de distinguer une tentative d'océanisation en liaison avec l'ouverture de l'Atlantique Central puis la cessation de cette activité par suite de l'apparition de l'Atlantique Nord. Lors du "rifting", la tectonique superficielle en distension sur un long linéament aurait facilité la remontée de l'asthénosphère et l'écoulement superficiel de la lithosphère.

Pour les ophiolites de la Méditerranée orientale, dont l'histoire magmatique fut plus brève, un système de coulissage s'impose, en particulier pour les ophiolites à structure verticalisée. On peut concevoir dans la même suture que certaines zones qui ont été plus longtemps en extension conduisent à des ophiolites bien stratifiées dans les bassins rhombiques. De la même façon, des zones qui ont subi une composante compressive plus importante peuvent apparaître plus tectonisées (zone de mélange, juxtaposition de blocs continentaux et océaniques). Enfin, une même chaîne peut connaître la répétition de ces évènements tectoniques

ce qui amène à la juxtaposition d'assemblages ophiolitiques d'âge différent localement séparés par des môles continentaux et des formations volcano-sédimentaires de marge : c'est le cas de la Cordillère Nord Américaine où le vecteur temps joue plus que le vecteur espace (fig. II.14A).Si le coulissage est induit par un poinçonnement (PELTZER et al., 1982) l'ouverture de bassins successifs peut avoir lieu le long de plusieurs accidents quelquefois fort éloignés les uns des autres comme dans la suture de l'Indus (BOULIN, 1981). Dans ce cas les vec-

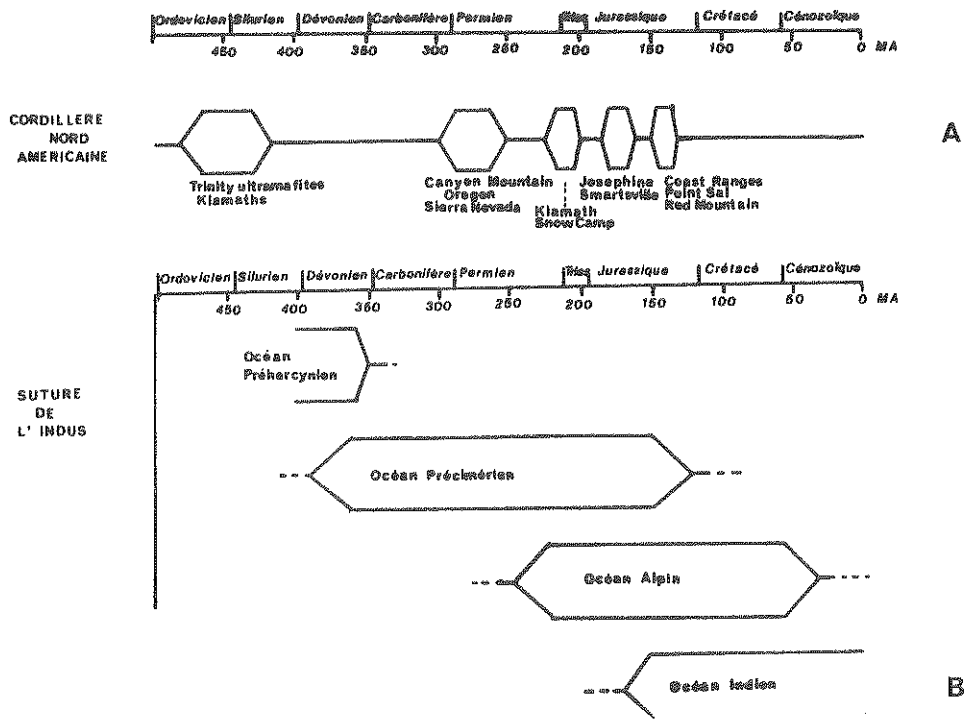


Fig. II.14 - Schéma des successions d'orogènes selon leur position dans le temps et l'espace ;A , modèle applicable à la Cordillère Nord Américaine ;B , modèle applicable à la Méditerranée orientale et à l'Indus.

teure temps et espace sont importants (fig. II.14B). Ce dernier aspect souligne que l'ouverture des bassins ophiolitiques intervientlors du déroulement d'un orogène. Leur création résulte de la collision et de la convergence plutôt que d'un épisode majeur de distension. En cela le magmatisme ophiolitique se déroule plutôt en bordure de continents qu'au sein d'océan entrant en subduction. Ces considérations rejoignent celles d'AUBOUIN (1982) pour qui la mise en place des couronnes péri-australienne et péri-asiatique constitue le modèle paléogéographique principal des chaînes alpines : les ophiolites seraient créées à la

marge d'un continent plutôt qu'au voisinage de zones de subduction telles qu'elles se présentent actuellement dans l'Ouest Pacifique. Dès lors, les ophiolites ne jouent pas un rôle passif par rapport à l'orogène mais actif.

Leur évolution inéluctable dans le temps -le diapir de péridotite n'approche de la surface que lorsque le bassin sédimentaire préexistant a atteint une certaine maturité- l'est aussi du point de vue magmatique. L'état d'équilibre est rarement atteint, sauf quelquefois dans les grands massifs ophiolitiques, et la composition des magmas évolue avec le temps au contraire des systèmes de ride médio-océanique. Le caractère épisodique du magmatisme intra-ophiolitique rappelle plutôt celui des arcs et bassins juxtaposés (RAYMOND et SWANSON, 1980). La présence locale de magmatisme calco-alcalin lié au développement d'un arc renforce l'analogie : cas de la chaîne de Sabzevar (BAROZ et al., II.3).

En résumé, si certaines ophiolites constituent la trace de tentatives d'océanisation comme en Méditerranée Occidentale, l'histoire de la plupart des autres notamment en Méditerranée Orientale coïncide avec le développement d'une marge passive puis active en bordure des continents majeurs où les mouvements transverses sont largement dominants. Ces derniers peuvent être induits par la dérive des plaques et surtout par la convergence des boucliers continentaux.

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The Mediterranean ophiolites and their related Mesozoic volcano-sedimentary sequences

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Abstract.

Despite some differences in their petrology and stratigraphy Mediterranean ophiolites were once considered to have a common origin. But their different ages and their petrological and chemical characters suggest that they do not belong to a single unit. Two main groups are distinguished here. The first group, represented in the Western Alps, the Apennines, Corsica and Calabria, are very similar to the formations described from present ocean floors. The ophiolites of the second group, including most of the Hellenic, Dinaric, Turkish, Cypriot and Syrian ophiolites are much more diversified and their origin is still discussed.

Both sets of ophiolites can also be related to their volcano-sedimentary sequences which are interpreted in terms of the paleogeographic evolution of the Mediterranean areas. In Western Mediterranean (the Alps, Corsica, the Apennines, Calabria), the pre-ophiolite magmatism consists of intrusions of peralkalic granites related to acidic and basic flows. Within the Middle and Upper Trias, alkalic igneous complexes and metabasalts are found. The post ophiolite volcanism consists of tholeiitic flows of Upper Cretaceous age related to the opening of the Valaisan trough.

The central area (Dolomites, Dinarides and Hellenides) is characterized by almost continuous volcanism from Permian to Recent e.g. Permian alkalic and tholeiitic basaltic and rhyolitic flows in Greece, Aniso-Ladinian volcanism, consisting of basalts, andesites, and acid lavas interbedded with pelagic or semi-pelagic sediments. The volcanism, contemporaneous to the ophiolite, outcrops only in the Vardar zone and consists of acid and basic flows. Basalts are either island arc or oceanic tholeiites. The calc-alkalic volcanism of Cretaceous – Eocene age follows the tectonic emplacement of the ophiolites which took place at the end of the Jurassic.

The third area is characterized by important Upper Triassic volcanism, the lack of magmatic activity during the Jurassic, and Middle to Upper-Cretaceous ophiolites. A highly differentiated alkalic volcanism is related to a continental rift system. Ophiolite volcanism is not always present. The post-ophiolite volcanism is known only in Northern Cyprus with both calc-alkalic and shoshonitic trends.

Middle-Upper Trias, Middle-Upper Jurassic and Upper Cretaceous paleogeographic maps illustrate the Mesozoic evolution of the Mediterranean areas. Rifting occurred in Upper Trias both in the Western and Eastern Mediterranean. In the Western area, sea floor-spreading follows immediately during the Jurassic. On the contrary in the East, oceanic basins are only of Middle to Upper-Cretaceous age. Subduction occurred in the Dinaro-Hellenic Range but may be probably absent in the Western Alps, Corsica and Apennines.

Introduction

A reconstruction of the Mesozoic paleogeographic evolution of the Tethyan area requires good knowledge of the location and the evolution of oceanic basins (oceans – marginal seas) present at that time. A study of the ophiolite suites which presumably represent ocean floors to these basins is therefore of great interest. However the geodynamic evolution of these oceanic basins is also recorded in their margins, where the structural setting of these basins and the different stages of their evolution (opening, development and closing) can be determined.

Among the formations geographically associated with the ophiolites and considered to be linked to their evolution, the volcano-sedimentary formations, abundant in all the Mediterranean area, are of special interest.

The different ages and the petrological characters of the ophiolites show that they do not belong to a single type. On one hand, the ages lead us to distinguish the Jurassic ophiolites of the Western Alps, Hellenides and Dinarides from the Cretaceous ones of Southern Turkey, Cyprus and Syria. On the other hand, petrological and geochemical data allow one to distinguish two main groups: the ophiolites of the first group, represented in the Western Alps, in the Apennines, Corsica and Calabria, are very similar to rocks from present ocean floors. The ophiolites of the second group are much more diversified and differ from the first one in various ways. Their origin is still a subject of debate. This second group includes most of the Hellenic, Dinaric, Turkish, Cypriot and Syrian ophiolites. This duality of the Mediterranean ophiolites has already been described in detail (Rocci et al., 1975).

Three geographic areas are discussed in this paper.

1. The Western Alps, Corsica and Calabria (Jurassic ophiolites belonging to the first group)
2. The Dolomites, Dinarides and Hellenides (Jurassic ophiolites belonging to the second group)
3. Southern Turkey, Cyprus and NW Syria (Cretaceous ophiolites belonging to the second group).

In every area, the pre-ophiolite, the contemporaneous and the post ophiolite magmatism will be considered.

Western Alps, Apennines, Corsica and Calabria

The magmatism related to the ophiolites consists mainly of pre-ophiolite sequences of Permian and Triassic age and a post-ophiolite one of Cretaceous age (Fig. 1).

The pre-ophiolite magmatism

During the Permian, alkalic granites intruded the two Hercynian calc-alkalic granite basements (Orsini, 1976). These different anorogenic granites shown in figure 2a modified after Bonin (1977) have been dated with Rb/Sr, K/Ar and $^{39}Ar/^{40}Ar$ by Bonin et al. (1972, 1978), Maluski (1974, 1976), Maluski and Lancelot (1976), Baubron et al.

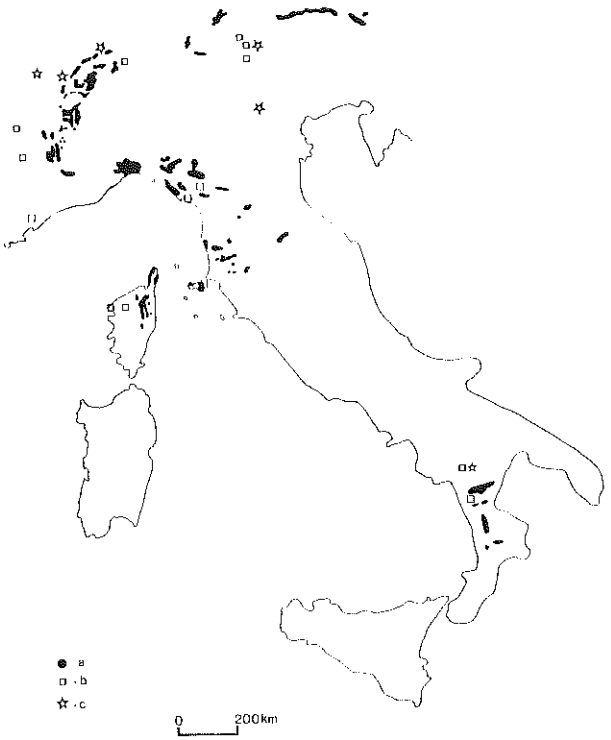

Fig. 1 *Distribution of the ophiolites and the pre- and post-ophiolite volcano-sedimentary formations in western Mediterranean areas.*

Legend

a- ophiolites

b- pre-ophiolite volcano-sedimentary sequence

c- post-ophiolite volcano-sedimentary sequence

(1975). The ages range from 262 ± 13 my to 194 ± 16 my.

Ignimbrites, rhyolites and olivine-basalts related to these granite ring complexes outcrop in Corsica, in the Esterel (SE France) and in the Apuan Alps (Italy) (Routhier and Meunier 1962, Barberi 1966, Bortolotti and Passerini 1970, Boucarut 1971, Brisset et al. 1974, Vellutini 1977). During the Upper Trias, metabasalt flows were interbedded in sandstones and limestones (Fig. 2b) in the French Alps (Alsac 1961, Tane 1961, 1967, Aumaître and Buffet 1973, Vatin-Pérignon et al. 1972, 1974). The alkalic metabasalts contain peridotite nodules. This volcanic event is associated with shallow shelf and lagoon deposits. Similar formations outcrop in the Apennines (Elter and Federici 1964, Ricci 1968, Baldaci and Cerrina-Feroni 1971, Ricci and Serri 1975) and in Calabria. In this latter region, tuffites are interbedded in *Halobia* cherty limestones (Dietrich and Scandone 1972) of Carnian age. In the Eastern Alps, the alkalic igneous complex of Predazzo, (Lucchini and Morten, 1977) ranging from gabbro to granite, intruded the carbonate platform formations during the Ladinian – Carnian.

These different magmatisms may be related to thermal events occurring in the Permian (Hunziker 1970) and in the Upper Trias (Ferrara and Innocenti 1974). These extrusive and intrusive rocks have been interpreted as the first stage of magmatic activity of a rift system in a sialic environment.

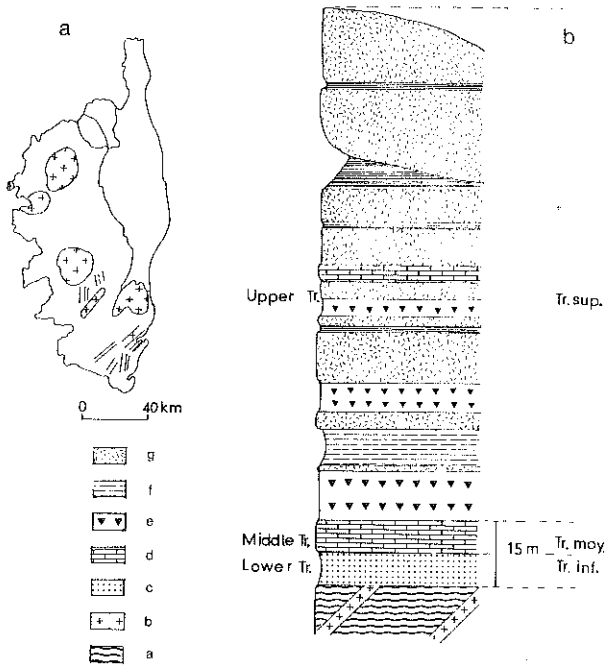

Fig. 2a *Distribution of Permian intrusive alkalic granites in Corsica (after Bonin, 1977).*

Fig. 2b *Diagrammatic section of Triassic volcano-sedimentary series in French Western Alps. From Aumaitre and Buffet (1973).*

Legend

a- substratum

b- intrusive basaltic dykes

c- fine grained sandstones

d- dolomitic limestones

e- volcanic breccia

f- clays

g- metabasaltic flow

The ophiolite sequence

The sea-floor spreading occurred during the whole Jurassic period, from Lias (180 my) to Malm (140 my). Interruption of sedimentation (Delcey et al. 1964; Grandjacquet et al. 1972) and block faulting with horsts and grabens are related to the formation of the ophiolites (Elter et al. 1966, Michard 1967, Elter 1971, Baud and Masson 1975). Ophiolites in the Alps, Apennines and Corsica are dated mainly from 190 to 140 my (Puszaszeri 1969, Bertrand 1970, Bigazzi et al. 1972, Bigazzi et al. 1973, Bertrand and Delaloye 1976, Ohnenstetter et al. 1977, 1979).

The ophiolite sequence of Corsica and the Apennines is from base to top: (Ohnenstetter and Ohnenstetter 1975, Ohnenstetter et al. 1975; Beccaluva et al. 1976, 1977; Piccardo 1977; Beccaluva and Piccardo 1978, Cottin 1978) (Fig. 3).

1. Lherzolite, mainly deformed, associated with minor clinopyroxene poor lherzolites, harzburgites and dunites. No chromite deposits.
2. The gabbroic cumulate sequence starting with troctolite to gabbro and ferrogabbro. Rare albite-granites sometimes intrude the volcanic sequence. They are often related to the upper part of the gabbroic sequence and have probably been formed by liquid immiscibility (Ohnenstetter et al. 1979).
3. The volcanic and hypabyssal rocks consist of massive dolerites, rare dykes and pillow-lava flows. They are petrologically and chemically similar to the mid-oceanic tholeiites.

The sedimentary cover begins with radiolarian cherts,

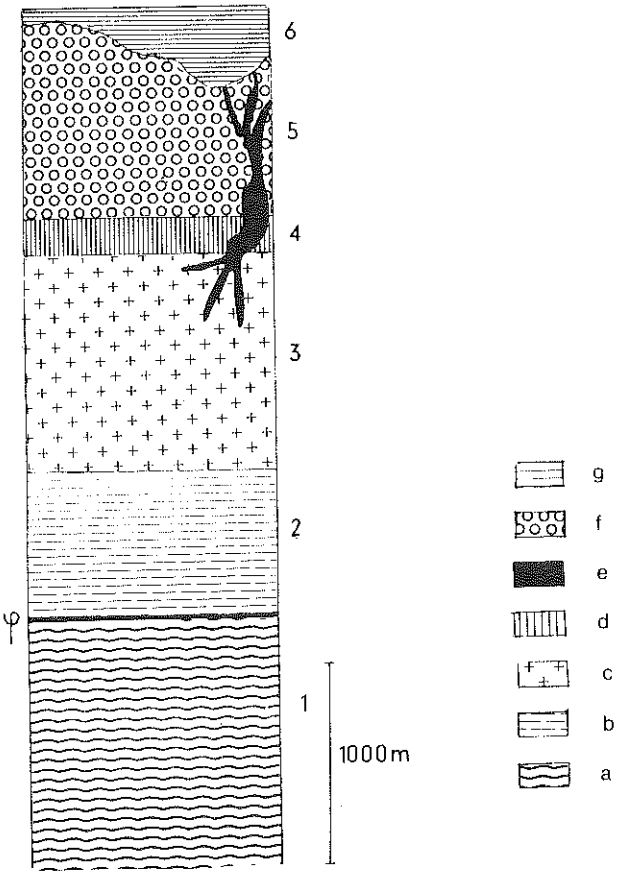

Fig. 3a *Diagrammatic section of the Corsican ophiolites*

Legend

a- metamorphic peridotite (lherzolites)
b- layered cumulate plutonic sequence
c- gabbros and ferro gabbros
d- hypabyssal rocks
e- plagiogranites: albitites
f- volcanic pile
g- sedimentary cover (radiolarian cherts)

associated with *Calpionella* Malm limestones, followed by the Lower Cretaceous Palombini shales and sandstone beds with clayey interlayers of Middle to Upper Cretaceous age (Lemoine 1971, Haccard and Lemoine 1970, Decandia and Elter 1972, Amaudric du Chaffaut et al. 1972, Caron 1977).

The whole sequence suggests the existence of an oceanic trough, the evolution of which is related to a passive continental margin (Bourbon et al. 1977, Laubscher and Bernouilli 1977).

The post-ophiolite volcanism

In the Valaisan domain, the post-ophiolite volcanism is of Cretaceous age. In the Visp valley, Dietrich and Oberhansli (1975) mention tholeiitic metabasaltic sills and pillow-lavas. In the Versoyern, Lasserre and Laverne (1976) and Lasserre and Vatin-Pérignon (1977) describe pillow-lava flows and highly differentiated tholeiitic sills of

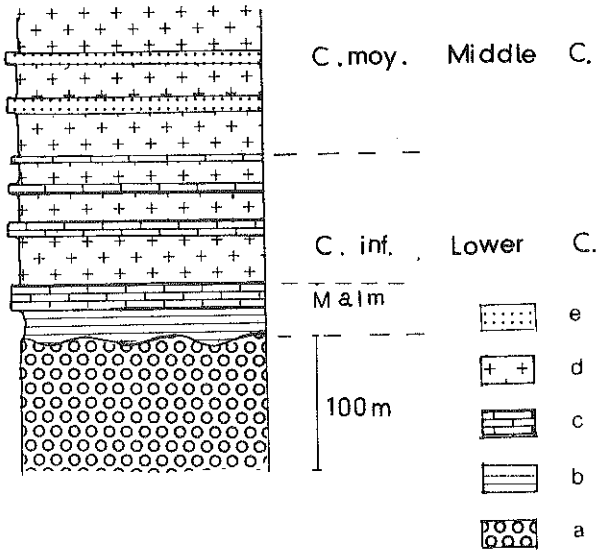

Fig. 3b *Diagrammatic section of the post ophiolite volcano-sedimentary cover in the Liguro-Piemontaise area (Chabriere Alps, Inzecca, Corsica, Bracco Apennines) after Lemoine et al. 1970; Decandia and Elter 1972; Amaudric du Chaffaut et al. 1972 and Caron 1977.*

Legend

a- ophiolitic substratum (mainly tholeiitic, basaltic pillow-lavas and pillow breccias)
b- radiolarian cherts
c- marmorean limestones (Calpionella limestones in Apenniines)
d- Schists and limestones of the Replatte or Erbejolo formation (metamorphic equivalent of Palombini shales)
e- schists and quartzites of the Roche Noire formation (equivalent of pp. Val Lavagua schist)

Upper Cretaceous (Senonian) age. These volcanics rest on limestones and are covered by Upper Cretaceous flysch (Antoine et al. 1973). They are related to the opening of the Valaisan trough.

During the same period, tectonic emplacement of ophiolites was accompanied by blueschist metamorphism. This high pressure event is dated at 100-78 my (Bocquet et al. 1974). It should be noted that if subduction occurred during the Upper Cretaceous and perhaps in the Eocene, there is no significant volcanism related to the supposed plate consumption. Rare andesitic debris is reported from the unmetamorphosed flysch succession (Aptian to Middle Eocene) on the Lucania – Calabria boundary (Lanzafame et al. 1977). Similar levels outcrop in the Alps and Apennines (Martini 1968; Elter et al. 1969; Barberi et al. 1971; Sawatzki 1975). The existence of dykes of calc-alkalic composition dated by K/Ar from Upper Cretaceous to Tertiary has been related to subduction processes (Gatto et al. 1976). Paleocene to Upper Oligocene trachytes and rhyolites occur in the Venetian (De Vecchi et al. 1976).

Dolomites, Dinarides and Hellenides

The importance of the ophiolite belts in the Dinaro-Hellenic range is well known. These areas as well as the Dolomites, are characterized by an almost continuous ophiolite-free volcanism from the Trias till Recent.

The pre-ophiolite magmatism

Basaltic and rhyolitic flows associated with Permian detritic sediments occur in Northern Greece, on the border of the Serbo-Macedonian massif, (Mercier 1966) and in Attica and Beotica, (Clément 1976; Clément and Lapierre 1979). The basalts present either tholeiitic or alkalic trends.

Volcano-sedimentary formations of Middle Triassic age outcrop in all the Dinaric-Hellenic ranges, except in the areas bounding the Serbo-Macedonian massif (Vardar zone) (Fig. 4).

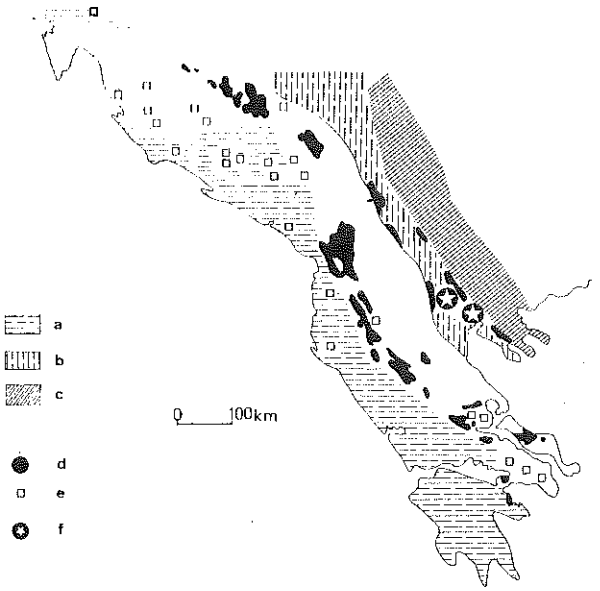

Fig. 4 *Distribution of the ophiolites and the pre and contemporaneous ophiolite volcano-sedimentary formations in the Dinaro-Hellenic area.*

Legend

a- External zones

b- Vardar zone

c- Serbo-Macedonian and Rhodope massives

d- ophiolites

e- pre-ophiolite volcano-sedimentary sequence

f- post-ophiolite volcano-sedimentary sequence

In Yugoslavia, the volcanism associated with radiolarian cherts, pelites and sandstones belongs to the *Porphyrite-Radiolarite* formation (Ciric 1954, 1960, 1964) of Aniso-Ladinian age. This formation, 50 to 300 m thick, lies on Anisian neritic limestones and dolomitic strata (Fig. 5a). It is overlain by either neritic or pelagic sediments. This volcanism is related to block faulting of the dinaro-Hellenic carbonate plate leading to the formation of ridges and troughs.

The volcanic flows consist of basalts, andesites, dolerite-diabases, acid lavas and their associated tuffs (Pietre Verde). Low-Ti basalts and andesites present a typical calc-alkalic trend (Bébien et al. 1978). Yugoslavian authors (Knezevic 1975, Djordjevic and Knezevic 1969 1972) have emphasized the frequent and important alterations of these volcanics (albitization and potassification). Pamic (1977a) related some intrusions of albite granites and syenites to the Aniso-Ladinian volcanism.

The Middle Triassic volcanism of the Dolomites is very similar to the *Porphyrite-Radiolarite*. Acid tuffs (Pietre Verde) are overlain by pillow-lava flows, often brecciated, of latite-basalt composition (Rossi et al. 1974-1976, Castellarin et al. 1977, Carraro and Fiora 1974). The chemical features of the basic lavas show calc-alkalic and/or shoshonitic affinities (Calanchi et al. 1978).

In Greece, this Middle Triassic volcanism is more scattered and diversified. The flows, mainly of basic composition, pillowed or unpillowed, associated with pyroclastic breccias and hyaloclastites are interbedded in radiolarian cherts, siliceous pelites and cherty limestones (Fig. 5b). They outcrop in Othris (Ferrières 1972, 1974; Celet et al. 1976), in the northern Pindos (Terry 1971) and in Akarnania (Fleury 1976). The uppermost Carnian-Norian levels are dated by *Halobia* limestones. Some of the Othris basalts exhibit alkalic affinities (Hynes 1974, Menzies 1976). These basaltic flows are related to the formation of an important trough – the Maliac trough – (Ferrieres 1974). The Vardoussia Aniso-Ladinian volcanism, 50 to 100m thick, consists of meta-basalts, meta-andesites and trachytes (Ardaens 1978). Basaltic flows overlain by Upper Triassic sediments have been described by Verriez (1976).

In Attica, Parnis and Central Eubea the Middle Triassic volcanism is intermediate/acid: meta-rhyolites, meta-dacite flows and tuffs interbedded either in sandstones or in cherty limestones (Clément 1968, Guernet 1971, Katsikatsos 1971). The Anisian volcanism in Albany occurs widely (Ndojaj 1978): basaltic, andesitic and acid flows show alkalic, calc-alklalic and tholeiitic trends.

The ophiolites and their contemporaneous volcanism

The Jurassic age of the Dinaric and Hellenic ophiolites is now generally accepted. According to structural, petrological and geochemical studies, these ophiolites are very diverse and do not belong to a single type (Dimitrijevic and Dimitrijevic 1973, 1976; Rocci et al. 1975).

Two ophiolitic belts can be distinguished (Maksimovic 1975; Pamic 1977b; Vergely 1977; Beccaluva et al. 1979). In the Western one (Rampnoux 1970, 1974; Cadet 1976; Charvet 1978), the lower units are formed of pillow-lavas and diabase associated with a mélange-type formation: the *Diabase-Radiolarite* formation (Diabase Horn stein-Ciric 1954; Celet et al. 1977). Slices of gabbro cumulates with ultrabasic levels and sheeted complex occur (Terry 1974). The uppermost tectonic unit is a sheet of peridotite, mainly lherzolites. Nevertheless, harzburgites may occur in some massifs. Chromite deposits are rare. Metamorphic slices may be found between the ultramafic sheet and the *Diabase-Radiolarite* formation (Karamata 1968; Pamic 1971; Pamic et al. 1973).

A tectonic phase with transcurrent movements has intensely affected the ophiolites of the Eastern belt. The ultramafic rocks are mainly harzburgites with abundant chromite deposits. Some of the ophiolites occurring along the edge of the Serbo-Macedonian massif are associated with granites (Bébien 1977), (Fig. 6c).

The exact nature of the differences between the Dinaro-Hellenic ophiolites of the Western and Eastern belts still requires further characterization. Nevertheless, taking into account present data, these ophiolites must have different origins.

The volcanism of the Vardar zone, contemporaneous with some ophiolites of the Eastern belt may be divided into two main formations (Mercier 1966; Mercier et al. 1975). In the Païkon (Fig. 6) acid (rhyolites - dacites) and basic lavas (meta-basalts) are associated with neritic limestones. The uppermost levels are Upper Jurassic (pre-Tithonic). The meta-basalts are low-K tholeiites, sim-

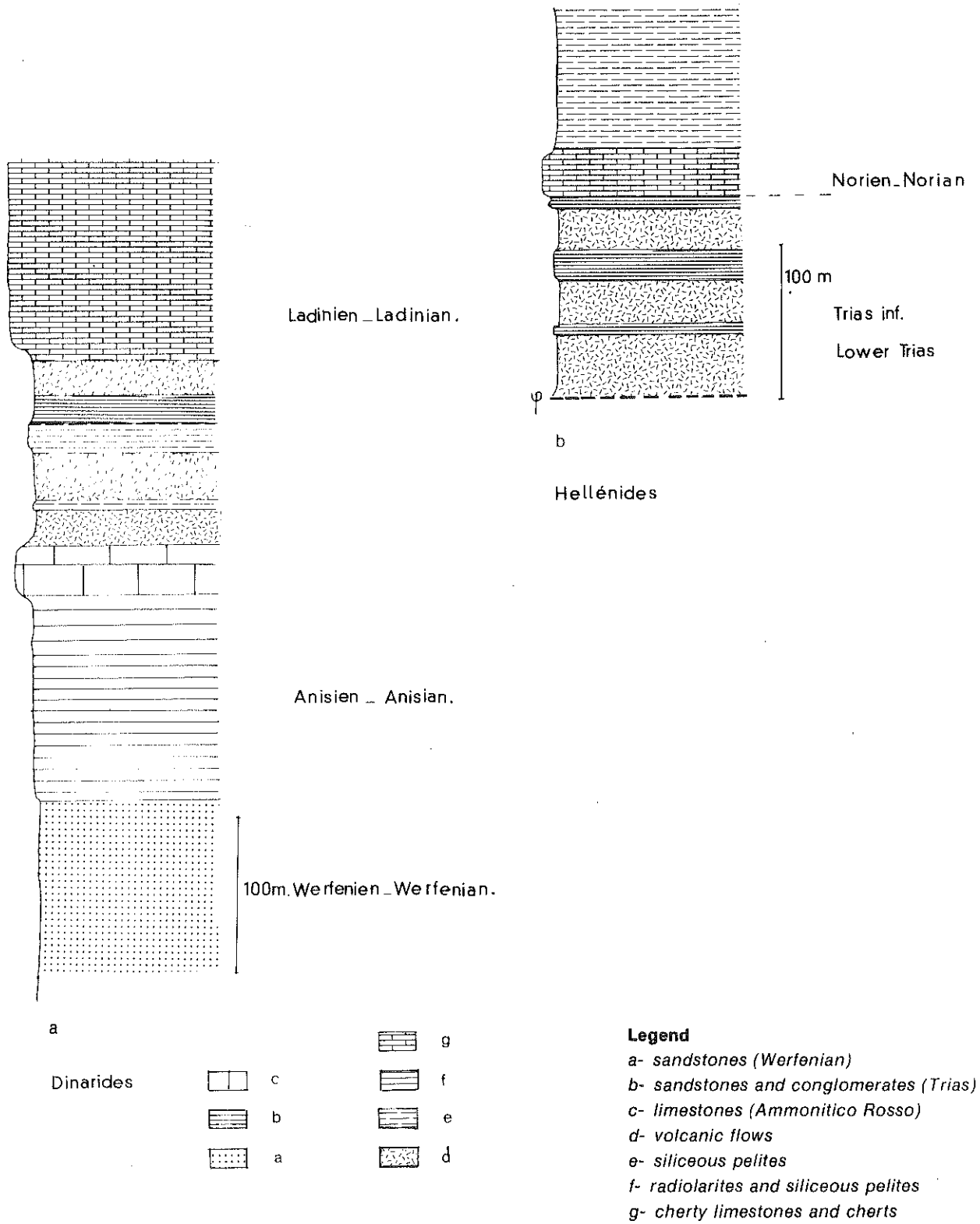

Fig. 5a *Diagrammatic section of the Aniso-Ladinian pre-ophiolite volcano-sedimentary formation in the Dinarides modified after Celet et al. 1977.*

Fig. 5b *Diagrammatic section of the Aniso-Ladinian pre-ophiolite volcano-sedimentary formation in the Hellenides from Celet et al. 1977.*

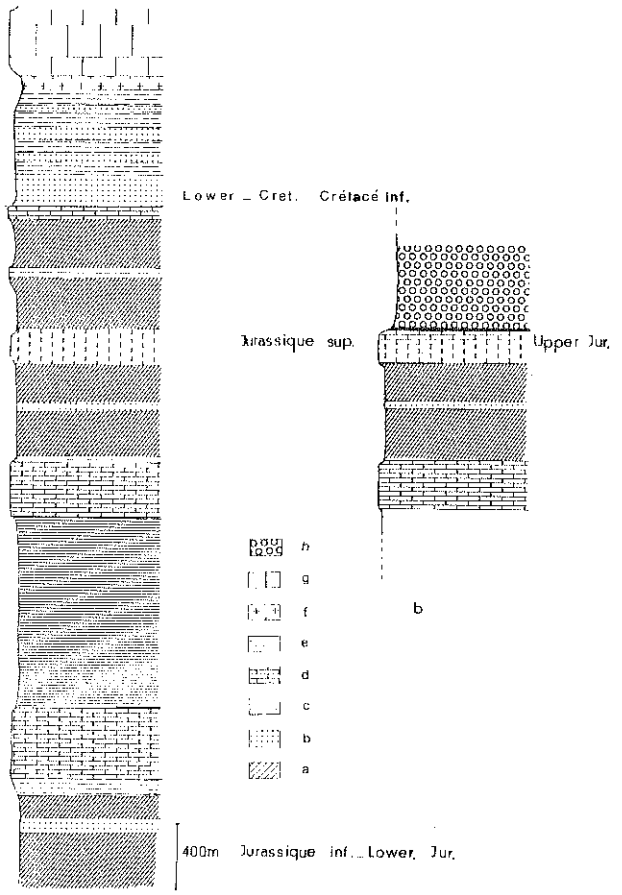

Fig. 6a-b *Diagrammatic sections of volcano-sedimentary sequences, contemporaneous to the ophiolites in the Vardar zone modified after J. Mercier (1966).*

Legend

a- basic and acid volcanic flows
b- sandstones
c- pelitic shales
d- limestones
e- green-schists
f- basic volcanic post-ophiolite flows
g- Middle Cretaceous limestones
h- ophiolites

ilar to those present in island arcs. Between the Païkon and the Serbo-Macedonian massif, basaltic flows and sills are interbedded in a Liasic pelagic sequence. The basalts are chemically similar to oceanic tholeiites.

Jurassic flows occur also in the Pelagonian massif and in the Almopias zone, between the Pelagonian and Païkon massifs (Mercier 1966).

The post ophiolite volcanism

A post-ophiolite volcanism of Cretaceous age occurs in Yugoslavia and Greece. In the Vardar zone and the Serbo-Macedonian massif, volcanics associated with flysch-type sediments were described by Ciric (1966), Mercier (1966) and Charvet (1978). In the central Peloponese, Upper Jurassic – Lower Cretaceous lavas are interbedded with limestones and radiolarites (Richter and Lensch 1977). This andesitic calc-alkalic volcanism is to be related to the tectonic emplacement of the ophiolites which took place at the end of the Jurassic in the whole Dinaro-Hellenic domain.

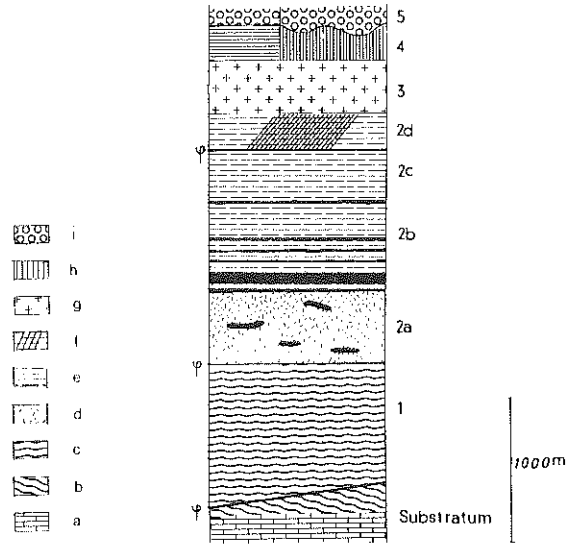

Fig. 6c *Diagrammatic section of Vourinos ophiolites (Greece) from A. Paupy (1976).*

Legend

a- substratum (Limestones)
b- metamorphic rocks
c- metamorphic ultrabasic rocks (1)
d- dunites with chromite deposits (2a)
e- layered plutonic complex (2a-b-c)
f- transition zone (2d)
g- plagiogranites (3)
h- sheeted complex (dykes and sills) (4)
i- volcanic pile (5)

Conclusions

The most important feature in the Dinaro-Hellenic ranges is the occurrence of three periods of volcanism in the Middle Trias, in the Middle and Upper Jurassic and at the end of the Lower Cretaceous. That is, before the formation of the ophiolites, during their crystallization and during their tectonic emplacement.

Eastern Mediterranean area

South Turkey, NW Syria and Cyprus (Fig. 7). This area is restricted to the southern Turkish ophiolite belt, because the Turkish ophiolites and their related volcanism are still incompletely studied; their geological, petrological and geochemical features are too fragmentary to draw any serious paleogeographical conclusions at present. Turkish ophiolites form three belts; from N to S: Pontides, Central and Southern Taurides (Lycian and Antalya nappes, Pozanti-Karsanti massif). The most southern ophiolite belt, bounding the Africo-Arabian platform is formed by the Troodos and Hatay ophiolites.

These two most southerly ophiolite belts are characterized by an important pre-ophiolite volcanism of Upper Triassic age.

The ophiolites of the Eastern Mediterranean area are younger than the western Alpine and Dinaro-Hellenic ophiolites. In contrast to the previously cited orogenic domains, there is an important break during the whole Jurassic, where no evidence of magmatism is found.

The pre-ophiolite volcanism

The volcano-sedimentary series linked to the ophiolites,

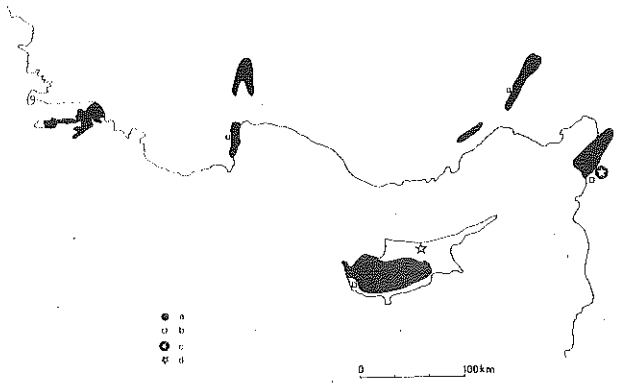

Fig. 7 *Distribution of the ophiolites and the pre, contemporaneous and post ophiolite volcano-sedimentary formations in Eastern Mediterranean area.*

Legend

a- ophiolites

b- pre-ophiolite volcano-sedimentary formations

c- volcano-sedimentary formation, almost contemporaneous to some ophiolites

d- post-ophiolite volcano-sedimentary sequence

are essentially composed of sandstones, *Halobia* limestones, calcareous silts, radiolarian cherts with different interbedded volcanic levels. Lapierre and Parrot (1972) have shown the similarity existing between these series both in NW Syria and Cyprus and more recently Delaune-Mayere et al. (1977) have extended this to Antalya and Lycian Taurus nappes. These series have also been found in the Pozanti – Karsanti massif (Çatakli et al. 1977; Çakir 1978).

This volcanism, closely related to *Halobia* and ammonite-bearing limestones of Norian – Carnian age, consists of pillowed and/or unpillowed flows (Fig. 8). In the Antalya nappes (Juteau 1974), thick alkalic basaltic strato-volcanoes existed with thin pelagic limestones. In SW Cyprus (Mamonia nappes – Lapierre 1972; 1975; Desmet and Lapierre 1974; Lapierre and Rocci 1976), this alkalic series is highly differentiated from alkalic basalts – mugearites to alkalic trachytes. In Syria (Parrot 1977), this volcanism presents tholeiitic to alkalic affinities.

In SW Cyprus the Jurassic shale and chert sedimentary sequence overlying this late Triassic volcano-sedimentary formation does not reach the Middle Cretaceous (Ealey and Knox, 1975; Robertson and Woodcock, 1979); in NW Syria on the contrary the same unit is continuous from Upper Triassic to Turonian (Delaune-Mayere and Parrot 1976). In Antalya, no Jurassic sediments are known (Marcoux 1978). The lack of Lower Cretaceous in Cyprus and Antalya would partly explain the absence of a Neocomian undersaturated and saturated alkalic volcanism (basanites, tephrites, phonolites and trachytes) observed in NW Syria (Baër-Bassit; Parrot 1977, Fig. 8).

The ophiolites

The southern Turkish ophiolites present both a different magmatic sequence and structural disposition compared to Troodos-Hatay.

The Antalya ophiolite sequence is not complete (Juteau 1974). The sheeted complex is reduced. No volcanic sequence is known, related to the cumulate sequence. The rocks are intensely sheared in a succession of more or less thick slices. The dating by K/Ar of the different parts of the ophiolite gives a Middle Cretaceous age (Thuizat et al. 1978, 1979). The Pozanti – Karsanti ophiolites seem to be more complete (Çatakli et al. 1977; Çatakli 1978; Bingöl 1978; Çakir 1978). No real sheeted complex occurs. The pillow lava sequence associated with radiolarian cherts and limestones, of tholeiitic trend, could represent the volcanic term of the ophiolite succession. The structural disposition is more or less similar to the Greek and Yugoslavian ophiolites. The uppermost tectonic level consists of an important and thick harzburgitic sheet on which rest slices of gabbro cumulates and peridotites. The lowermost levels are formed by pillowed basalts and their associated sediments. Between the volcanic sequence and the harzburgite sheet outcrop metamorphic slices (Whitechurch 1977; Whitechurch and Parrot 1974; Parrot and Whitechurch 1978). The Troodos - - Hatay ophiolites (Juteau et al. 1973; Parrot 1973) are somewhat different. The main characteristics of these two ophiolites is the occurrence of an important sheeted complex (Rocci 1973, Kidd and Cann 1974; Kidd 1977; Desmet et al. 1977 a-b, 1978 a-b) and its related volcanic sequence (Smewing et al. 1975; Gass and Smewing 1973; Desmet 1977). The Troodos ophiolite suite is quite well known

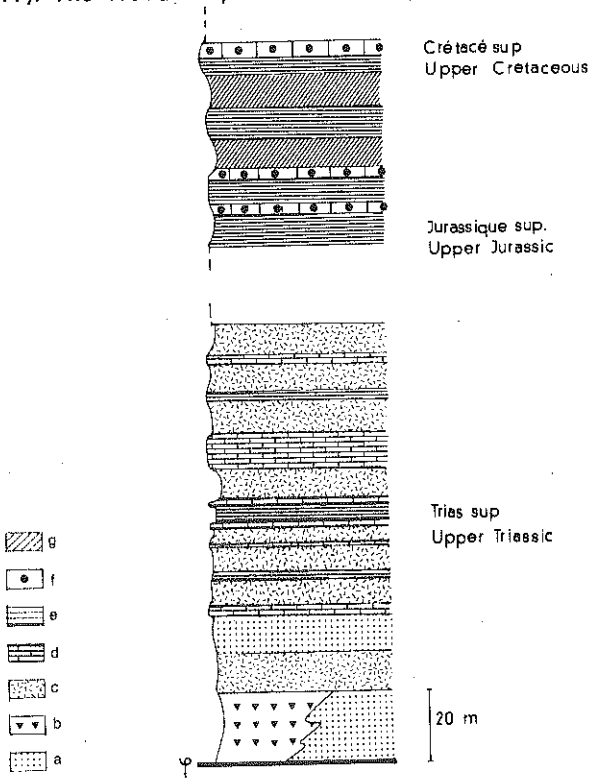

Fig. 8 *Diagrammatic section of the Upper-Triassic and Neocomian volcano-sedimentary formation of Eastern-Mediterranean areas.*

Legend

a- sandstones

b- volcanic breccia

c- basaltic flows

d- pelagic limestones (± Halobia)

e- shales and cherts

f- calcarenites

g- Neocomian under-saturated volcanic flows

(Wilson and Morel 1959; Bear 1958; Gass 1968; Gass and Masson-Smith 1963; Moores and Vine 1970; George 1978) (Fig. 9). The Troodos and Hatay rocks have been dated by K/Ar; they are of Upper Cretaceous age (Delaloye et al. 1977; Vine et al. 1973; Desmet et al. 1978b; Delaloye and Desmet 1979). From a structural point of view, the Troodos plutonic core is formed of a succession of slices overthrust by an important harzburgite sheet (Turner 1973; Bortolotti et al. 1976; Biju-Duval et al. 1976) (Fig. 9).

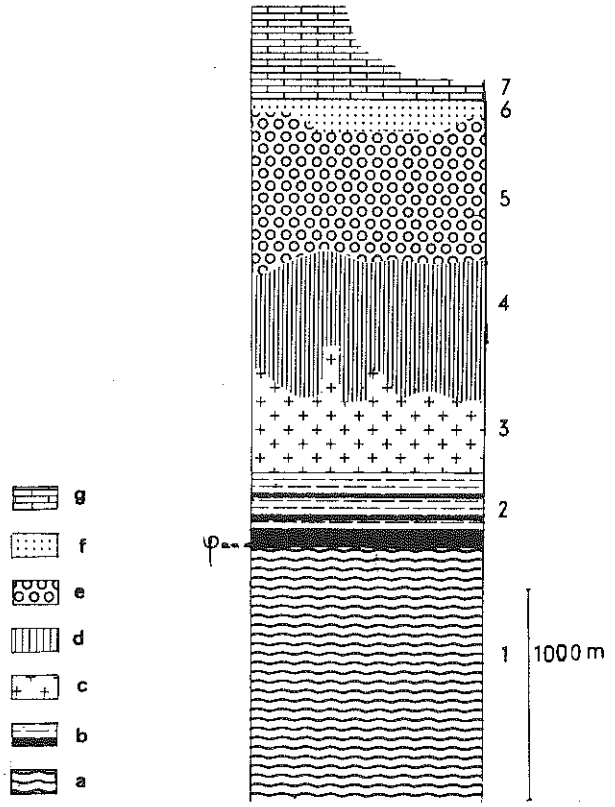

Fig. 9 *Diagrammatic section of the Troodos ophiolite (Cyprus)*

Legend

a- metamorphic harzburgites

b- cumulate layered plutonic complex

c- doleritic gabbros and plagiogranites

d- sheeted complex

e- volcanic pile

f- Upper Cretaceous umbers and volcanic sandstones (Kannaviou formation)

g- pelagic limestones (Lefkara formation)

The post-ophiolite volcanism

The Eastern Mediterranean post-ophiolite volcanism is, at present, only known in the Pentadaktylos range, in Northern Cyprus (Bear 1959; Baroz et al. 1976; Baroz 1976).

This volcanism belongs to the pelagic limestone sedimentation of the Lapithos group (Fig. 10), which rests with an angular unconformity on the platform carbonate deposits of the Trypa group. Two main volcanic levels occur. The first one, of Maastrichtian age, consists of a basalt – dacite – rhyolite series with no andesite. The geochemical features indicate a calc-alkaline trend. The second level is Paleocene. Pillowed basaltic flows and trachytes show a potassic alkalic differentiation trend, comparable to shoshonites.

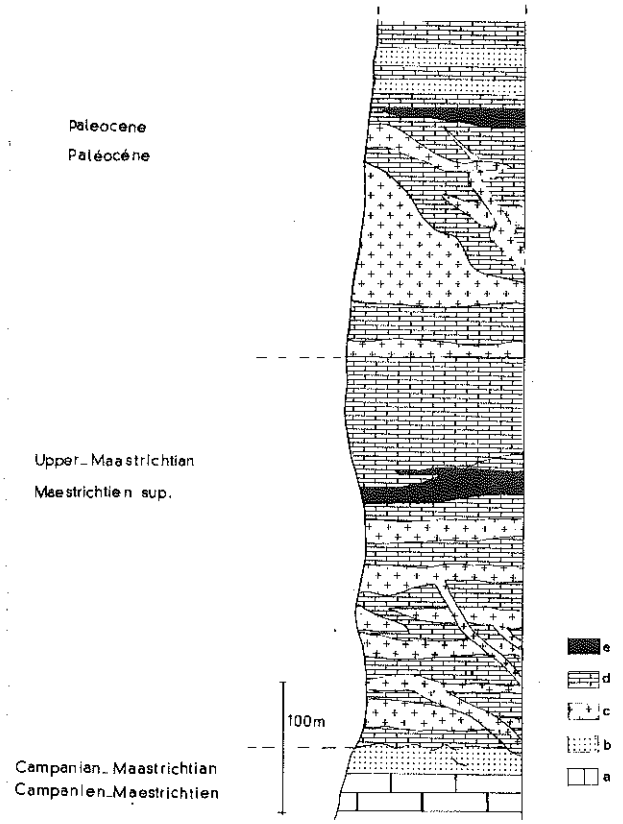

Fig. 10 *Diagrammatic section of the post ophiolite volcano-sedimentary formation of Northern Cyprus.*

Legend

a- platform limestones of Jurassic and Middle Cretaceous age

b- sandstones

c- basic volcanic flows

d- pelagic limestones

e- Maestrichtian rhyolite – Palaeocene trachyte

Interpretation

In the three areas distinguished here, the volcano-sedimentary formations show three different types of evolution of the Tethyan domain during the Mesozoic.

As emphasized by Smith (1971), the evolution of the Mediterranean oceanic domain is controlled by the plate motions of Eurasia versus Africa with at least, three phases of relative movement: during Lower Jurassic, Upper Cretaceous – Late Eocene and Eocene to Recent times. In this paper, only the first two opening phases are discussed, the kinematics of which were described by Dewey et al. (1973) and Biju-Duval et al. (1977).

1. In the Western Alps, Apennines, Corsica and Calabria, the magmatism is presumed to be related to continuous opening, first in a continental domain, evolving into an oceanic one, during Permian to Jurassic times. Taking into account the length of evolution, the supposed ocean should not be very wide (a few hundred kilometres).

The Upper Cretaceous magmatism of the Valaisan domain is more difficult to explain. Although this volcanism is contemporaneous with a phase of high pressure-low temperature metamorphism of the ophiolites, it could be due to a restricted crustal extension, perhaps related to strike-slip movements (Marignac 1977).

The lack of calc-alkalic island-arc volcanism in the post-ophiolite formations of the Western Alps comparable

to that related to active continental margins, shows presumably the absence of subduction of the oceanic lithosphere produced during the Jurassic.

2. The highly differentiated mainly basic Upper Triassic volcanism of the Eastern Mediterranean (Southern Turkey, Cyprus and Syria), is related to block faulting in the Arabian – African plate, the classic example of alkalic volcanism in a continental rift system. Such movements reoccur, after the last eruptions of under-saturated alkalic lavas in Syria during the Neocomian. The correlation between the Eastern Mediterranean and the Dinaro-Hellenic volcanism is still difficult. Nevertheless, the Upper Triassic volcano-sedimentary formations of Turkey, Cyprus and Syria could be correlated with the volcanism of Crete and Pindos of about of the same age (Pindos facies of Aubouin 1959). The origin of the Arabian – African plate rifting is still unknown.

The development of the calc-alkalic volcanism of Maastrichtian and Paleocene age in the Pentadaktylos range, northern Cyprus, is related to a subduction and strike slip movement along the Kythrea fault.

3. The non-ophiolite volcano-sedimentary sequences present in the Dinaro-Hellenic ranges show another type of geotectonic evolution. These lavas are, in fact, similar to the lavas erupted along active continental margins and present-day island arcs. They could show, therefore, the existence of subduction and lead us to admit the presence of an oceanic domain in this area during the Trias. The age of the volcanic rocks allow us to distinguish two areas. The Western one, includes the main part of the Hellenides and Dinarides, the Eastern one, corresponds roughly to the Vardar zone (or only its Eastern part, Charvet 1978).

In the Western area, the volcanism is mainly Triassic. It presents typical calc-alkalic features in Yugoslavia and Dolomites. On the contrary, it includes tholeiitic or alkalic lavas in Albania and Greece, and therefore, seems to be different from the volcanism of present areas of converging plates. This Triassic magmatism, whatever its chemical characteristics may be, is related to the individualisation of troughs which might develop after into oceanic basins with the formation of ophiolites (Maliac zone and perhaps Serbian and Bosniac zones).

The Eastern area is characterized by an important Jurassic volcanism occurring mainly in Greece. The basic lavas are mainly low-K-tholeiites, similar to those found in present-day island arcs. This volcanism is contemporaneous with the ophiolite and granite formation. It could represent an extension towards the West of the Jurassic volcanism known in the Pontides (Fourquin 1975, Bergougnan 1976) and in the Caucasus (Khain 1975).

Subduction zones towards the West during the Trias and later towards the N-East during the Jurassic, bounding an ocean (Tethys) which divides the Apulian plate from Europe, can explain the distribution of the magmatism in space and time (Mercier et al. 1975, Cadet 1976, Blanchet 1977, Charvet 1978, Bébien et al. 1978, Le Pichon et al. 1978). The origin of the ophiolites tectonically emplaced during the Upper Jurassic must be looked for in the basins opened at the end of the Trias and in the Jurassic to the East as well as to the West (Maliac, Serbian, Bosniac zones) or in the Tethyan ocean itself (Vergely 1976, 1977). Their diversified petrological characteristics are probably due to their varied origins.

The volcanism continued in the Hellenides and Dinarides, after the Upper Jurassic – Eocretaceous tectonic phase by the eruption of andesites in the Cretaceous.

It these different volcanisms are located on a schematic paleogeographic map, inferred from those published by Biju-Duval et al. (1977), their relations with the Mesozoic evolution of the Mediterranean domain is obvious. During Middle and Upper Trias (Fig. 11), the volcanic eruptions almost completely surround the Apulian plate. To the East, the Hellenic and Dinaric volcanism must be at least partly related to the subduction of the Tethyan ocean towards the West. To the West and to the South, the alkalic basic volcanism of the Western Alps, Apennines, Calabria, south Turkey, Cyprus and Syria, is related to the opening and the formation of an oceanic area between the Apulian plate and its extension towards the East (Anatolian plate) on one hand and the European – Iberic and African plates on the other hand. The eruptions are perhaps located preferentially at triple points between Europe – Iberia – Apulia, Iberia – Africa – Apulia and Apulia – Anatolia – Africa. The problem of a possible relation between the volcanism of the Eastern and Southern boundaries still claims our attention.

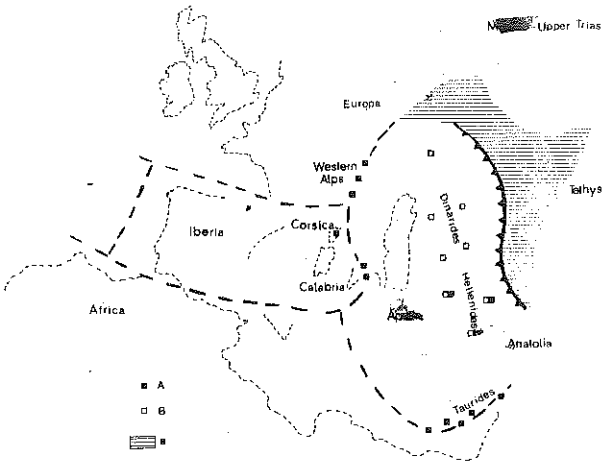

Fig. 11-A *Palaeogeographic map for the Middle and Upper Triassic.*

Legend

A- alkalic volcanism

B- calc-alkalic volcanism

a- tethyian oceanic domain

subduction zone

plate boundaries

The age of the Eastern Mediterranean ophiolites being Middle to Upper Cretaceous (Delaloye et al. 1977; Thuizat et al. 1978, 1979; Delaloye and Desmet 1979), no data allow us to assume that an ocean existed during the Jurassic between the Apulian and African plates. We have therefore supposed that Africa moved along with the Apulian plate towards the East. The Ligurian-Piedmont ocean, which began to open, seems to be an extension towards the North of the Atlantic ocean.

The cessation at the end of the Trias of the calc-alkalic volcanism, east of the Apulian plate corresponds probably to the end of the westerly dipping subduction zone of the Tethyan ocean. The troughs individualized at the same period expand to become oceanic basins. The important volcanism during the Lias in the Vardar zone and the Pontides shows the existence of a new subduction zone now dipping north.

At the end of Jurassic (Figs. 12-13), the Apulian plate impinged against Europe. The shock was preceded by

dextral shear movements due to an anticlockwise rotation of the Apulian plate. It is in such a structural setting that inner-arc basins opened in the Vardar zone (Bébien and Gagny 1978). These shear movements were quickly followed by crustal shortening in the Hellenides and Dinarides with the tectonic emplacement of the ophiolites (Dercourt et al. 1977).

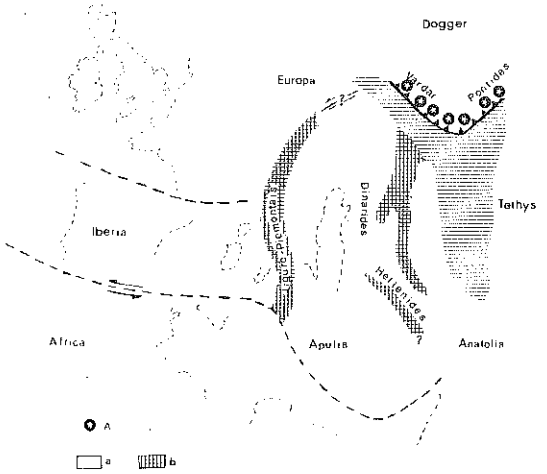

Fig. 12-A *Palaeogeographic map for the Dogger.*

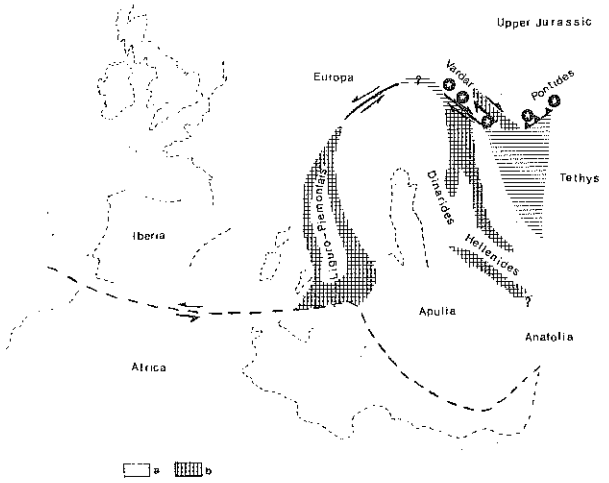

Fig. 13-A *Palaeogeographic map for the late Jurassic. same symbols as fig. 11 and 12*

Legend

A- contemporaneous ophiolite volcanism

B- an opening ocean basin

b- Mesogean oceanic domain

same symbols as fig. 11

The Apulian plate was only separated from Africa during the Cretaceous (Fig. 14) perhaps after the eruption of the Syrian Neocomian under-saturated lavas, provoking the opening of the Mesogea and therefore the formation of the ophiolites of Southern Turkey, Cyprus and Syria. At the same time, the Tethyan ocean was probably completely subducted, since an important andesitic volcanism characterizes the area bounding the Apulian and European continental blocks. To the West, the Ligurian-

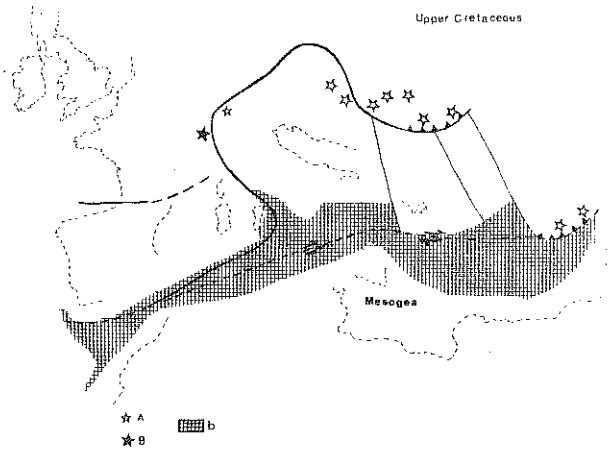

Fig. 14-A *Palaeogeographic map for the late Cretaceous.*

Legend

A- calc-alkalic post-ophiolitic volcanism

B- tholeiitic post-ophiolitic volcanism

same symbols as fig. 11 and 12

Piedmont basin has also completely disappeared at the same time. Nevertheless, local extension movements related to a tholeiitic volcanism occur in the Valais domain. The late-Cretaceous (Campanian – Maastrichtian) of the North Atlantic, is related to the reversal of the strike-slip movement between Europe and Africa which changed sinistral to dextral. This movement has provoked the bringing together of the Arabo-African and Anatolian plates (the Arabian plate moved quickly from South to North). This movement led to the main tectonic phase, along the northern boundary of the Arabian plate, provoking the thrusting onto this plate of all the ophiolitic suites (Southern Turkey – Troodos – Hatay) and their related Bassit slices). Perhaps the same compressive movement radiolarian sheets (Antalya and Mamonia nappes – Baër- with related thrustings led to local subduction zones, marked by calc-alkalic volcanism as in the Pentadaktylos range in Northern Cyprus.

This evolution of the peri-Mediterranean area, defined more accurately from a study of the volcanism, easily accounts for the variety of the peri-Mediterranean ophiolites.

Before the separation of North America and Europe, the Ligurian-Piedmont basin constituted during the Jurassic the extension of the Atlantic towards the north. The similarities noted between the Western Alpine, Corsican and Calabrian ophiolites and rocks sampled on the ocean floor are not astonishing.

Compared to the opening of the Ligurian-Piedmont basin, the opening of Mesogea is long and hesitant. Already announced during the Trias by an important alkalic volcanism, this opening took place only during the Late Cretaceous after a break during the Jurassic. The lack of an important mantle activity immediately before the formation of the Eastern Mediterranean ophiolites could perhaps explain their peculiar petrological characteristics. They could be formed from a mantle, depleted by the alkalic eruptions. It is also possible that this mantle depletion is related to the subduction (Sun and Nesbitt 1978) of the Tethyan ocean under the Apulian block.

The Dinaric and Hellenic ophiolites were formed in troughs individualized along active continental margins. The fact that the ophiolites belonging to the Apulian margin were formed after the volcanism related to the subduction zone and that the ophiolites of the Vardar zone were formed during this same volcanism, could explain their petrological differences. In another way, it is not impossible, although it has not been proved, that some ophiolites came from the main Tethyan ocean.

Summary and conclusions

The ophiolites have for a long time been considered to be remnants of an oceanic crust. It appears, however, that the geotectonic significance of the ophiolites must be reconsidered. We propose a schematic evolution of the Mediterranean areas from the Trias until the end of the Cretaceous.

1. In the Middle and Upper Trias, mainly alkalic volcanism related to rifting appears in the Western Mediterranean area and the South-East Mediterranean. On the contrary, calc-alkalic, tholeiitic and alkalic basalts related to subduction zones occur in the Dinarides and Hellenides. It is not impossible that this event partly provoked the destruction of the Permian Tethys and the modification of the upper mantle in Eastern Mediterranean.

2. During the Jurassic, a seaway, bounded by two passive continental margins, occurred in the Alps. The ophiolites are similar to present-day oceanic crust. In the Central zone of the Dinarides, Pindos and Othris, the ophiolites also show some affinities with present-day oceanic crust. On the other hand, the Vardar zone and Vourinos were formed in back-arc basins related to subduction zones.

3. During the Cretaceous in Western Mediterranean the main magmatic event stopped and compression associated with shearing began. On the contrary in the Eastern Mediterranean the opening of basins leading to the formations of Southern Taurides, Cyprus and Syrian ophiolites occurred.

The main features inferred from the study of this paleogeographic evolution of the Mediterranean areas during the Mesozoic are the following:

1. The diversity of the Mediterranean ophiolites which can be classified in two main groups.

   The scarcity of true oceanic ophiolites.

   The nature of the oceanic crust which shows no variation with the width of the ocean.
2. Rifting is followed immediately in the Western Mediterranean by the opening and the formation of an oceanic basin. On the contrary, in the Eastern Mediterranean there is a long break between the eruptions related to the rifting and the individualization of the oceanic basins.
3. Passive continental margins occur in the Western Mediterranean and active continental margins in the Central Mediterranean.
4. Subduction may probably be absent in the Western Mediterranean and present in the Dinarides and Hellenides. Thus, a new mechanism for the formation of blue schists must be proposed.
5. The evolution of the Mediterranean areas is mainly controlled by triple junctions (at least three) and by transcurrent faults between continental blocks.

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DIVERSITY OF THE GREEK OPHIOLITES : BIRTH OF OCEANIC BASINS IN TRANSCURRENT SYSTEMS

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Dedicated to J. H. BRUNN

ABSTRACT

A review of most of the Greek ophiolites has been compiled. Sedimentary and tectonic data, petrographic and geochemical features are used to define their precise geological setting. It appears that Greek ophiolites have resulted from the opening of several small basins, at least three in a continental environment. Transcurrent movements may play an important role in the shearing and the thinning of the continental crust and in the emplacement of the subsequent oceanic crust and upper mantle. This mechanism is in good agreement with the structural evolution of the Hellenides.

INTRODUCTION

Since the beginning of the 19th century, Greece has been used as a geological reference as for example in the development of theories on geosynclines (Aubouin, 1965) and in the description of ophiolitic massifs (Brunn, 1956 ; 1961) made up of rocks similar to oceanic samples (Brunn, 1959). The role of ophiolites as remnants of oceanic crust in plate tectonic models, has been emphasized (Dercourt, 1970 ; 1972 ; Smith, 1971 ; Dewey et al., 1973). In this context, the occurrence of two main ophiolitic belts (Kober, 1952, in Brunn et al., 1977) must put some constraints on geological reconstructions : the ERO zone, the most westerly ophiolites (sub-Pelagonian) and the IRO zone ("Vardar"). Either one or two oceanic basins have been proposed for the origin of these ophiolites. In the first case, the ophiolites would have an ultra-Pelagonian origin (Kober, 1952 ; Bernouilli and Laubscher, 1972 ; Dercourt, 1972), whereas in the second case the ophiolites would come from the sub-Pelagonian and ultra-Pelagonian zones (Dercourt, 1970 ; Aubouin, 1976). The occurrence of relatively closed marginal basins is more easily compatible with an active ocean margin as in the Southwest Pacific (Mercier et al., 1975 ; Aubouin et al.,

1977 a and b ; Blanchet, 1977). This type of geotectonic setting with adjacent ocean, arc and marginal basins may best explain the petrographical and geochemical diversity of Greek ophiolites (Rocci et al., 1975 ; Beccaluva et al., 1979). Thus, the evolution of the different ophiolitic basins must be recorded together with the geological history of the associated formations : the pre and contemporaneous volcano-sedimentary series (Celet et al., 1977 ; Rocci et al., 1980), and the nature of the post-ophiolitic sediments and the tectonometamorphic events (Vergely, 1977) in order to define the geotectonic situation of the Greek ophiolites.

SHORT SURVEY OF THE GEOLOGY OF GREECE

In Greece, the Mesozoic and Cenozoic geological formations, including the ophiolitic complexes, occur in a remarkable system of sheets, thrust during the Tertiary toward the West and the South-West. In this system, the structural, metamorphic and magmatic evolution allows one to contrast Eastern and Western areas (fig. 1).

1. Structural evolution. A distinction is always made between the external Hellenides, to the West, in which no folds occur before the late Eocene, and the more easterly internal Hellenides, affected by early tectonic movements during the late Jurassic and the Eocretaceous (Brunn, 1956).

Owing to more accurate studies, two early tectonic phases are now recognized (Mercier, 1966 ; Vergely, 1977).

- The first one (JE_1), occurring from the border of the Serbo-Macedonian massif to the Pelagonian massif characterizes an internal area sensu stricto. Its related structures and microstructures result from very important tangential tectonics with a West or South-West vergence.

- The second phase (JE_2) is early Cretaceous, and affects only Westerly units qualified as "intermediate" (Vergely, 1977). The vergence of the resulting structures is towards the East or the North-East.

These JE_1 and JE_2 phases both produced the tectonic setting of the ophiolites (Vergely, 1976 ; Ferrière and Vergely, 1976).

2. Metamorphism in the greenschist or in the glaucophane blueschist facies affect, the internal areas up to the Pelagonian massif in Macedonia. Farther to the South, they affect external areas and reach the South of Peloponnesos. From field arguments and geochronological data, one can attribute early Cretaceous, Eocene and Oligocene ages to these metamorphisms : they are closely linked to the various phases of deformation (Yarwood and Dixon, 1977 ;

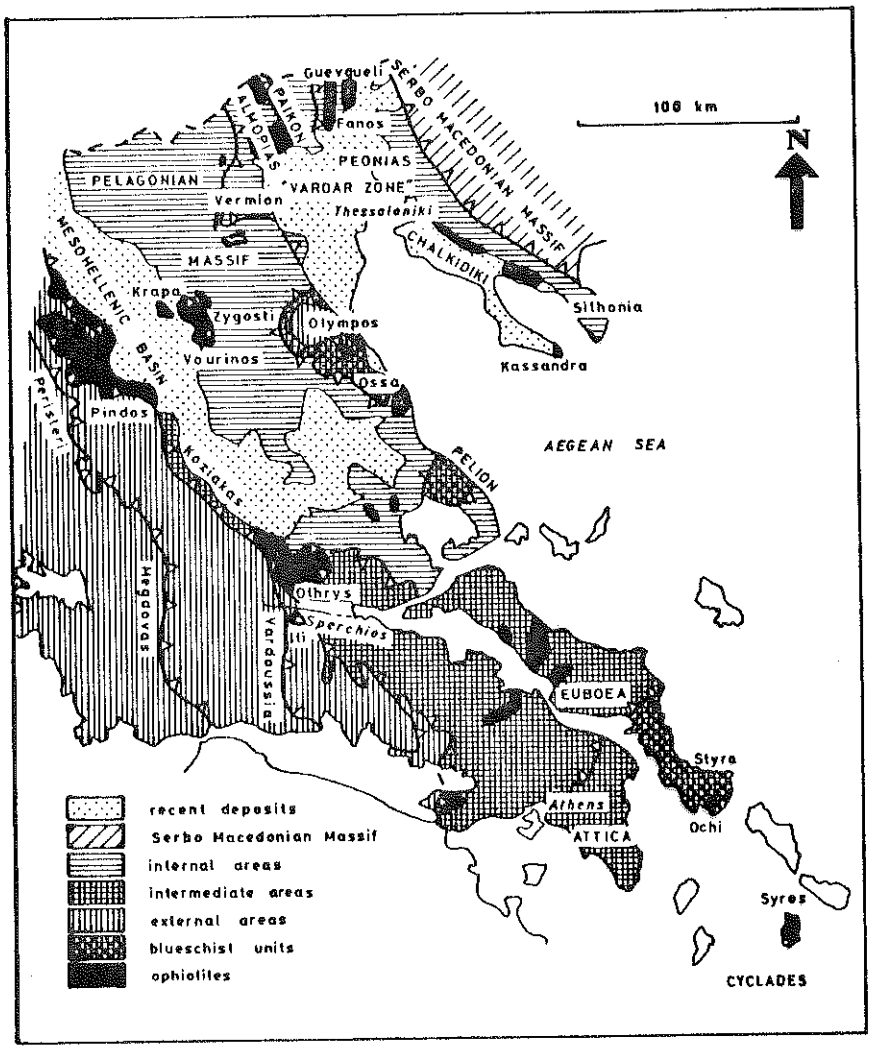

Fig. 1 - Distribution of the ophiolitic complexes in Greece.
The main tectonic and metamorphic zones are represented (see text).

Bavay et al., 1980 ; Bonneau et al., 1980 a and b).

3. The Mesozoic magmatism is spectacularly represented by ophiolitic complexes in the intermediate and internal areas of Greece. The other effusive and plutonic formations, disrupted by tectonic phenomena and observed by metamorphism, are often poorly known (Celet et al., 1977 ; Rocci et al., 1980). Thus, the following list will be incomplete.

Some Triassic volcano-sedimentary outcrops, including mainly pillow lavas, tuffs and breccias of basic composition, are found in external areas (Peristeri, Megdovas, Vardoussia). More Easterly, effusive series, also of Triassic age, and made up of basalts and spilites with some keratophyres and dacites, are spatially linked to the ophiolites of Northern Pindos and of Othrys, from which they are separated by faults. In Attica and central Euboea, metabasalts, metarhyolites and metadacites give evidence of the existence of basic and acid volcanism in the middle Trias.

In the internal areas, between the Pelagonian and the Serbo-Macedonian massifs, many pre-Cretaceous volcanic and volcano-sedimentary formations may be found ; unfortunately, their ages are precisely known only in a few cases. In the Paikon massif, thick acid and basic formations (metabasalts, metarhyolites, keratophyres) are tightly intermingled with Upper Jurassic neritic limestones. More to the East near the Serbo-Macedonian massif, basaltic flows and sills are interbedded with sediments of probable Liasic age.

The Fanos granite massif with a K/Ar age of 150 m.y. (Upper Jurassic) outcrops among the ophiolitic rocks on the Eastern side of the Paikon massif.

Thus, an important and well diversified non ophiolitic Jurassic magmatism occurs in the internal areas of Greece, whereas it is completely unknown in the external areas. In the latter a Triassic volcanism occurs, the Eastern extension of which has yet to be precisely determined.

DESCRIPTION OF THE MAIN OPHIOLITIC COMPLEXES IN GREECE

Numerous ophiolitic outcrops of varied size are known in the internal and intermediate Hellenic areas (fig. 1). In this paper, we shall consider only the main ophiolitic massif from the internal (Chalkidiki, Guevgueli, Almopias, Vourinos) and intermediate (Pindos, Othrys, Iti, Euboea) regions, the ophiolites belonging to the metamorphic blueschist formations of Ossa and Southern Euboea will also be rapidly discussed. In the next chapter we shall try to bring out and to discuss the main characteristics of the Hellenic ophiolites.

1. Ophiolites of Chalkidiki

Ophiolitic rocks in Chalkidiki strike NW-SE with a general dip of 60° - 80° NE. They present an axial extension of about 80 km from Thessaloniki to the southern parts of the Sithonia and Kassandra peninsulas ; their width does not exceed a few kilometres. They rest tectonically on highly deformed basement rocks of continental origin.

A study of these ophiolites is presently in progress (Jung et al., 1979 ; Burgath et al., 1979 ; Sapountzis, 1980 ; Dimitriadis, 1980). Several papers describing limited areas or dealing with metallogenetic subjects (chromitite, magnesite) are devoted to them (Hiessleitner, 1951 ; Monod, 1965 ; Ricou, 1965 ; Sapountzis, 1969, 1975, 1979 ; Kockel et al., 1971 et 1976).

Their coarse-grained mafic and ultramafic (Tabl. 1) are well exposed in the northern areas. Dunites with chromitite of podlike nature are interlayered with harzburgites in sequences which exhibit cumulate features ; these rocks are intruded by websterites and two-pyroxene gabbros. They are overlain by gabbroic cumulates composed of plagioclase (An_{95-91}), clinopyroxene ($Ca_{47}\ Mg_{46}\ Fe_7$-$Ca_{46}\ Mg_{42}\ Fe_{12}$), orthopyroxene ($Ca_3\ Mg_{79}\ Fe_{18}$ - $Ca_2\ Mg_{65}\ Fe_{33}$) and olivine ($Fo_{81.5-73.5}$).

These mafic and ultramafic rocks are tectonically overlain by slightly metamorphosed quartz-diorites, which are intensively intruded by a subparallel dyke swarm striking in the same regional direction as the complex. The quartz-diorites and the associated dykes have a more or less pronounced calc-alcaline character.

A sheeted dyke complex and pillow lavas (Tabl. 1) are exposed on the seashore in the southern parts of the Sithonia and Kassandra peninsulas. The lower dykes interfinger with a gabbro unit. In the uppermost part, the pillow lavas are intercalated by pelagic and detritus-rich sediments. A sulphide-bearing layer is associated with these lavas ; it grades upwards into pelagic manganese metasediments. The TiO_2 - FeO^*/MgO ratio in the volcanic rock suite follows the trend for abyssal tholeiites.

According to Burgath et al. (1979), the ophiolites of Chalkidiki formation started at the begining of the Jurassic by a rift phase : the "newly formed ocean" must have been rather small : lavas are associated with sediments of continental origin. Lower Tithonian molasse and shelf sediments transgressively cover the ophiolites. However according to Sapountzis (1979), radiometric ages of gabbros argue against a Mesozoic age of these ophiolites ; in fact, datations by K - Ar method yielded highly discordant results, ranging from 42 ± 8 my to 325 ± 30 my.

One can wonder, in these conditions, whether the basic and ultrabasic rocks of Chalkidiki belong to a single complex. It would be perhaps advisable to distinguish :

	1	2	3	4	5	6	7
SiO_2	40.06	41.63	34.80	47.26	46.94	45.57	47.20
TiO_2	0.02	0.01	0.11	0.04	0.16	0.73	0.50
Al_2O_3	0.48	0.59	1.30	12.98	18.72	20.38	18.15
Fe_2O_3	8.49	8.41	4.39	1.82	1.63	3.16	2.39
FeO	-	-	2.91	3.67	5.98	7.56	5.21
MnO	0.13	0.13	-	0.10	0.15	0.15	-
MgO	37.64	40.75	43.70	15.42	8.99	6.70	10.68
CaO	1.07	0.50	0.14	17.39	15.90	14.31	13.40
Na_2O	0.04	0.07	0.13	0.42	0.49	0.82	0.79
K_2O	0.01	0.01	0.08	0.02	0.02	0.03	0.10
P_2O_5	-	-	-	tr	0.01	0.01	-
L.O.I.	10.58	6.87	12.00	0.61	0.52	0.75	0.99
	98.52	98.97	99.56	99.73	99.51	100.17	99.41

Table 1 : Chemical composition of the Chalkidiki ophiolites

1. harzburgite from Triadhi, Bébien, unpub. data
2. harzburgite from Triadhi, Bébien, unpub. data
3. harzburgite, Desprairies et al, 1974
4. olivine-gabbronorite, Sapountzis, 1979
5. gabbronorite, Sapountzis, 1979
6. gabbronorite, Sapountzis, 1979
7. gabbro, Desprairies et al, 1974

- on the one hand, the gabbros and the ultrabasic rocks to the SE of Thessalonika with possibly an ante-Mesozoic age,

- on the other hand, the sheeted dyke complex and pillow lavas of Jurassic age in the Sithonia and Kassandra peninsulas. These last rocks would belong to a unit hypabyssal and eruptive formations with tholeiitic characteristics, stretching to the North towards the Yugoslavian border (Aspro Vrissi, Metalikon), and related to the Guevgueli complex.

2. Ophiolites of Guevgueli

The ophiolite complex of Guevgueli (Mercier, 1966 ; Bébien, 1977 ; Mercier and Bébien, 1977) outcrops in Greece on the Eastern side of the Paikon massif. It extends widely towards the North in Yugoslavia : its Yugoslavian part is described in this volume (part I) as the complex of Dren (Karamata et al., 1980). Its size is comparable to the largest ophiolitic complex of Greece : the Pindos one (60 km from North to South, 20 km from East to West). It presents the particularity of having the upper part of an ophiolitic complex (ultramafic rocks are scarce) closely associated with granites, microgranites and rhyolites.

Gabbros with cumulate textures are very abundant ; the cumulus minerals are clinopyroxene ($Ca_{48}\ Mg_{47}\ Fe_5$ - $Ca_{43}\ Mg_{43}\ Fe_{14}$) and plagioclase ($An_{90-50}$), the presence of olivine (Fo_{78-66}) is exceptional. Orthopyroxene ($Ca_2\ Mg_{72}\ Fe_{26}$) is an intercumulus phase in the uppermost gabbros were titanomagnetite appears. This latter phase is very important in the overlying ferrodiorites. In the Western part of the complex, these coarsed grained mafic rocks are covered by a thin and irregular layer of quartz-diorites, tonalites and trondhjemites.

Hypabyssal formations (Tabl. 2) consist of sills in the Western part and dykes in the Eastern part of the complex. The sills rest directly on the plutonic acidic layer ; they are made up of rocks with basic or intermediate composition, showing tholeiitic differentiation, with a marked FeO and TiO_2 enrichment (fig. 2) : This trend probably corresponds to the evolution of the residual liquids during the formation of the cumulate gabbros and ferrodiorites. The dykes form a sheeted complex ; their lower parts cut the underlying gabbros. They present well exposed flowage differentiation phenomena which can account for their petrographic diversity and for their weak enrichment in FeO^* and TiO_2 (Bébien et al., 1979 a et b ; Bébien et Gagny, 1979 a et b).

Effusive rocks rest on the hypabyssal formations. The lava flows are intercalated with volcanic breccias and hyaloclastites. Sedimentary rocks (radiolarites) are very scarce. The effusive layers are made up of spilitic pillow lavas, massive quartz-keratophyres and abundant rocks of intermediate composition.

The upper part of the effusive rocks is hidden by detritic formations of

	1	2	3	4	5	6	7	8	9	10
SiO_2	48.10	49.94	43.40	51.31	52.93	60.22	47.29	50.97	52.61	70.57
TiO_2	0.25	0.49	4.93	1.47	2.13	1.47	0.62	0.94	0.95	0.41
Al_2O_3	16.96	17.79	13.06	13.88	14.15	14.07	17.43	15.03	14.53	13.69
Fe_2O_3	5.47	6.12	19.54	12.00	13.17	10.45	7.37	8.74	8.24	2.83
FeO	-	-	-	-	-	-	-	-	-	-
MnO	0.10	0.07	0.33	n.d.	n.d.	n.d.	0.20	0.19	0.19	0.03
MgO	10.86	7.82	5.83	6.30	5.63	1.67	8.59	5.12	3.67	0.31
CaO	14.57	12.10	7.23	7.00	2.76	3.69	8.02	7.26	6.33	1.43
Na_2O	1.50	2.36	3.52	5.01	5.63	6.19	1.39	4.89	6.26	7.23
K_2O	0.16	0.39	0.28	0.13	0.09	0.05	4.19	1.51	0.53	0.48
L.O.I.	2.21	2.05	1.39	3.00	2.90	1.60	4.95	5.24	6.59	2.69
	100.18	99.13	99.51	100.10	99.39	99.41	100.05	99.89	99.90	99.67

Table 2 : Chemical composition of the Guevgueli ophiolites

1. gabbro
2. gabbro
3. ferrodiorite
4. sills from the Western Unit
5. sills from the Western Unit
6. sills from the Western Unit
7. doleritic dyke, porphyritic centre
8. same doleritic dyke, aphyric margin
9. microdiorite, selvage
10. microtonalite

1 to 10 data after Bébien (1981)

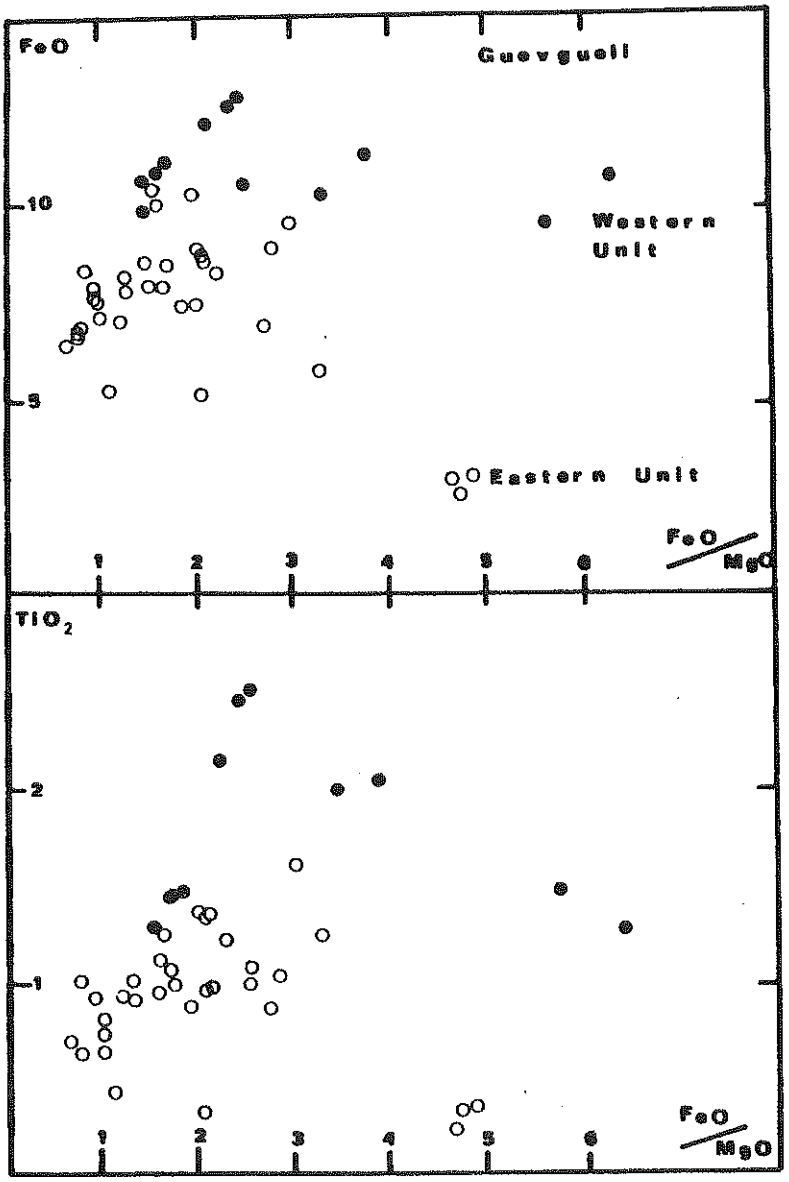

Fig. 2 - FeO total or TiO_2 versus FeO total/MgO for hypabyssal rocks of the Guevgueli ophiolites.

Pontian age ; however, a few kilometers to the East of the Guevgueli complex, associations of gabbros, diorites and effusives rocks forming little massif, and probably corresponding to members of the Guevgueli ophiolite scattered by tectonic phenomenon, are transgressively overlain by sandstone and by reef limestones. These latter give fauna of Middle and Upper Kimmeridgian and Portlandian age.

The Guevgueli complex contains magmatic breccias, exposed over an area about 15 km long by 3 km wide, and intruded into gabbroic cumulates. The matrix is made up of leucocratic rocks with dioritic or, more rarely, granitic compositions. The inclusions, doleritic or microdioritic, are irregularly shaped ; their border is often sinuous, crenelated and chilled, suggesting a simultaneous cooling of the matrix and the inclusions.

A granite massif, the Fanos granite, is intrusive into the ophiolites of Guevgueli ; the biotites of this granite, dated by K-Ar and Rb-Sr methods, are 150 my old (Upper Jurassic, Borsi et al., 1966). Biotite-orthoclase-sillimanite-cordierite migmatites (migmatites of Piyi) also occur along with the granite. All these rocks have been affected by diastrophic processes which occured during the formation of the complex. An important angular unconformity can be observed in the plutonic rocks : part of the gabbroic cumulates (gabbros I), chemically characterized by low FeO/MgO and $(Na_2O + K_2O)/CaO$ ratios, had a vertical magmatic layering during the crystallization of gabbro richer in iron and alkalis (gabbros II), of magmatic breccias and of hypabyssal and effusive rocks. One can observe in these gabbros I shear fractures, mechanical twinning of plagioclase and granulation phenomena, whereas the other rocks are generally undeformed. On the other hand, studies of structural petrology (Bébien and Gagny, 1978) in the sheeted complex and the magmatic breccias show that the ophiolites of Guevgueli were affected during their formation, by transcurrent movements with an extensional direction perpendicular to the layering of the gabbros I. Gabbro from the ophiolite of Guevgueli rests tectonically on the volcanic and sedimentary formations of Jurassic age in the Paikon massif. In the other hand, in the Southern part of the complex, ophiolitic effusives rest conformably on Upper Jurassic limestones.

3. Basic and ultrabasic Almopias formations

The area situated between the Pelagonian massif on the West and the Paikon massif on the East (Almopias zone, Mercier, 1966) is rich in basic and ultrabasic formations, but, owing to its tectonic complexity, it does not conveniently lend itself to a study of the ophiolites. However, this region is regarded, in certains models, as the source of the ophiolites thrust westwards.

In this Almopias zone, under thick middle and upper Cretaceous deposits, one can recognize two kinds of formations (Mercier, 1966 ; Mercier and Vergely, 1977) :

	1	2	3	4	5
SiO_2	57.67	52.44	52.68	47.19	51.18
TiO_2	1.08	0.71	1.66	1.73	2.55
Al_2O_3	17.18	15.49	14.82	13.85	12.78
Fe_2O_3	7.22	7.94	9.39	12.05	15.72
FeO	-	-	-	-	-
MnO	0.13	0.18	0.48	0.29	0.24
MgO	3.00	8.01	7.39	6.99	4.07
CaO	4.95	7.12	2.88	12.80	6.14
Na_2O	5.35	3.50	4.85	2.03	4.84
K_2O	0.63	1.97	0.26	0.19	0.15
L.O.I.	2.28	2.18	5.03	3.78	2.20
	99.49	99.54	99.44	100.90	99.87

Table 3 : Chemical composition of the Almopias "ophiolites"

1. quartz-diorite
2. basalt
3. basalt
4. basalt
5. ferro-basalt

1 to 5 after Bechon (1980)

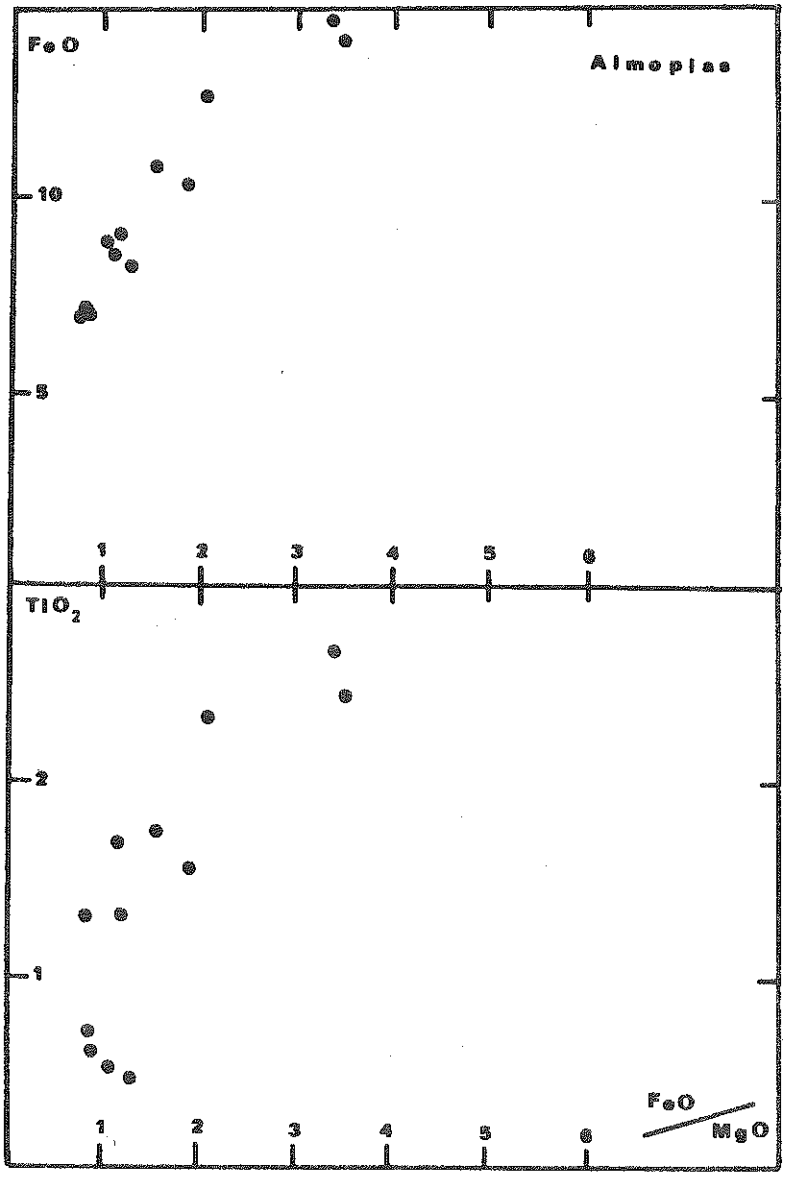

Fig. 3 - FeO total or TiO_2 versus FeO total/MgO for effusive and hypabyssal rocks of the Almopias formation.

- basic and acid tuffites, limestones interstratified with "Nerinees" limestones,

- pillow lavas, volcanic breccias and dolerites occasionally associated with radiolarian cherts.

The formations are often associated with abundant serpentinites showing a chaotic aspect (coloured mélange). Mercier and Vergely demonstrated that these mélanges are mainly of tectonic origin. They result largely from important transcurrent movements latter than the JE_{1-2} (Upper Jurassic - Eocretaceous tectonic phases and earlier than the Upper Cretaceous. These transcurrent movements affect a chaotic sedimentary and volcanic unit formed at the base by a sheet principaly made up of ultrabasic rocks and emplaced during the JE_1 phase.

In the Western area (Northern Vermion), associations of plutonic and volcanic basic and acid rocks make up the lower part of tectonic slices thrust onto the Pelagonian massif (Braud, 1967 ; Braud and Brunn, 1972). Serpentinites occur probably arising from peridotites, pyroxenites, metagabbros, quartz-diorites, metadolerites, spilites and keratophyres ; acid rocks are rather abundant. Detritic Cenomanian - Turonian formations overlie these eruptive rocks.

In the Eastern part of the Almopias zone, important massifs of effusive and hypabyssal rocks outcrop (Bechon, 1980). The pillow lavas are often broken up into decimetric fragments cemented by hyaloclastites. Sedimentary intercalations, when they are present, are made up of radiolarian cherts. These volcanic formations are locally intruded by swarms of dykes or sills with an approximatively N-S direction. The rocks the least affected by weathering or metamorphism are made up of plagioclase, clinopyroxene and sometimes abundant titanomagnetite. From a chemical point of view (Tabl. 3) they present very low K_2O contents, and the FeO^* and TiO_2 contents are very variable. The analysed samples define a trend comparable with the abyssal tholeiites one (fig. 3).

Neither the geological setting, nor the petrographic and geochemical characteristics permit us to distinguish these formations from the different parts of many ophiolitic complexes : the presence, in the Almopias zone, of ophiolites scattered by tectonic phenomena is very probable.

4. The Vourinos

The Vourinos ophiolitic complex lies on the Western side of the Pelagonian massif. It is about 30 km long by 15 km wide and is divided into two main units : the Western Vourinos unit (Krapa and Asprokambo) 5 by 8 km and the Eastern 30 by 10 km (Moores, 1969). This complex rests tectonically on the Pelagonian paleozoic schists and their marble cover (Brunn, 1956) dated by Diplores and Halobies of Triassic age, and on amphibolites, amphibole-schists and tuffaceous schists with

pillow lavas alternating with marbles and jaspers (Pichon and Brunn, 1977). Theses rocks beneath the ophiolites have been interpreted by Zimmerman (1972) as a "mélange". However Naylor and Harle (1976) have shown that the Ayios Nicolaos complex ("the mélange") is a clastic sequence composed of shales, siltstones, tuffs, radiolarian cherts and olistostromes. These authors assign a late Jurassic to early Cretaceous age by analogy with Othrys. The amphibolite dated by $^{40}Ar/^{39}Ar$ gave an age of 179 ± 4 m.y. (Spray and Roddick, 1980). The tectonic origin of the ophiolite is controversial : for Naylor and Harle (1976, 1977) and Smith (1977), it was thrust from the Othrys zone with a movement towards the NE (D_2 phase = JE_2), whereas for Vergely (1975, 1976, 1977) the Vourinos originated in the "Vardar zone" with a displacement towards the SW during the upper Jurassic (JE_1 phase) and was backthrust during the lower Cretaceous (JE_2 phase) towards the East. It must be noted that the folding at the base of the ultramafics is comparable to the deformation affecting the volcano-sedimentary substratum of the ophiolites (Vergely, 1976). The main rocks of the ophiolitic complex are ultramafics, mainly harzburgites with tectonic fabrics forming a large massif. The harzburgites are composed (Harkins et al., 1980) of 5 - 20 % enstatite ($En_{90.7-89.3}$; $Wo_{1-1.5}$; $Fs_{8.2-9.1}$), 80-90 % Ol ($Fo_{91.8-90.8}$) ; 1-3 % Sp, and 1 % or less clinopyroxene ($En_{49.2-50.5}$; $Wo_{48.2-46}$; $Fs_{2.6-3.5}$). The peridotites are supposed to have arisen in a diapir at a spreading centre (Moores et Vine, 1971 ; Ross et al., 1980) during the formation of the oceanic crust. The peridotite massif has been divided into three areas (Moores, 1969). In the central part coarse grained rock, mainly dunite, peridotite and chromitite is slightly deformed and arranged in isoclinal folds with steep axial plunge varying widely in direction (Moores, 1969). The poikiloblastic texture shown by these rocks is supposed to have formed in the diapir at a depth of 45 - 70 km in which partial fusion occurs (Ross et al., 1980) under a high pyroxene geotherm (Mercier, 1980). At a lower depth under higher mylonitisation a new foliation - S_2 - was produced (Ross et al., 1980). Towards the Eastern border the layering has a flatter dip (Moores, 1969). Pyroxenite abundant in the central part is rare in the NW part, and occurring parallel to the layering or in dykes perhaps related to conjugate joint systems is younger than the chromitite-dunite and the dunite-peridotite layering and may represent interstitial magmatic liquid (Moores, 1969). The podiform chromites (Maratos, 1964 ; Karamata, 1966) exhibit isoclinal folds with the development of axial-plane structures (Ayrton, 1968). They may also occur as pipelike, columnar and lenticular deposits (Zachos, 1964). The formation of the dunite bodies is linked to multiple intrusive events before and after the deformation (Harkins et al., 1980). Due to similar olivine fabric in both chromitite-dunite and harzburgite, a common origin by accumulation during laminar flowage has also been envisaged accompanied by syntectonic recrystallization and intra-crystalline gliding (Burgath and Weiser, 1980). Olivine as well as chromite have compositions similar

to the early cumulus phases (Harkins et al., 1980).

The transition zones between the ultramafic and mafic rocks comprise dunites interlayered with websterites and gabbros (Moores, 1969). This stratiform unit is composed of cyclic units with strong lineated lamination. They are characterized by the crystallization of : Ol, Chr, Ol, Ol + Cpx, Cpx, Cpx + Opx, Cpx + Opx + Pl, Cpx + Pl + Hb (Jackson et al., 1975). The S_2 foliation may be recognized at the base of the cumulate sequence (Ross et al., 1980).

Within each cycle chromite, olivine and pyroxene show fractionation trends. The gabbroic cumulates, mainly two pyroxene-anorthite gabbro-norites may contain up to 8 % of iron-titanium oxides. Banded Opx-Cpx-An gabbros have been observed by Rassios (1979) in the ultramafic cumulates. The upper part (Tabl. 4) of the complex characterized by the appearance of primary poikilitic amphiboles shows transition from noritic gabbros to diorites, quartz-diorites and granophyres (Moores, 1969 ; Paupy, 1976).

At the top, appear magmatic breccias (Brunn, 1956 ; Bortolotti et al., 1969). Three main types have been described by Paupy (1976) :

- a magmatic breccia of dioritic composition with xenoliths of amphibole, plagioclase and pyroxene clusters favoured perhaps by hydrothermal circulation.

- cumulative rocks reworked in a dynamic environment in contemporaneous basic and acid liquids. Corrosion and remelting of minerals (e.g. plagioclase) occur. Microdolerites considered to be the roof of the magmatic chamber (Bébien et al., 1975) have also been included in magmas.

- magmatic breccias between diorites and microdolerites which show interaction between two magmas may represent magmatic stoping (Brunn, 1956).

In the mafic cumulates clinopyroxene ranges from $En_{46.2}$, Wo_{49}, $Fs_{4.8}$ in pyroxenite towards $En_{40.7}$, $Wo_{48.3}$, Fs_{11} in the gabbros, and orthopyroxens varies from En_{84}, $Wo_{2.6}$, $Fs_{13.4}$ to $En_{67.3}$, $Wo_{3.2}$, $Fs_{29.5}$. The An content in plagioclase from the gabbros is always high $An_{95.7}$ to $An_{83.7}$ except in the dioritic term, where it is only $An_{38.7}$ (Ohnenstetter unpub. data). Olivine evolves from $Fo_{92.9}$ in dunite in the lower cumulates (Harkins et al., 1980) to $Fo_{83.7}$ in pyroxenite. Above the magmatic breccia zone, diorite and quartz-diorites from the Asprokambo area are intruded by several dykes. Massive granophyres crop out in the Mandania river area.

Hypabyssal rocks occur as dykes and sills. In the Asprokambo area, the dykes are dolerites, porphyritic basalts, basaltic andesites and dacites (Tabl. 5) On the contrary, on the Krapa hills hypabyssal rocks are mainly sills with doleritic or microlitic fluidal textures with a compositions evolving also from basalt to dacite (Tabl. 5). Above occur massive or pillowed lava flow (Paupy, 1976). Ewing (1977) has noted that three different transition may exist between the sheeted dyke unit and the magmatic cumulate suite :

	1	2	3	4	5	6	7	8	9
SiO_2	37.70	43.47	42.95	51.54	47.08	48.25	53.50	68.82	73.84
TiO_2	tr	0.03	0.01	0.07	0.09	0.44	0.32	0.29	0.23
Al_2O_3	0.28	0.61	1.09	1.96	19.66	15.43	15.95	12.52	11.75
Fe_2O_3	2.64	1.23	6.52	5.30	6.78	8.11	10.21	4.70	4.60
FeO	4.63	6.63	4.17	-	-	-	-	-	-
MnO	0.10	0.12	0.17	0.10	0.09	n.d.	n.d.	0.12	0.09
MgO	48.09	45.02	31.50	18.35	7.85	6.15	5.45	1.56	0.97
CaO	0.04	0.60	5.15	20.51	14.05	10.19	8.17	6.80	1.47
Na_2O	0.17	0.56	0.05	-	1.09	3.30	3.33	2.43	3.37
K_2O	0.15	0.14	0.10	-	0.05	0.46	0.34	0.14	0.61
P_2O_5	0.01	0.02	0.02						
H_2O^+	2.45	0.75	4.52	1.98	3.20	7.90	2.82	2.20	2.15
H_2O^-	2.49	0.48	2.39						
CO_2	1.23	0.28	0.72						
Cr_2O_3	0.27	0.40	0.48	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
NiO	0.35	0.29	0.12	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
	100.60	100.63	99.96	99.71	99.94	100.23	100.09	99.58	99.08

Table 4 : Chemical composition of the Vourinos ophiolite : plutonic members

1. dunite
2. harzburgite-tectonite
3. lherzolite-cumulate
4. websterite
5. gabbro
6. gabbro
7. diorite
8. quartz-diorite
9. granophyre

1 and 2 after Moores (1969), 3 to 9 after Paupy (1976)

	1	2	3	4	5	6	7	8	9
SiO_2	51.85	53.98	55.35	68.38	72.40	48.92	53.65	58.50	64.60
TiO_2	0.44	0.41	0.30	0.32	0.25	1.01	0.99	0.78	0.93
Al_2O_3	14.35	14.90	13.58	12.90	10.90	18.42	15.40	15.12	13.51
Fe_2O_3	8.68	8.70	9.80	6.57	5.50	12.52	10.95	10.52	9.16
MnO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	6.75	5.92	7.76	1.72	1.45	4.12	4.92	4.55	2.63
CaO	12.06	10.97	7.22	2.60	3.10	8.15	3.03	2.86	1.30
Na_2O	1.21	0.75	2.86	4.18	3.51	3.85	5.59	4.41	5.22
K_2O	< 0.01	< 0.01	0.17	0.12	0.03	0.33	0.05	0.37	0.15
P_2O_5	0.02	0.04	0.01	0.03	0.04	0.06	0.05	0.06	0.07
L.O.I.	3.66	3.44	3.12	2.06	1.75	1.78	3.87	3.16	2.62
	99.02	99.11	100.17	98.88	98.93	99.16	98.50	100.33	100.19
Zr	29	32	19	36	35	58	39	48	60
Cr	402	285	289	92	81	28	69	68	68
Ni	105	96	92	31	15	30	45	27	13

Table 5 : Chemical composition of the Vourinos ophiolite : hypabyssal and effusive rocks

1. basalt from Asprokambo
2. basaltic andesite from Asprokambo
3. basaltic andesite from Asprokambo
4. dacite from Asprokambo
5. rhyodacite from Asprokambo
6. basalt from Krapa
7. basaltic andesite from Krapa
8. andesite from Krapa
9. dacite from Krapa

1 to 9 from Paupy (1976) and Beccaluva et al (submitted)

- basaltic to andesitic dykes intruded into quartz-hornblende-diorites gradational zone, between fine grained dyke material and coarser diorites, with magmatic lineations parallel to the dykes margins
- intrusion of dykes into cumulate gabbros.

The effusive and hypabyssal terms, mainly basaltic andesites, evolve chemically from low-K tholeiitic basalts towards dacites (fig. 4). Besides clinopyroxene, and orthopyroxene, amphibole and alkali feldspar have been reported (Ohnenstetter unpub. data). The chemical variation of the liquids reflects the removal of the main phases in the cumulates. More or less perfect fractional crystallization trends (Ohnenstetter et al., 1979) are marked by early clinopyroxene and late plagioclase fractionation at a mafic index of about 50. The existence of smalls magma chambers with different accumulative processes as in the Krapa and Asprokambo areas, respectively characterized by adcumulates and orthocumulates has been proved by trace element behaviour (Beccaluva et al., submitted). The geochemical evolution of liquids exhibits large discrepancies with that of the Pindos even if REE are considered (Montigny, 1975). Firstly, dolerite from Vourinos presents a slight LREE enrichment compared to the flat REE pattern of the Pindos. Secondly the cumulates in Vourinos show slight to absent europium anomaly due to the restricted role played by plagioclase (Montigny, 1975, Cocherie, 1978). Therefore, the petrographic and geochemical evolution of the Vourinos is more compatible with an origin in an island-arc setting (Miyashiro, 1973 ; Beccaluva et al., 1978, 1979).

Radiolarian cherts of upper Jurassic age overlie and cement the ophiolitic pillow lavas. They give the following fauna : *Amphisphaera, Cenosphaera, Ropalastrum* and *Dictomitra* (Deflandre in Brunn, 1956) followed by Calpionelids limestones (Moores, 1969). The Langadhakia cross-section above the Krapa hill has been studied from a geological and a micropaleontological point of view (Pichon, 1976, 1977 ; Pichon and Lys, 1976 ; Mavridis et al., 1977). Biomicrites and bioclastic limestones contain *Stomiosphaera moluccana* Wanner, *Lagenidae, Saccocoma sp., Aptychus,* polyps : *Cladocoropsis mirabilis* Felix, alguas : *Thaumatoporella parvovesiculifera* Raineri, passing into Calpionella micritic limestones (*Calpionella alpina* Lorenz). This pelagic microfauna overlying the ophiolitic complex gives a Tithonian to Berriasian age. The sedimentation presents both pelagic and reef facies. After an emergence period corresponding to the JE_2 tectonic phase, unconformable calcareous sandstones contain *Globigerinelloides* of Aptian ? - Albian age. A second hiatus may perhaps exist before the Cenomanian transgression. These bioclastic to clastic limestones contain neritic microfauna alternating with pelagic levels, corresponding to a continental slope sedimentation.

To summarize, three main events (Pichon and Lys, 1976) occur after the deposition of the radiolarian cover on the ophiolitic lavas and before the upper Cretaceous transgression :

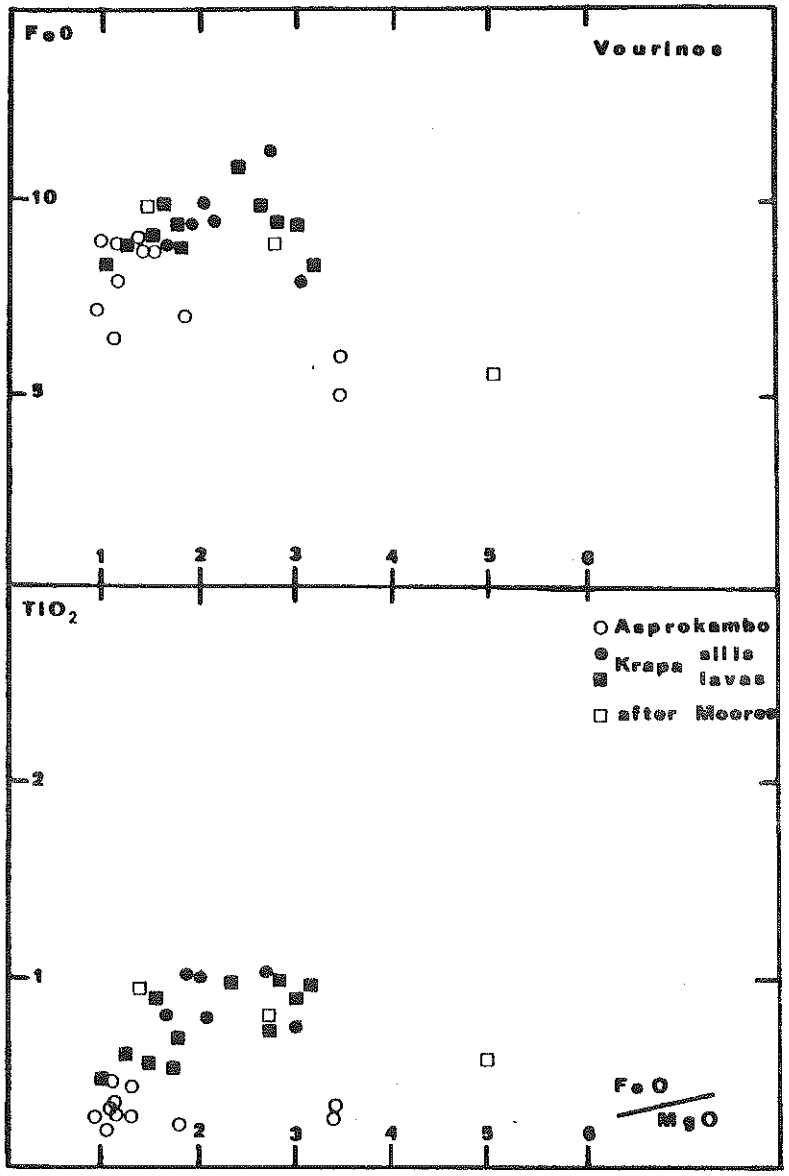

Fig. 4 - FeO total or TiO_2 versus FeO total/MgO for effusive and hypabyssal rocks of the Vourinos ophiolites.

- the transgression of Tithonian to Berriasian age with the pelagic Langadakia limestones
- a post-Berriasian ante-Cenomanian tectonic phase (JE_2) with erosion and uplift
- a transgressive series of rhythmic settling of Albian ? to Cenomanian with cornstone, limestone, volcanodetritic sandstone, biodetritic and bioclastic limestones characteristic of the alternance of pelagic and neritic levels of a continental slope.

5. Ophiolites of Vermion and Zygosti

Several small outcrops principally of serpentinite can be observed in the Pelagonian massif between the Almopias zone and the Vourinos ophiolites.

In the Pelagonian part of the Vermion massif, near the western border of the Almopias zone, ophiolites stretch in a N - S belt about forty kilometers long for a width smaller than a few hundred meters. These rocks are largely serpentinised and eroded ; however, in some places, one can recognize olivine-hypersthene-chromite-harzburgites (Pichon, 1976). At their base amphibolite sheets sometimes appear. These formations rest on Pelagonian schists and marbles comparable to those described under the Vourinos massif.

A calcareous and sandstone unit transgressively overlies these ophiolites or sometimes the Pelagonian substratum. Some determinable fossils give a probably Upper Jurassic - Lower Cretaceous age. After a tectonic phase and an emersion, a middle Cretaceous (Vraconian-Cenomanian) formation is deposited.

At the Southern end of the ophiolite belt of the Vermion (Ayios Dimitrios area) has been described (Vergely, 1976) a synmetamorphic fold with a SW vergence (JE_1 phase) reaffected by decametric to hectometric folds with an E vergence (JE_2 phase).

In the Zygosti, a few kilometers eastward from the Vourinos massif, ophiolites with chromitites form a narrow massif (Brunn, 1956). Their radiolarian cover is overlain by a crystalline limestone unit with an Upper Jurassic and Lower Cretaceous fauna. Upper Cretaceous strata discordantly rest either on the ophiolites or on the crystalline limestones (Pichon, 1977).

Thus, like those of the Vourinos, the Vermion and the Zygosti ophiolites rest upon Pelagonian marbles and schists, and are discordantly overlain by formations of Upper Jurassic and Lower Cretaceous age. This cover is detritic eastward (Vermion), where its deposition follows an important erosion of the ophiolites, but it is calcareous westward (Vourinos). On the other hand, in the Vourinos and in the Vermion, one can recognize synmetamorphic folds belonging to the JE_1 phase. Hence, the Vourinos, Zygosti and Vermion ophiolites would belong to the same sheet thrust westward, and arising from the Almopias zone.

6. The Pindos

The Pindos ophiolitic complex was well described since Brunn (1956). The ophiolite was thrust towards the West overly the Eocene flysch of the Pindos-Olonos zone and is overlain unconformably by molasse of Oligocene age (Brunn, 1956). Between the ultramafics and the Eocene flysch four formations may exist :

- an upper Jurassic flysch
- a volcano-detritic unit
- volcanic series of Triassic age
- amphibolites beneath the serpentinized harzburgites. The flyschoid series of Upper Jurassic to Eo-Cretaceous age (Berriasian) found in the Pindos (Terry and Mercier, 1971) and called the Beotian flysch towards the South (Celet and Clément, 1971) is systematically found between the Internal and External zones of the Dinarides (the Bosniac flysch of Blanchet et al., 1969).

Parrot (1967) has described near Avomintsa a series (150 m) composed of radiolarian cherts and limestones with calpionellids : *Calpionella alpina* (Brunn, 1956). The volcano-detritic series (Dercourt et al., 1977) has been subdivided into three units (Terry, 1975) :

- the graywackes and conglomerates of ophiolitic and Triassic volcanic material from Panagia of Dogger to Malm age (*Nautiloculina sp.*, and *Protopeneroplis striata*),
- the graywackes and sedimentary klippes near Kandila ridge,
- the pelite and sedimentary klippes from Avommintsa enclosing lenses of *Halobia* limestones of Triassic age.

The reworked and redeposited ophiolitic material reflects an initial Jurassic tectonism comparable to the event described in the "Vardar" zone by Mercier (1966) and Vergely (1976) and in Yugoslavia by Dimitrijevic and Dimitrijevic (1973, 1976) and Charvet (1980).

Near Avdela village a volcanic series up to 200-300 m of Triassic age is considered as the equivalent of the Diabas-Hornstein formation of Yugoslavia. The Conodontes (*Gondolella sp.*) limestones cement the volcanic breccias (Terry, 1971). The volcanics of alkalic affinity are mainly vesicular, Pl-Cpx and sometimes Ol porphyritic pillow lavas associated with scarce doleritic dykes, dacites and keratophyres (Terry, 1979) - see Tabl. 6).

Slices of amphibolites (T = 400-500°C, P = 2 Kbar) have been described in detail by Whitechurch and Parrot (1978) near Pericoli and Milea. In this latter area the metamorphic rocks lie between the peridotites and the Eocene flysch and are composed from base to top of gabbros, lavas and limestones included within micaschists corresponding to the tectonic base of the slice, followed by quartzites rich in muscovite, quartzose micaschist and amphibole-schist. Near Metsovo the amphibolite sometimes Pl or epidote bearing are tectonically included in the serpentinites as well as near

	1	2	3
SiO_2	50,67	48.40	72.75
TiO_2	1.24	1.40	0.35
Al_2O_3	18.82	17.38	13.64
Fe_2O_3	9.17	9.99	3.1
MnO	0.15	0.19	0.06
MgO	4.98	8.69	0.24
CaO	9.21	9.17	2.06
Na_2O	4.42	3.94	6.28
K_2O	1.18	0.78	1.54
L.O.I.	(4.40)	(7.16)	(2.00)
	99.84	99.94	100.02

Table 6 : Chemical composition of the Pindos Triassic alkaline effusives

1. dolerite, mean of 6 analyses
2. pillow lava, mean of 8 analyses
3. keratophyre, mean of 3 analyses

1 to 3 recalculed water free, after Terry (1979)

Perivoli, where biotite and garnet may also be present. These amphibolites have been folded three times. North to Metsovou amphiboles from a hornblende - Pl An_{23-32} - apatite - sphene - pyrite - saussurite - epidote - calcite - amphibolite (see Tabl. 7) of the metamorphic sole has been dated by $^{40}Ar/^{39}Ar$ at 172 ± 3 m.y. (Spray and Roddick, 1980). This confirm the age of 182 ± 4 m.y. given by Roddick et al. (1979) and led to the conclusion of these authors that this age may be the earlier Jurassic period of ophiolite emplacement. Similar ages (172 ± 5 m.y.) were obtained by Thuizat et al. (submitted) by the K-Ar method.

Foliated peridotites compose 2/3 of the whole ophiolitic massif. The peridotites are mainly harzburgites, dunites and plagioclase lherzolites. Layers of pyroxenites with strong Opx lineation are abundant in blastomylonitic Pl-free peridotites. The foliation planes are parallel to the pyroxenite layering. Some pyroxenites dykes are considered as the last partial melting product of the ultramafics (Montigny et al., 1973). The Pl-peridotite crops out in the Mount Gomara and Vassilitsa. The plagioclase is concentrated in clinopyroxene-rich bands. Gabbroic dykelets cuts the foliation at a low angle. All gradations exist from dunite-harzburgite to Pl-peridotite (Terry, 1974).

	1	2	3	4	5	6
SiO_2	48.52	46.65	46.09	44.87	45.44	46.03
TiO_2	2.43	1.61	2.49	2.38	1.40	3.06
Al_2O_3	13.49	14.05	13.51	18.23	15.02	14.05
Fe_2O_3	4.17	2.78	4.05	0.33	1.32	4.37
FeO	8.35	6.62	9.40	10.47	10.17	9.36
MnO	0.22	0.15	0.18	0.06	0.01	0.18
MgO	4.19	6.85	6.16	7.98	9.00	7.51
CaO	11.77	17.94	14.02	10.38	12.80	10.48
Na_2O	- 3.50	1.85	2.20	2.82	2.25	2.54
K_2O	1.10	0.11	0.49	0.31	0.05	0.53
P_2O_5	0.16	0.20	0.36	0.31	0.06	0.48
H_2O^+	1.21	1.19	1.12	1.86	2.48	1.41
H_2O^-	0.24	-	-	-	-	-
CO_2	-	-	-	-	-	-
	99.35	100.00	100.07	100.00	100.00	100.00

Table 7 : Chemical composition of the amphibolite soles beneath the Greek ophiolites

1. amphibolite from Euboea
2. amphibolite from Euboea
3. amphibolite from the Vourinos
4. amphibolite from Othrys
5. amphibolite from Othrys
6. amphibolite from the Pindos

1 from Parrot and Guernet (1972), 2 to 6 from Spray and Roddick (1980)

The gabbroic dykes are rodingitized in two stages : the first one occurs probably under hydrothermal conditions within the oceanic litosphere, the second one may be related to the ophiolite emplacement (Capedri et al., 1978). Terry (1974) has also observed brecciated foliated peridotites cemented by plagioclase peridotite having cumulus textures within the foliated peridotites.

The ultramafic cumulates : pyroxenite (enstatitites and websterites) wehrlites and Pl-peridotites grade in a few meters into allivalites (Brunn, 1956 ; Parrot, 1967, 1969). Zr reaches a maximum in the Pl-peridotites (Parrot and Rambaud, 1969).

The mafic cumulates (Brunn, 1956 ; Parrot, 1967 ; Terry, 1974 ; Capedri et al., 1980) are composed of troctolite, eucrite (76 % Pl (An_{80}), 17 % augite, 3 % bronzite, 2 % Ol (Fo_{80}), 2 % opaque minerals), gabbro with doleritic textures (57 % Pl (An_{65}), 24 % augite, 1 % bronzite (En_{70}), 18 % alteration products and few opaques minerals) and hornblende-gabbro (52 % Pl (An_{65}), 40 % green Hb, 4 % opaque minerals and 4 % transformation products). The order of crystallization is Ol - Chr Sp - Pl - Cpx - Opx, and the minerals often exhibit adcumulus growth (Capedri et al., 1980).

The cumulate may sometimes be deformed before the tectonic emplacement of the ophiolite and before dyke injection. These metamorphisms have been related to an ocean-floor metamorphism of amphibolite and greenschist facies contemporaneous with igneous processes (Terry, 1973). Within the plutonic rocks, olivine evolves from Fo_{91} in harzburgite, $Fo_{88.1-85.1}$ in allivalites and troctolites, to $Fo_{80.3}$ in Ol gabbros. Orthopyroxene ranges from $En_{86.9}\ Wo_{5.5}\ Fs_{7.5}$ in harzburgite $En_{86.1}\ Wo_{3.8}\ Fs_{10.1}$ in pyroxenite, $En_{92.8}\ Wo_{12.1}\ Fs_{10.9-6.2}$ in troctolite, to $En_{85.6}\ Wo_{3.4}\ Fs_{10.8}$ in gabbros.

Clinopyroxene evolves from diopside ($En_{52.3-49.4}\ Wo_{46.5-47.7}\ Fs_{5.7-4}$) and endiopside ($En_{52.4}\ Wo_{40.2}\ Fs_{7.4}$) in Pl-peridotite and troctolite to diopside ($En_{46.3}\ Wo_{40.2}\ Fs_{7.4}$) and augite ($En_{45.8}\ Wo_{37.4}\ Fs_{16.8}$) in gabbro. Plagioclase evolves from bytownite ($An_{87.5}$) in eucrite to labradorite ($An_{61.7}$) in gabbro and andesine ($An_{42.2}$) in diorite. Hornblende in gabbro has a Mg/Mg+Fe+Mn ratio ranging from 0.87 to 0.68 (mineralogical data from Parrot and Verdoni, 1976 ; and Venturelli unpub. data).

The transition between gabbro and lava is characterized by multiple dyke injections and by quartz-diorite and granophyre intrusions with magmatic breccias (Terry, 1974).

The effusive rocks (up to 450 m thick) are mainly pillow lavas. The volcanics exhibit a spilitic paragenesis : actinolite, chlorite albite, quartz, titanite, pumpellyite, zeolites. In some samples primary assemblages may be conserved : augite, pigeonite, labradorite (Terry, 1974). The pillows have textures ranging from pilotaxitic to intersertal and subophitic. They are mainly aphyric but some Cpx, Pl and

	1	2	3	4	5	6	7	8	9	10
SiO_2	48.60	52.40	55.70	61.80	47.80	48.40	48.90	48.50	52.50	54.60
TiO_2	1.11	1.35	1.33	0.92	1.64	0.87	1.30	0.20	0.21	0.46
Al_2O_3	15.20	14.30	14.10	13.40	14.90	17.60	14.00	10.40	11.20	14.30
Fe_2O_3	11.60	12.90	12.40	9.60	11.80	9.59	12.10	10.60	9.39	11.10
MnO	0.13	0.17	0.14	0.11	0.17	0.14	0.18	0.17	0.15	0.16
MgO	3.10	5.25	4.78	3.31	8.54	7.07	8.00	10.20	11.30	5.42
CaO	11.80	4.50	3.08	2.56	9.05	11.60	11.30	15.80	8.75	7.00
Na_2O	0.96	4.54	6.11	4.64	3.21	3.64	2.54	0.16	4.23	4.01
K_2O	0.24	0.48	0.05	0.27	0.09	0.03	0.08	0.03	0.05	0.23
P_2O_5	0.08	0.08	0.11	0.13	0.12	0.07	0.10	0.02	0.02	0.03
L.O.I.	7.20	4.00	3.20	3.20	2.70	1.00	1.60	3.90	2.20	2.70
	100.02	99.97	100.02	99.94	100.02	100.01	100.10	99.98	100.00	100.01
Ni	15	10	12	7	177	70	65	1830	240	25
Cr	11	10	9	7	380	270	200	550	750	45
Zr	58	58	58	97	113	40	65	9	9	22

Table 8 : Chemical composition of the Pindos ophiolite : effusive and hypabyssal rocks

1. pillow lavas
2. pillow lavas
3. pillowlavas
4. pillow lavas
5. massive basalt
6. dolerite
7. dolerite
8. dyke
9. dyke
10. dyke

1 to 10 after Capedri et al. (1980)

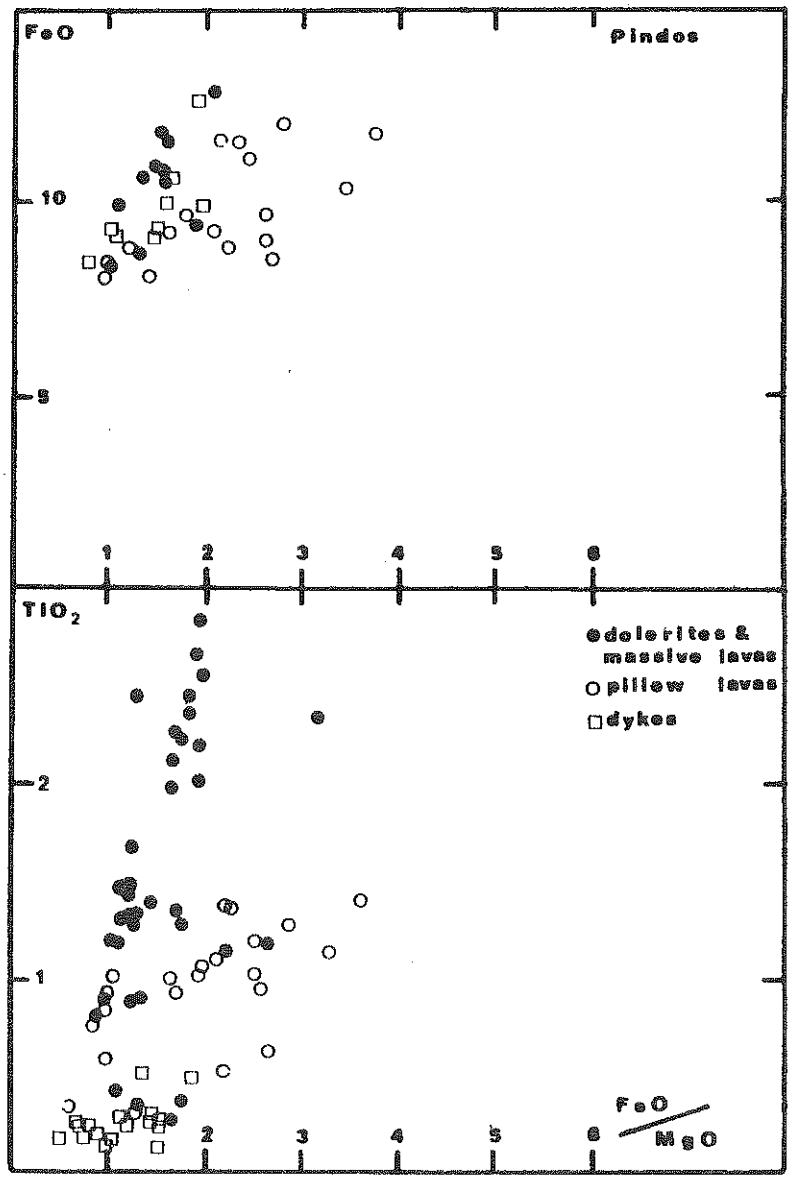

Fig. 5 - FeO total or TiO_2 versus FeO total/MgO for effusive and hypabyssal rocks of the Pindos ophiolites.

	1	2	3	4	5	6	7	8
SiO_2	37.56	41.80	39.30	41.25	41.50	47.70	47.88	72.62
TiO_2	0.03	0.10	0.25	0.20	0.06	0.15	0.37	0.54
Al_2O_3	0.55	2.45	5.60	15.80	10.70	22.70	18.47	11.84
Fe_2O_3	9.49	3.55	6.35	3.75	9.90	4.96	7.06	6.39
FeO		2.45	5.80	2.30				
MnO	0.11	n.d.	n.d.	n.d.	0.13	0.07	0.08	0.21
MgO	37.99	36.40	26.20	23.55	24.00	7.81	6.99	2.10
CaO	0.83	5.25	6.55	8.90	6.49	13.00	13.11	2.25
Na_2O	0.19	0.05	0.40	0.25	0.48	1.35	3.56	3.66
K_2O	n.d.	tr	0.05	0.05	< 0.02	< 0.02	0.05	0.51
P_2O_5	n.d.	n.d.	n.d.	n.d.	< 0.02	0.02	0.07	n.d.
L.O.I.	12.66	8.55	8.65	3.80	6.70	2.20	2.88	1.70
	99.41	100.60	99.15	99.85	99.96	99.96	100.52	100.12
Ni	1700	380	2200	-	1100	275	55	-
Cr	1930	6000	2400	-	335	590	115	-

Table 9 : Chemical composition of the Pindos ophiolite : plutonic members

1. serpentinized dunite
2. lherzolite
3. plagioclase peridotite
4. allivalite
5. troctolite
6. gabbro
7. gabbro
8. granophyre, near of 3 analyses recalculed water free

1 and 7 after Montigny et al. (1973)
2, 3, 4 after Parrot (1967) and Parrot and Rambaud (1969) for the trace elements
5 and 6 after Capedri et al. (1980)
8 after Terry (1979)

Ol phyric samples exist also. The massive basalt has intersertal to subophitic texture and is rarely porphyritic : Ol, Cpx, Pl phenocrysts (Capedri et al., 1980). The dolerite is composed of 51-43 % of Pl (An_{55}), 32-38 % Cpx (augite and sometimes pigeonite) 1 to 4 % of opaque minerals. Orthopyroxene has also been described in a sample (Parrot, 1967). Most of the clinopyroxene from the dykes and the pillow lavas exhibits low TiO_2 content 0.05 to 0.14, whereas the dolerites and massive basalts show higher TiO_2 values ranging from 0.44 to 0.87 (Capedri and Venturelli, 1979 ; Capedri et al., 1980).

The discrepancies between dolerite and lava have been emphasized by Paupy (1976), Beccaluva et al. (1979), Terry (1979) and Capedri et al. (1980). These latter authors distinguish four groups :

- A massive basalts
- B dolerites
- C pillow basalts
- D dykes.

The dyke rocks show very low Zr, P, Y and TiO_2 contents (Tabl. 8) and also have low-Ti pyroxenes ; they correspond to Terry's (1979) second group, made up of the latest dykes cutting cumulates. Capedri et al. (1980) concluded that the pillow lavas are not co-magmatic with the dolerites and the massive basalts. The main groups may easily be observed on Miyashiro's (1973) diagram (fig. 5). The massive basalts and dolerites closely ressemble present-day oceanic basalts (Terry, 1979). The analogies have also been emphasized by trace element and REE data (Allègre et al., 1973 ; Montigny et al., 1973 ; Montigny, 1975).

The harzburgites, websterite and gabbros (Tabl. 9) exhibit a positive Eu anomaly. Depleted LREE similar to MORB are shown by dolerite and albitophyres. The latter are enriched in K and Rb, the former having a K/Rb ratio comparable to oceanic tholeiites . The whole sequence, however, shows high $^{87}Sr/^{86}Sr$ ratios ranging from 0.7046 to 0.7085 (Montigny, 1975). However, there are some differences between the albitophyres and spilites and MORB, but Montigny et al. (1973) considered that these rocks were tholeiites altered by hot pore-water circulation. The geochemical discrepancies between some Pindos lavas and dykes, especially the low-TiO_2 dykes, lead to the conclusion that these ophiolites may have been formed in an island-arc-marginal basin system (Beccaluva et al., 1979 ; Capedri et al., 1980) or in an island-arc -off-shore oceanic crust environment (Beccaluva et al., 1979).

7. Othrys

According to Smith et al. (1975), the rocks of the Othrys mountain can be divided into three main groups :

- the Othrys group comprising Paleozoic carbonates, Mesozoic dolerites, pillow lavas

cherts and carbonates (Maliac zone),
- the Mirna ophiolitic complex, in this paper called the Othrys ophiolitic complex,
- the Dinai group of upper Cretaceous conglomerates and limestones followed by a lower Tertiary flysch.

The ophiolites are overlain by Triassic-Jurassic volcano-sedimentary series of the Othrys group. In the younger sequence (Dinai group) rudist fragments (Renz, 1927 ; Aubouin, 1957) near the base gave an Cenomanian to Coniacian age (Stornay and Fleury, in Hynes et al., 1972) and locally an Aptian age (Ferrière, 1974). The Othrys group has been subdivided into three sequences (Smith et al., 1975 ; 1979) or five stratigraphic series (Ferrière, 1972) :
- shallow water platform carbonates resting unconformable on Paleozoic metamorphics : the Strimbes limestone sequence in Central and Eastern Othrys = neritic limestones of Flambouri = (perhaps) the Trilophon series in Western Othrys. Two distinct structural units have been defined by Ferrière (1974) : the Flambouri neritic limestones resting on the Pelagonian and the Chatala series, transitional between the Pelagonian and the Maliac.
- The Poulia sequence or Garmeni Rachi series, a complete series from Permian to Malm, is characterized by a clastic limestone-radiolarian cherts formation. A volcanosedimentary unit is composed mainly of pillow lavas and radiolarites similar to the Agrilia formation (see later) overlie by cherty limestone of Norian age. Fragmental limestones, turbidites and sedimentary gravity flows (the Meterizia formation) of upper Triassic - Jurassic age have been considered to have formed in a small submarine fan (Price, 1976).
- The pelagic limestone Loggistion = Neokhorion sequence in Eastern and Central Othrys, and the series of Katsichorachi and Moschokarya in Western Othrys (Courtin, 1979), comprise a volcano sedimentary unit : the Argilia formation. The lavas of continental-rift type (Hynes, 1974) consist of picritic basalts, diabase, keratophyric tuffs, pillow lavas and pillow breccias, and trachytes (Tabl. 10). A wehrlite related to the picrite and a dacite, respectively ages of 251 ± 5 m.y. and 231 ± 4 m.y. by the $^{40}Ar/^{39}Ar$ method (Roddick et al., 1979). K-Ar ages of the Triassic volcanics of the Argilia formation (a rhyolite and a spilite-dolerite) gave an age of 181 ± 6 and 156 ± 6 m.y. (Hynes et al., 1972). However, conodonts above the dolerites and pillow lavas of the Katsichorachi series (Courtin, 1979) gave a Carnian-Norian age. In the Moschokarya series the pillow lavas are overlain by radiolarian cherts followed by limestones with conodonts of Norian age (Courtin, 1979). Ophiolitic pillow lavas tectonically overlie the whole Loggistion series. The Profitis Ilias series (Ferrière, 1977) is characterized by a Triassic Amonitico rosso facies with Conodonts overlying volcanics and by their peculiar structural position : they are tectonically overlain by ophiolitic ultramafics above their amphibolites soles. The amphibolites (Tabl.7) near Lamia gave an age of about 177 ± 4 m.y. (Spray and Roddick,

	1	2	3	4	5	6	7	8
SiO_2	47.23	53.83	49.11	79.21	61.93	68.80	46.19	49.56
TiO_2	0.33	1.48	1.19	0.14	0.37	0.36	1.40	1.41
Al_2O_3	5.19	16.18	13.71	11.38	17.21	14.27	17.41	14.45
Fe_2O_3	}	}	}	}	}	} 4.59	} 9.50	} 9.78
FeO	} 8.13	} 6.58	} 7.40	} 0.93	} 3.19	}	}	}
MnO	0.17	0.21	0.15	0.02	0.09	0.05	0.18	0.08
MgO	33.63	7.48	7.62	0.21	0.41	0.20	3.59	6.79
CaO	4.41	8.96	15.76	0.53	4.24	0.61	8.95	5.59
Na_2O	0.28	3.93	3.87	5.26	5.74	5.60	4.64	3.83
K_2O	0.28	0.86	0.09	2.26	6.84	4.64	0.97	2.71
P_2O_5	0.35	0.58	1.02	0.07	0.30	-	-	4.26
L.O.I.	-	-	-	-	-	0.72	3.50	-
CO_2	-	-	-	-	-	-	3.72	0.85
	100.00	100.09	99.92	100.01	100.32	99.84	100.05	99.31

Table 10 : Chemical composition of the Othrys Triassic volcanics

1. picrite
2. diabase
3. pillowlava
4. tuff
5. trachyte
6. trachyte
7. pillow lava
8. massive basaltic flow

1 to 5 after Hynes (1974)
6 to 8 after Courtin (1979)

1980). The garnet-pyroxene amphibolite (high amphibolite to granulite facies) reached conditions of ≈ 860° C and 7 Kbar (Spray and Roddick, 1980). To the Western part of Othrys similar series to the Profitis Ilias one have been described by Courtin (1979 : the series of Tourla and Grammeni. Near Tourla Conodonts and Ostracods limestones of Ladinian to Carnian age fill the interstices of the pillow lavas (Kozur in Courtin, 1979). The ostracods need sea-water temperatures of 10 to 14° C and may exist in a deep neritic environment over 200 m (Colin in Courtin, 1979) perhaps 500 m, but less than the CCD (Carbonate compensation depth). The detritic formations of the Grammeni and the Profitis Ilias series may correspond to facies developed near a stable continental margin initiated through shear movements (Boillot, 1979) on the Sperchios transform fault (Aubouin and Dercourt, 1975 ; Ferrière, 1977 ; Courtin, 1979). Petrographically these lavas (Tabl. 10) are transitional basalts and peralkaline trachytes (Courtin, 1979). Pyroxenes evolve from augite through salite and aegyrine-augite to aegyrine and are associated with plagioclase (labradorite to albite) and K-feldspar and arfvedsonite (Lapierre et al., submitted). Therefore, since the Trias a shoal with neritic facies existed at the Southern border of the Maliac zone (Ferrière, 1974). This zone extended to the North in the Koziakas area (Ferrière, 1974 ; Celet et al., 1978), with effusive rocks are interbedded with limestones of Liassic age (Scandone and Radoicic, 1974).

The ophiolites

Smith et al. (1975) distinguish several outcrops :

- the Agoriani plagioclase-lherzolite with metamorphic fabric and gabbroic schlieren ;
- the kedhron harzburgitic tectonites with interlayered dunites overlying the Kournovon dolerites, pillow lavas and sediments ;
- the Mega Isoma magmatic lherzolites with intersertal plagioclase with wehrlite and gabbro and lying also tectonically on the Kournovon dolerites. Vertically banded anorthosite dykes and irregular masses of gabbro occur in the serpentinized peridotites ;
- the Dereli homogeneous gabbro with sporadic pegmatitic patches. Olivine gabbros display cumulus textures. Rare diorite has also been reported ;
- the Kournovon dolerites are mainly fine grained dykes cutting the gabbroic cumulates. Rhyolite flows, pillow lavas and basalt flows are associated within this unit.
- The Sipetorrema pillow lavas unit consists of pillow lavas, lava flows (containing Ol and Cpx phenocrysts) associated with thin beds of red siltstone and cherts.

The main ophiolite descriptions are taken from Courtin's thesis (1979).
The harzburgites with tectonic fabric (porphyroblastic texture) crop out near Fournos Kaitsa. They are composed of Ol (65-80 %), Opx (10-30 %), Cpx (2-5 %), Sp (1-1.5 %) and contain layers of pyroxenite. The lherzolites from Pirgarki Thrapsimion and from

Mount Koukos exhibit a similar porphyroblastic texture, however, more clinopyroxene rich (5-10 %). The dunites near Katachloron (90-95 % Ol, 1-3 % Sp, 0-5 % Opx) show a granular banded texture. Foliated plagioclase-peridotites crop out within the massif of Mega Isoma, Karapa, Mount Koukos, Platia Rachi Agoriani and Kouramos. Two main facies may be observed :

- layered plagioclase-lherzolites (60-70 % Ol, Sp 1-2 %, Pl (An_{95}) 5-7 %, Opx 12-20 %, Cpx 7-10 %) with porphyroblastic texture ;
- homogeneous foliated Pl-lherzolites containing up to 20 % of Pl (An_{72}).

The peridotites suffered syntectonic recrystallization with crystals elongated in a parallel direction to the foliation plane followed by a post tectonic recrystallisation (late annealing) with triple point boundaries. In general the foliation is parallel to the layering, small folds are linked to the development of the foliation. Within the feldspar free peridotites, the gabbroic dykes are undeformed, whereas they are often deformed within the Pl-peridotites.

The entire spectrum ranging from Pl-lherzolite towards residual harzburgite has been interpreted to be the residua left after one or more partial melting events (Menzies, 1973 ; 1977 ; Menzies and Allen, 1974).

The transition between harzburgite and Pl-lherzolites is gradual with an increase in Pl and Cpx (Menzies and Allen, 1974). Rare wehrlite (Ol 70-80 %, Cpx 15-20 %, Opx 0-5 %, Pl 0-2 %, Sp 1 %) has been observed within the Pl-peridotites always near dykes with diffuse edges. Pl-dunites (An_{79}) exhibit an unclear tectonic fabric. Chromitites have been exploited near Domokos where gabbroic dykes cut serpentinized dunite and chromitites (Maratos, 1964).

Near Kedros and Trilophon other chrome occurrences have been described (Marinos et al., 1962 in Courtin, 1979). Gabbroic dykes are scarce within foliated peridotites and more abundant within the plagioclase-peridotites.

Some troctolites with ortho and adcumulate textures (for the cumulus texture terminology refer to Wager and Brown, 1968) as well as Ol-gabbros, gabbros and gabbro-norites sometimes exhibit deformation phenomenon.

The troctolites are cumulates without clear layering but with ratio contacts (Jackson, 1974). This unit ranges from Pl-dunites (Ol 60-70 %, Pl 30-40 %, opaque 1-2 %) with orthocumulate texture to troctolites (Ol 30-35 %, Pl 65-70 % (An_{76}), opaque (1 %) with adcumulate texture. When clinopyroxene is present (up to 7 %), it is poikilitic and the troctolite shows a heteradcumulate texture.

The olivine-gabbro (Pl An_{75-65} 47 %, Cpx 30 %, Ol 22 %, opaque 1 %) presents heteradcumulate texture, the plagioclase is often prehnitized.

The gabbros (Pl An_{50-55} 58 %, Cpx 40 %, opaque 2 %) are sometimes homogeneous and sometimes layered. Heteradcumulus clinopyroxene is often transformed into tremolite. The diallage gabbro (euphotide of the alpine geologists see vol. 1) show weak layering. Near Domokos (Petra Mokos and Souffi) leucocratic (Q-albite) apophyses

	1	2	3	4	5	6
SiO_2	40.32	39.92	43.63	42.30	45.30	49.69
TiO_2	-	0.01	0.11	0.11	0.10	0.36
Al_2O_3	0.53	0.11	1.85	4.60	16.28	17.33
Fe_2O_3	-	4.87	2.48	2.20	1.58	1.81
FeO	8.91	4.84	6.18	6.40	3.59	3.07
MnO	0.15	0.18	0.16	0.21	0.25	0.13
MgO	49.01	49.00	42.58	37.23	17.41	10.91
CaO	0.88	0.19	2.15	6.14	12.77	14.15
Na_2O	0.16	0.05	0.14	0.20	1.20	2.75
K_2O	0.05	0.01	0.02	0.01	0.06	0.04
Cr_2O_3	-	0.45	0.42	0.33	-	-
L.O.I.	-	-	-	-	2.80	(2.80)
	100.01	99.63	99.72	99.73	101.34	100.24

Table 11 : Chemical composition of the Othrys ophiolite : plutonic rocks

1. harzburgite
2. dunite, mean
3. lherzolite, mean
4. plagioclase-rich lherzolite
5. gabbro
6. gabbro

1 after Menzies (1973)
2 including 0.38 of NiO
3 including 0.28 of NiO
4 including 0.19 of NiO after Menzies and Allen (1974)
5 and 6 after Menzies (1976)

have been observed by Maratos (1964). They are mainly diorite, quartz diorite and tonalite (trondhjemite) from Asproklissia (Courtin, 1979) Q 45 %, feldspath 45 %, green hornblende, biotite, apatite and epidotes 10 %).

The effusive rocks are composed of :

- lava dykes with Pl (An_{60}) phyric and aphyric textures,
- picritic flows with Cpx-Ol-phyric and aphyric textures,
- pillow lavas sometimes Pl-phyric. They are always spilitized.

More recent considerations indicate that normal and inverse series of ophiolites may be primary (Courtin, 1979 ; Celet et al., 1980). The ophiolites of Domokos with harzburgite, chromitite, low-Ti dykes and some diorite may have been formed along a transcurrent fault bordering a bassin with a normal ophiolitic sequence consisting from base to top of harzburgite and Pl-lherzolite, olivine-gabbro, dolerite, and lavas with high Ti (Courtin and Ohnenstetter, in preparation).

Selected chemical analyses of the whole ophiolitic sequence are given in Tabl. 11 and 12. As in the Pindos the effusive rocks show affinity with ocean-floor tholeiites, and some dykes, especially within the Domokos area*, show low TiO_2 (fig. 6), high SiO_2 contents comparable to island-arc volcanics (Bickle and Nisbet, 1972). The existence of boninites has been signalized (Cameron et al., 1980).

The REE patterns indicate that these tholeiites could be generated by 10-30 % melting of the tectonic lherzolite (Menzies, 1976 a). These lherzolites have chondritic REE abundances and may represent undepleted mantle material - refractory residue - liquid extract system (Menzies, 1976 b).

The $Sr^{87}/^{86}Sr$ ratios range from 0.703 to 0.707 in basalts and gabbros, and from 0.703 to 0.725 in peridotites (Menzies, 1976 b). These high values may be caused by interactions with fluids during serpentinization ; separated minerals (diopsides) from the same peridotites gave a $^{87}Sr/^{86}Sr$ ratio of 0.7037, similar to those of ocean ridge tholeiites (Menzies and Murthy, 1976).

The Opx varies from En_{92-89}, Wo_{0-2}, Fs_{8-9} in the ultramafics to En_{88-85}, Wo, Fs_{15-12} in the mafics. Clinopyroxene evolves from En_{54-48}, Wo_{50-38}, Fs_{2-10} to En_{51-53}, Wo_{65-36}, Fs_{2-20} and olivine is close to Fo_{82-78} (Menzies, 1973). Harzburgites contain Cr-spinel and lherzolites more Al-rich spinel. The Cr-Al variation in the spinels may result from interaction with interstitial aluminous liquid (Menzies, 1975).

The ophiolitic complex is affected by early-tectonic events of uppermost Jurassic - lower Cretaceous age (Ferrière, 1972 ; 1974 ; Smith et al., 1975 ; Smith and Woodcock, 1976 a and b). They have been transported to the NE (Ferrière and Vergely, 1976 ; Smith, 1977 ; Smith et al., 1979). The activation of the margin is caused by the initiation of a subduction zone to the West of the Pelagonian zone (Smith and Woodcock, 1976 b ; Smith et al., 1979).

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* New chemical data kindly provided by Lapierre (Lapierre and Ferrière, in preparation) confirm the observation of Bickle and Nisbet (1972).

	1	2	3	4	5	6	7
SiO_2	48.60	49.20	54.06	45.56	54.12	57.94	58.43
TiO_2	1.39	1.32	1.04	1.60	0.49	0.33	0.44
Al_2O_3	15.30	14.72	13.51	14.34	15.19	14.60	13.37
Fe_2O_3	8.67	11.05	6.84	13.02	10.71	9.68	4.79
MnO	0.13	0.16	n.d.	n.d.	n.d.	0.14	n.d.
MgO	6.24	9.19	6.92	9.91	4.01	4.45	4.82
CaO	9.32	6.28	9.03	9.03	6.67	8.46	4.48
Na_2O	3.56	4.13	3.21	1.81	3.97	1.95	3.78
K_2O	0.90	0.30	0.25	0.27	0.06	0.35	0.15
P_2O_5	0.07	0.14	n.d.	n.d.	n.d.	n.d.	n.d.
CO_2	0.24	0.07	n.d.	n.d.	n.d.	0.15	n.d.
L.O.I.	4.34	3.19	4.91	3.55	4.39	1.53	3.89
	98.76	99.75	99.77	99.09	99.62	99.58	99.15

Table 12 : Chemical composition of the Othrys ophiolite : effusive and hypabyssal members

1. basaltic pillow lava
2. basaltic pillow lava
3. basaltic pillow lava
4. basaltic dyke from Trilophon
5. basaltic andesite dyke from Domokos
6. andesitic dyke from Domokos
7. andesitic dyke from Domokos

Kindly supplied by J. Ferrière to be published in Ferrière Thesis (1981) in preparation.

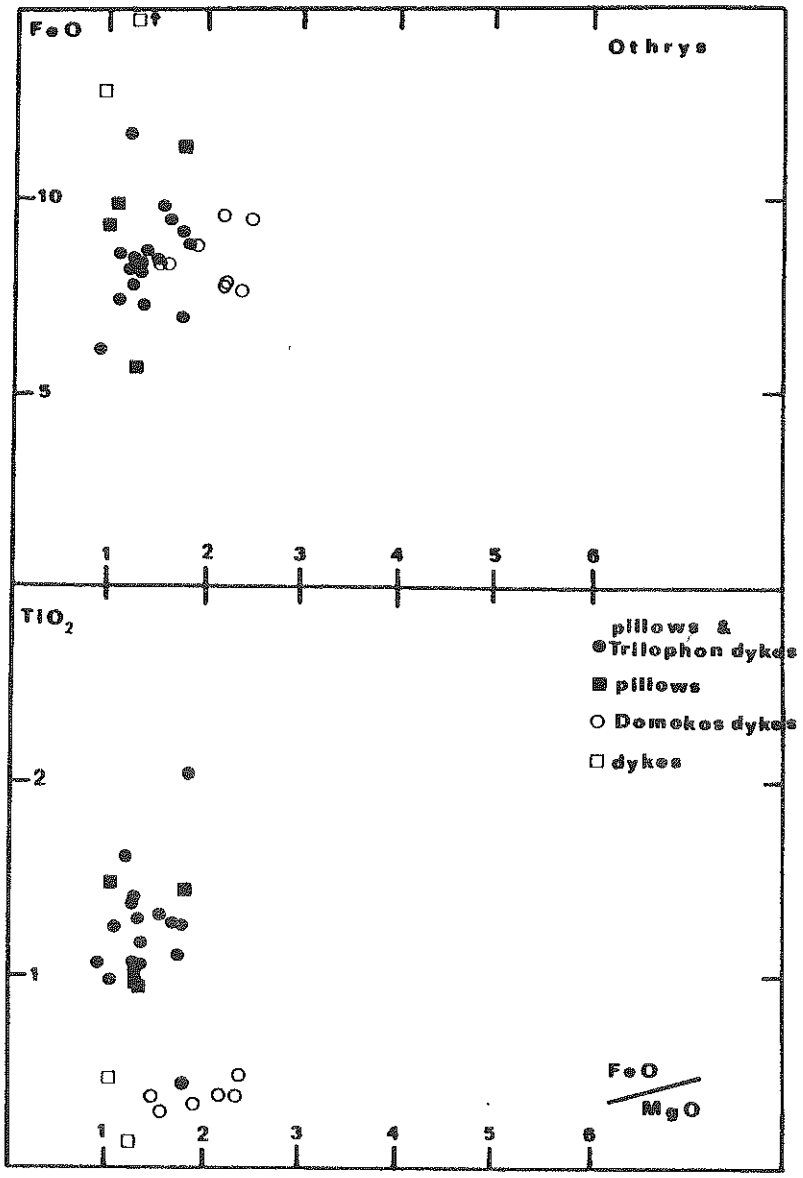


Fig. 6 - FeO total or TiO_2 versus FeO total/MgO for effusive and hypabyssal rocks of the Othrys ophiolites.

8. Iti

In the Iti mountains S of the Sperchios river, Wigniolle (1977) has described different series coming from numerous zones. Ophiolitic rocks (Tabl. 13) considered as olistoliths by Celet (1976) crop out within a mêlange occurring as klippes in the external domain. It has been tectonized after deposition during the upper Jurassic, and overlain tectonically by limestones of Malm age. Celet (1976) considered that this "mêlange" was formed near a continental slope. The ophiolitic nappe was destroyed and the debris transported to the front of the nappe in movement (Celet et al., 1977).

The ophiolitic rocks (Wigniolle, 1977) are mainly serpentinized peridotites (harzburgites and lherzolites), gabbros and olivine gabbros, dolerites, diabases, granophyres and lavas ; most of them are spilitic (fig. 7) and highly altered and are associated with micaschists and granitic blocks. Near Pyra occurs an ophiolitic massif of 15 km^2 formed by tectonites (harzburgites and dunites), cumulate gabbros, isotropic gabbros, hornblende and apatite-rich diorites. Doleritic dykes occur within the harzburgites and gabbros. Some gabbros have been metamorphosed into the amphibolite facies. The volcanics are agglomerates and massive aphanitic lava flows ; no pillows have been encountered. Lavas occur on top of the serpentinites as well as on top of the gabbros and the amphibolites.

9. Euboea

North and Central Euboea or the "non-metamorphic Pelagonian zone" is mainly composed from bottom to top of neo-Paleozoic formations, limestones and metabasic volcanites of lower to middle Triassic age, neritic limestones and dolostones of upper Triassic to Jurassic age, ophiolites followed by transgressive limestones of upper Cretaceous age (Katsikatsos, 1977). In Central Euboea, more or less serpentinized peridotites constituting the unique member of the ophiolitic trilogy overlie a volcano-sedimentary sequence.

Harzburgite (Tabl. 14) up to 600 m thick is composed of 65 % Ol, 26 % Opx, 7 % Cpx and 2 % brown chromian spinel. A few lherzolite cumulates have also been observed (Parrot and Guernet, 1972). Cumulus textures are exceptional, in this case poikilitic pyroxene encloses olivine. Euhedral spinel may be concentrated in thin layers. These cumulus peridotites underwent recrystallization in presence of a liquid and were later affected by a strong plastic deformation (Capedri, 1974). Olivine varies from $Fo_{91.5}$ to $Fo_{90.5}$; enstatite is relatively uniform $En_{90.7-90.1}$; and the clinopyroxene is diopsidic $En_{48.7}\ Wo_{48.5}\ Fs_{2.8}$; spinel has Cr_2O_3 contents varying from 51.98 to 37.79 % and Al_2O_3 from 16.53 to 31.28 (Capedri, 1976). The spinel compositional variation is related to the change from a porphyroclastic to a microgranular texture. This latter may be in equilibrium with a tholeiitic liquid.

	1	2	3	4	5	6	7	8
SiO_2	48.99	50.53	49.08	75.43	50.54	59.22	50.55	52.70
TiO_2	0.20	1.93	1.19	0.35	0.56	0.70	1.24	2.59
Al_2O_3	21.17	14.07	13.14	13.94	16.47	18.79	14.82	15.24
Fe_2O_3 FeO	6.43	11.78	9.33	2.28	5.78	3.94	10.28	10.71
MnO	0.59	0.66	0.20	0.03	0.12	0.08	0.19	0.12
MgO	6.25	6.49	6.13	0.10	8.60	1.49	7.57	4.32
CaO	3.37	5.12	12.71	< 0.01	12.98	6.35	9.65	5.50
Na_2O	4.52	4.67	2.44	6.24	2.43	6.53	2.76	5.76
K_2O	0.74	0.38	0.01	0.39	0.12	0.38	0.84	0.02
L.O.I.	7.85	4.36	5.19	-	1.07	2.14	1.65	2.60
	100.11	99.99	99.42	98.76	98.67	99.62	99.55	99.56

Table 13 : Chemical analyses of the Iti ophiolite

1. gabbro, Iti
2. dolerite, Iti
3. basalt, Iti
4. granophyre, Iti
5. gabbro, Pyrna
6. diorite, Pyrna
7. dolerite, Pyrna
8. basalt, Pyrna

1 to 8 after Wignolle (1976)

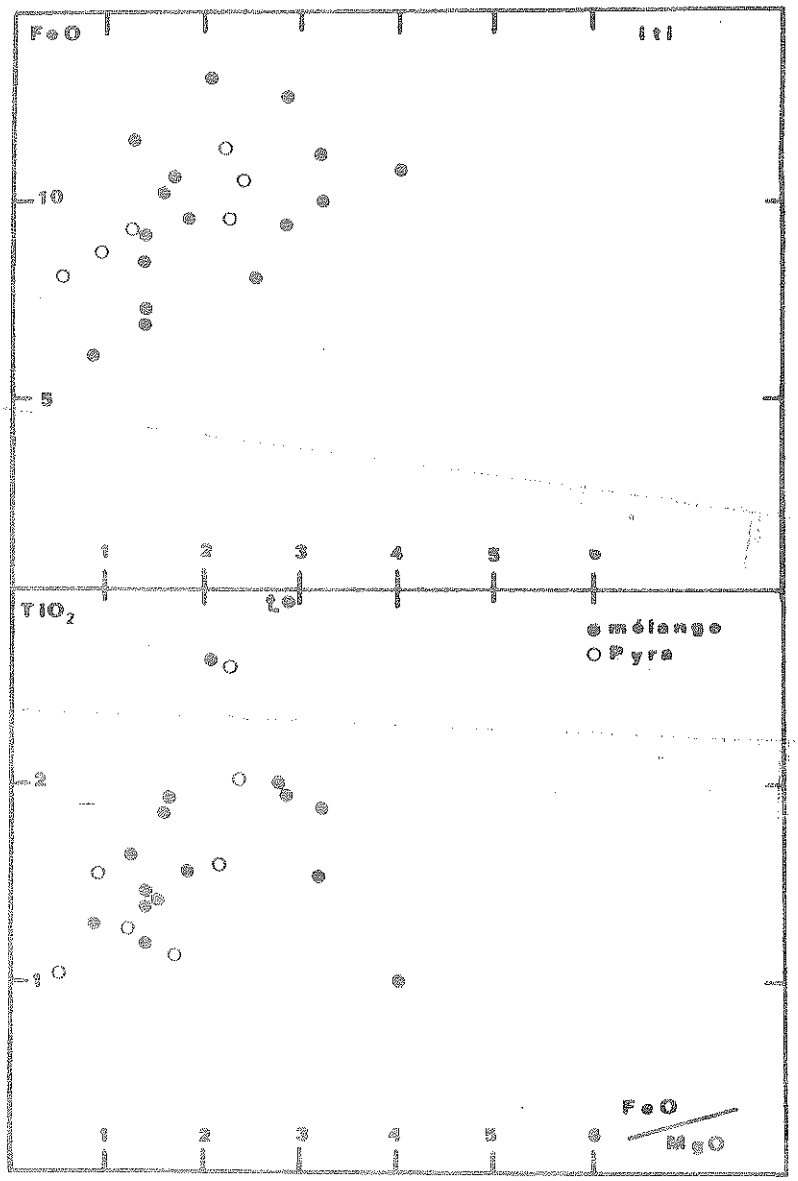


Fig. 7 - FeO total or TiO_2 versus FeO total/Mg for effusive and hypabyssal rocks of the Iti ophiolites.

	1	2	3	4
SiO_2	48.55	47.32	44.62	44.76
TiO_2	1.80	2.04	1.95	0.05
Al_2O_3	15.19	14.77	13.56	1.65
Fe_2O_3	-	-	6.04	0.87
FeO	10.61	9.49	4.26	6.86
MnO	0.18	0.17	0.16	0.13
MgO	5.51	7.18	4.38	43.29
CaO	8.91	8.02	14.19	1.71
Na_2O	4.80	3.40	3.10	0.02
K_2O	0.20	1.60	2.30	0.01
P_2O_5	0.16	0.15	0.08	-
Cr_2O_3	-	-	-	0.31
NiO	-	-	-	0.25
H_2O^+	2.87	4.43	4.07	0.57
H_2O^-	0.60	0.94	0.57	0.14
CO_2	-	-	1.20	
	99.38	99.51	100.48	100.62

Table 14 : Chemical composition of the volcanosedimentary sequence and of the harzburgite of Euboea

1. basalt from the Triassic volcanosedimentary sequence
2. basalt from the Triassic volcanosedimentary sequence
3. basalt from the Triassic volcanosedimentary sequence
4. harzburgite

1 to 3 after Parrot and Guernet (1972)
4 after Capedri (1976)

Re-equilibration occured at 900° C and 20-30 Kbar in a subduction zone environment (Capedri, 1976). Diabase and gabbro dykes cutting the serpentinized peridotites exhibit a rodingitic mineralogy : hydrogrossular, prehnite, clinozoisite, zoisite, xonotlite, vesuvianite, calcite, tremolite and chlorites (Bertrand et al., 1980).

These peridotites were eroded before the Cenomanian transgression (Guernet and Parrot, 1972) and exhibit a lateritic alteration with Ni deposits (Mossoulos, 1964). Near Troupi metamorphic rocks occur between the peridotites and the volcano-sedimentary sequence.

The amphibolites below the peridotites contain hornblende + oligoclase + epidote + sphene and derived from metabasites. They have been dated through the $^{40}Ar/^{39}Ar$ method (Spray and Roddick, 1980) and give an age of 180 ± 5 m.y. (modified spectrum).

The volcano-sedimentary sequence (about 100 m) is composed of radiolarian cherts and volcanic rocks mainly pillow lavas (Tabl. 14), rich in augite (up to 31 %) and plagioclase (48 %) ranging from An_{55} to An_{45}. Olivine is rare at the base of the lava flows and more abundant towards the top. These rocks are affected by a zeolite facies metamorphism (Parrot and Guernet, 1972). The diabase-chert formation has an upper age of late Jurassic to early Cretaceous, probably Berriasian to Valanginian (Baumgartner and Bernouilli, 1976). The radiolarian cherts overlie conformably a thick sequence of shallow-water limestones of Kimmeridgian age (Bassoulet and Guernet, 1970).

10. The ophiolites of the blueschist units of Ossa, Southern Euboea and Cyclades

Several units comprising basic and ultrabasic rocks metamorphosed in the blueschist facies, have been described in Greece. One can quote :

- the Ossa unit, thrust upon the Olympos limestones, dolomites and flysch of external origin (Godfriaux, 1965), and tectonically overlain by a sheet of internal Pelagonian origin (Derycke and Godfriaux, 1976 and 1977) ;
- the Southern Euboea unit (Styra - Ochi) thrust upon the Almyropotamos internal formations (marbles with Triassic and Upper Cretaceous fauna), and overlain by the Central Euboea sheet showing intermediate characteristics (Katsikatsos et al., 1976 a and b, Katsikatsos, 1977) ;
- the Cycladic blueschist unit thrust upon a pelitic unit intruded by granites and metamorphosed in the greenschist facies in the Syros island (Bonneau et al., 1980 a et b).

The Styra - Ochi unit, in Southern Euboea has been folded and metamorphosed during the Upper Jurassic-Lower Cretaceous, the Lower-Middle Eocene and the Upper Eocene-Lower Oligocene (Bavay et al., 1980). One can rezognize Eocene folding and metamorphic phases in Ossa and in the Cyclades.

On Syros island (Bonneau et al., 1980 a et b) and in Southern Euboea (Andronopoulos, 1962 ; Bavay and Romain-Bavay, 1980), the ophiolites appear as an enormous olistostrome in a metaflysh : they are made up of serpentinites, metagabbros, and metabasalts with glaucophane, sodic pyroxene, garnet, phyllite, and chloritoid ; Fe-Ti gabbros with omphacite, riebeckite and deerite have been described in Syros. Similar parageneses have been described at Milos (Kornprobst et al. (1979). A comparison of these various metamorphic windows (amongst which it is advisable to quote the High Pelion one, which probably is not ophiolitic does not lead to the establishment of clear correlations (Katsikatsos et al., 1976 ; Celet et al., 1978). It appears, for example, that these units have followed various tectono-metamorphic evolutions : in Southern Euboea, ante Middle Cretaceous phases of folding and metamorphism have been recognized, whereas on Syros island or in Ossa, the first phase is probably of Upper Cretaceous - Eocene age.

These windows give evidence of the importance of tangential tectonics in Greece, particularly during the Eocene, in which the thrusting of internal or intermediate units on external formations, contributes to throw into confusion the data concerning the Jurassic paleogeography and the origin of the ophiolites.

GENESIS OF THE GREEK OPHIOLITES RELATED TO THEIR ENVIRONMENT

1. The associated volcanism

Within the Othrys-Pindos-Euboea zone the Triassic volcanism has a transitional to alkalic character and corresponds to the rifting and early development of the ophiolitic trough (Hynes et al., 1972 ; Rocci et al., 1980 ; Celet et al., in preparation) and is similar to the volcanism described in the Internal Dinarides (Ciric and Karamata, 1960). Both differ from the volcanism of the same age from the Vardoussia (Bérard-Bergery, 1980) which shows a calcalkaline tendency similar to that found in the external zones of the Dinarides (Ciric and Karamata, 1960 ; Bébien et al., 1978) as well as the continental tholeiitic volcanism of Atalanti (Verriez, 1980).

In the Paikon massif between the ophiolites of Almopias and Guevgueli, low-K tholeiitic basalts, andesites and rhyolites of Jurassic age occur. In the Peonias area in a more internal position, amongst formations probably of Liassic age outcrop rhyolites, calcalkaline andesites and tholeiites with abyssal affinity (Metalikon, Aspro Vissi). Low-K tholeiitic basalts, calcalkaline andesites and rhyolites can be related to island-arc volcanic activity in a continental area, whereas tholeiites with abyssal affinity are linked to the magmatism of the Guevgueli massif (Bébien, 1976).

To summarize, it appears that before the ophiolite formation (during the

Trias), the external zone was characterized by calcalkaline volcanism and the intermediate zone by transitional to alkaline volcanism where ophiolites appear late (Euboea, Othrys, Pindos). In the "Vardar" zone during the Jurassic, the Paikon plays the role of an island-arc separating two oceanic basins (Almopias and Peonias).

2. The amphibolite soles

Amphibolites have been described beneath most ophiolites in the Hellenides as well as in the Dinarides (Karamata, 1974). They derived from the metamorphism of a volcano-sedimentary sequence which ranges from greenschist to amphibolite (epidote - amphibolite) facies and in some places (e.g. Othrys) to granulite facies up to 860° C, 7 Kbar (Spray and Roddick, 1980). The metamorphic rocks result from the emplacement of ophiolites on the subjacent volcano-sedimentary sequences (Woodcock and Robertson, 1977), sometimes during subduction processes (Whitechurch and Parrot, 1978 ; Nicolas and Le Pichon, 1980).

At the present day metamorphic soles beneath ultramafics exist on several island-arc systems : the Yap trench system (Shiraki, 1971 ; Hawkins and Batiza, 1977) and in the Southwest Pacific (Parrot and Dugas, 1980).

If the age, nature and condition of metamorphism of the amphibolites are considered together, their origin may give some indications about the origin of ophiolites.

a) Somes amphibolites may be derived from lavas, as pillow structures and tuffite are sometimes preserved (Pichon and Brunn, 1977) . They possess a transitional to alkaline character plotting in the Within-plate basalt discriminant fields (Spray and Roddick, 1980), as do the Dinaric amphibolites (Pamic, 1977).

It is highly improbable that they derive from the overlying ophiolitic basic rocks, most often strongly improverished in alkali and incompatible elements. For example, a Vourinos amphibolite has TiO_2 = 2.49 % and Zr = 130 ppm, whereas ophiolitic lavas have TiO_2 ranging from 0.19 to 1.01 % and Zr from 16 to 58 ppm. Therefore, the original material must be looked for in another volcanic sequence such as the Triassic one from the Intermediate zone characterized by transitional to alkalic nature. Associated metasediments-marbles, micaschists, metacherts, quartz-phyllites and quartzites derived either from cherts or sandstones (Karamata and Lovric, 1978) - look like the Triassic sedimentary sequences associated with volcanism as in the Maliac zone.

b) The age of the underlying amphibolites lies in the range between 170-180 m.y. (Spray and Roddick, 1980 ; Thuizat et al., submitted) similar to the ages of

Yugoslavian amphibolites (Lanphere et al., 1975 ; Karamata and Lovric, 1978 ; Okrush et al., 1978). They are about 30 m.y. older than the age of emplacement on the continental margin during the Upper Jurassic (JE_1) or Lower Cretaceous JE_2 phase). Therefore, the age of amphibolites appears to be close to the time of ophiolite formation.

c) The high temperature indicated by mineral assemblages in Greek and Yugoslavian amphibolites is incompatible with a simple frictional heat of thrusting and thus hot ophiolitic slabs are needed to obtain this high temperature (Pamic et al., 1973 ; Karamata, 1974 ; 1979 ; Pamic, 1977 ; Woodcock and Robertson, 1977 ; Malpas, 1979). This dynamothermal event (Williams and Smith, 1973) in the Hellenides and Dinarides has been attributed to obduction from narrow oceanic areas (e.g. Red Sea type) or marginal basin with high heat flow (Karamata, 1979). Consequently, the formation of amphibolites occurs near the site of ophiolite generation.

The above considerations indicate that the amphibolites derive from a dynamothermal metamorphism of a volcano-sedimentary sequence resembling the Triassic one and occurring nearly at the same time and place as the ophiolite formation, that is before thrusting onto the continental slope. In a narrow oceanic basin diapiric emplacement of hot mantle material may cause the deformation and recrystallization of the previous Triassic volcano-sedimentary sequences giving the amphibolite sole.

3. Ophiolites and tectonic phases

Important data concerning the history of the Greek ophiolite complexes can be obtained from a tectonic and microtectonic study.

The JE_1 (ante-Tithonic) and/or JE_2 (lower Cretaceous) phases, used in the definition of the internal and intermediate areas, have been recognized in and under most of the ophiolites.

Folds with a West or Southwest vergence, caused by the JE_1 phase, have been observed under the Guevgueli ophiolites, in the mélanges of the Almopias zone, in the serpentinites of the Vermion massif and in the lower part of the Vourinos ophiolites. A greenschist-facies metamorphism is often associated to these deformations.

One can recognize structures with an East vergence, produced by the JE_2 phase, in the Othrys formations, in Euboea, and in most of the ophiolites affected by the JE_1 phase (Vourinos, Vermion, Almopias), with the exception of the most eastern one (Guevgueli). A metamorphism may be linked to these deformations ; it is of the "greenschist" or sometimes of the "blueschist" type (Styra-Ochi unit in Southern Euboea, Milos island).

These early phases are also responsible for the formation of units with ophiolitic fragments in Pindos, Othrys, Iti, Southern Euboea and Syros.

The recognition of these tectonic phases JE_1 and JE_2 permits us to consider the hypothesis of a multiple origin both in space and time for the Greek ophiolites

- the ophiolites affected by the JE_1 phase would have an eastern (Vardar) origin,
- those affected only by the JE_2 phase would have a western origin (Ferrière and Vergely, 1976 ; Vergely, 1977).

4. The sedimentary cover of the ophiolites

The deposition of the transgressive sediments on the Greek ophiolitic complexes took place chiefly during :

- the Upper Jurassic and Eocretaceous,
- the Middle and Upper Cretaceous.

Sediments of Upper Jurassic-Eocretaceous age, transgressive on ophiolites, have been described in the Pelagonian Massif (Vourinos, Zygosti and Vermion ophiolites). The Jurassic radiolarian cherts, in close association with ophiolitic lavas, are similar to the cherts and diatomaceous oozes from the Japan sea and thus a large distance from a continent and very deep seas are not essential to the development of this siliceous sedimentation (Steinberg et al., 1977). Above the radiolarian cherts, the sedimentary formations are characterized by the presence of limestones with siliceous layers, giving evidence of the existence of sheltered shallows ; they are more detritic to the East (Vermion), where the deposits follow an important phase of erosion, than to the West (Vourinos, Zygosti) (Pichon, 1976, 1977).

Towards the East, basic plutonic and volcanic formations linked to the Guevgueli complex are transgressively overlain by reef limestones of Upper Kimmeridgian and Portlandian age associated with detritic layers (Mercier, 1966). These formations of Upper Jurassic and Eocretaceous age are not affected by the JE_1 tectonic phase, whereas the underlying ophiolites and sediments are affected by this phase.

In many other regions, the first observable formations on ophiolites are of Middle and Upper Cretaceous age. In the Almopias zone, Middle and Upper Cretaceous deposits overlie various kinds of volcanic and sedimentary formations, including ophiolites ; the transgression began during the Upper Aptian - Albian and went on up to the Upper Maestrichtian. In the Western border of the Almopias zone, thick deposits can be found near the Pelagonian ridge ; Cenomanian pelagic sediments characterized its central part (Mercier, 1966).

The ophiolites and the Triassico-Jurassic volcanic and sedimentary formations of Othrys are also discordantly covered by Cretaceous deposits (Dinai group) made up of detritic rocks and limestones overlain by a flysch. The transgression began with the Aptian (Ferrière, 1974).

In the same way, the peridotites of Central Euboea are transgressively covered by formations made up of Upper Cretaceous limestones and of a flysch.

These facts are to be linked with the existence, in Greece, of the JE_1 (Upper Jurassic) and JE_2 (Lower Cretaceous) tectonic phases. The emplacement of formations of Upper Jurassic and Eocretaceous age in the Pelagonian Massif or on the Guevgueli ophiolites follows an emersion caused by the JE_1 phase. In Euboea or in Othrys, only the JE_2 phase has been recognized : the transgressive sediments on the ophiolites are of Middle and Upper Cretaceous age. However, the lack of Upper Jurassic and Eocretaceous sediments does not prove the non-existence of the JE_1 phase : in the Almopias zone, this phase is clearly expressed.

5. The ophiolites

Taken as a whole, the ophiolitic massifs from Greece present all the various geochemical and petrographical characteristics defined in ophiolites (Rocci et al., 1975 ; Church and Riccio, 1977 ; Pamic, 1977). The different members may be successively considered. Both harzburgite and lherzolite exist in some massifs (generally large and in an external position), whereas some others like the Vourinos possess mainly harzburgite. Some massifs with subvertical foliation (Vourinos) may be related to diapiric emplacement either in an oceanic-ridge (Moores and Vine, 1971 ; Ross et al., 1980) or in a transverse-fault system (Ohnenstetter, 1980), whereas some others with original platy foliation (Othrys) may be formed farther from the ridge axis (Courtin and Ohnenstetter, in preparation).

The order of crystallization of minerals from some cumulate assemblages indicate an early precipitation of plagioclase in the Othrys and Pindos ophiolites, whereas in the Vourinos the two pyroxenes or at least clinopyroxene appear before plagioclase (see fig. 8 and 9). Moreover, the Fe/Mg and Na/Ca ratios exhibit a different behaviour during these low pressure fractionations. Large variation of Fe/Mg coupled with elevated An content in plagioclase (except in diorites) is observed in Vourinos, whereas decrease of An content occurring with slight increase of Fe/Mg ratio is shown by Othrys and Pindos cumulates. The large range of composition and evolution of liquids may be considered in the TiO_2 - FeO/MgO diagram (fig. 10). Three fields have been defined high Ti (HT) (corresponding to the anisotitaniferous field of Bébien (1980), low Ti (LT) and very low Ti (VLT) corresponding respectively to the evolutionary domain of abyssal tholeiites, island-arc tholeiites and boninites. It appears that each ophiolitic massif possesses group

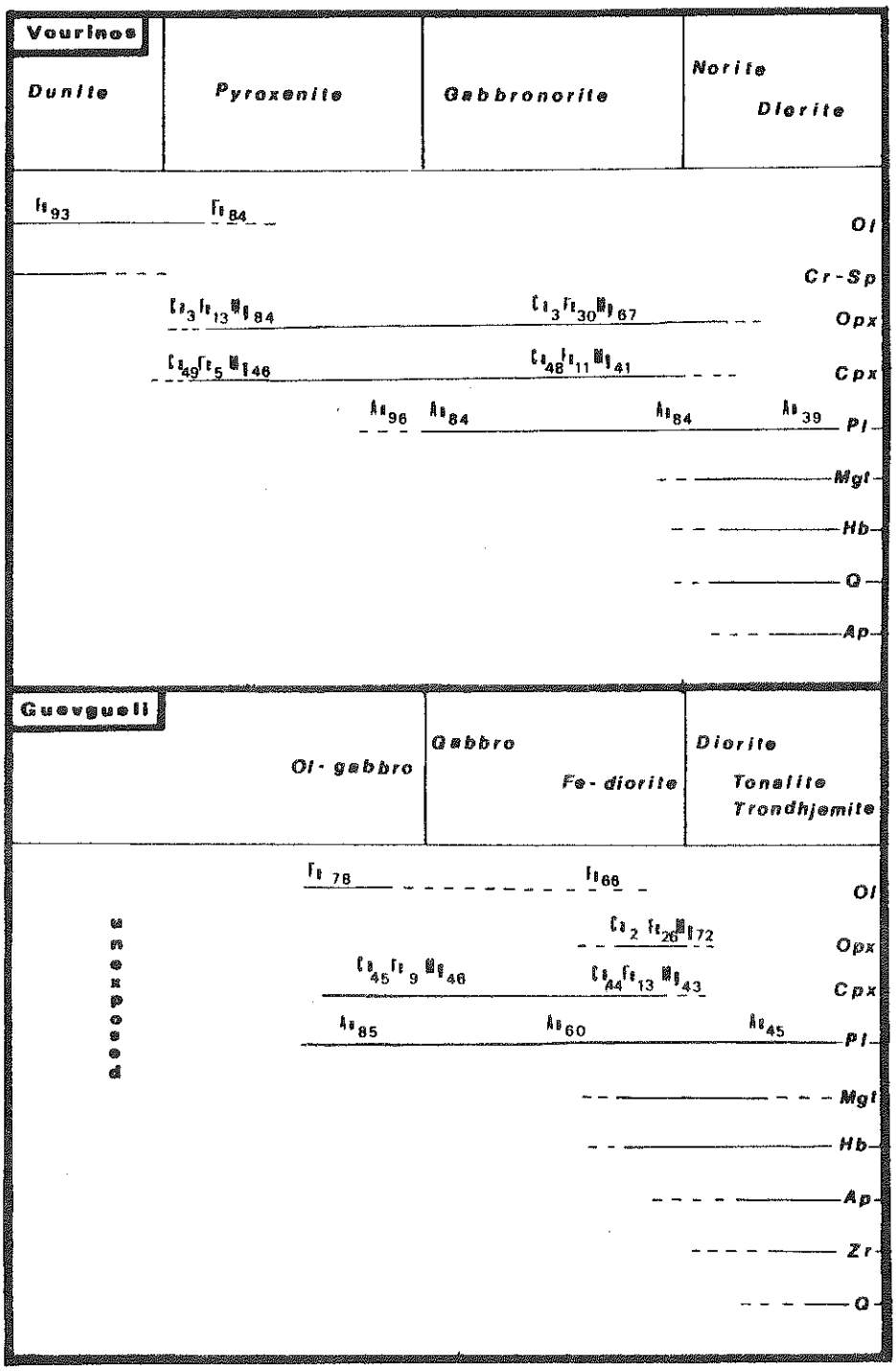
of lavas falling in two juxtaposed fields or sometimes in the three fields, as for Pindos. As in some other Mediterranean ophiolites, later dykes more often falling in the VLT group show a picritic character (Beccaluva et al., 1979). In a more general way, in Greece, HT basalts occur in massives having at least cumulate with early plagioclase and lherzolitic peridotites. On the contrary VLT lavas are associated with harzburgites and pyroxene-rich lower cumulate The spatial association of peridotites, cumulates and lavas with pronouncedly different character as well as the evolution of magmatism with time exclude the presence of a true ocean of Atlantic, Indian or Pacific types. It is better explained in a system of marginal basin and island-arc environment (Beccaluva et al., 1979 ; Upadhyay and Neale, 1979). Morever, in a single marginal basin like Othrys, geochemical and petrological variation are assigned to different ophiolitic segments each characterized by a peculiar internal geometry compatible respectively with an extensional regime or with strike-slip systems (Courtin and Ohnenstetter, 1981). Therefore, it appears that in Greece some ophiolites may have appeared in extensional ridge (part of Othrys and Pindos), whereas some others from Guevgueli, Vourinos and the Domokos area in Othrys may be related to transcurrent faulting (Bébien and Gagny, 1978 ; Ohnenstetter, 1980 ; Courtin and Ohnenstetter, 1981).

6. Association of ophiolites and granites

The association of ophiolites and granites found in the Guevgueli complex is rather unusual and is worth discussing more precisely. This kind of association is not restricted to this complex : it characterizes a 200 km long belt stretching northerly up to Skopje in Yugoslavia (Bébien, 1977).

The ophiolitic nature of the Guevgueli complex has often been discussed during these last years (Jones et al., 1977 ; Brunn et al., 1977). In fact, its petrological and structural characteristics, though they do not correspond exactly to the traditional picture of an ophiolitic complex, are not exceptional. For example, ophiolites of Guevgueli and of Southern Chili (Dalziel et al., 1974 ; Brunn et al., 1978) present many common features : scarcity of outcrops of ultramafic rocks, presence amongst gabbros of granitic rocks, non tectonic relationships between the two complexes and volcano-sedimentary formations ; on the Hokkaido island in Japan, the Hidaka metamorphic belt (Hashimoto, 1975 ; Miyashiro, 1977) is also made up of ultrabasic and gabbroic rocks showing ophiolitic characteristics, associated with granites and migmatites. On the other hand, a structural organisation comparable to that of the Guevgueli complex can be observed in the Canyon Mountain ophiolites in Oregon (Avé Lallement, 1975 ; Thayer, 1978).

The presence of ophiolites in previously mobilized continental crust can account for their association with granites and migmatites (Bébien, 1981).


Vourinos
Dunite
Pyroxenite
Gabbronorite
Norite
Diorite
Fo93
Fo84
Ol
Cr-Sp
Ca3Fe13Mg84
Ca3Fe30Mg67
Opx
Ca49Fe5Mg46
Ca48Fe11Mg41
Cpx
An96
An84
An84
An39
Pl
Mgt
Hb
Q
Ap
Guevgueli
Ol-gabbro
Gabbro
Fe-diorite
Diorite
Tonalite
Trondhjemite
Fo78
Fo66
Ol
Ca2Fe26Mg72
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Ap
Zr
Q
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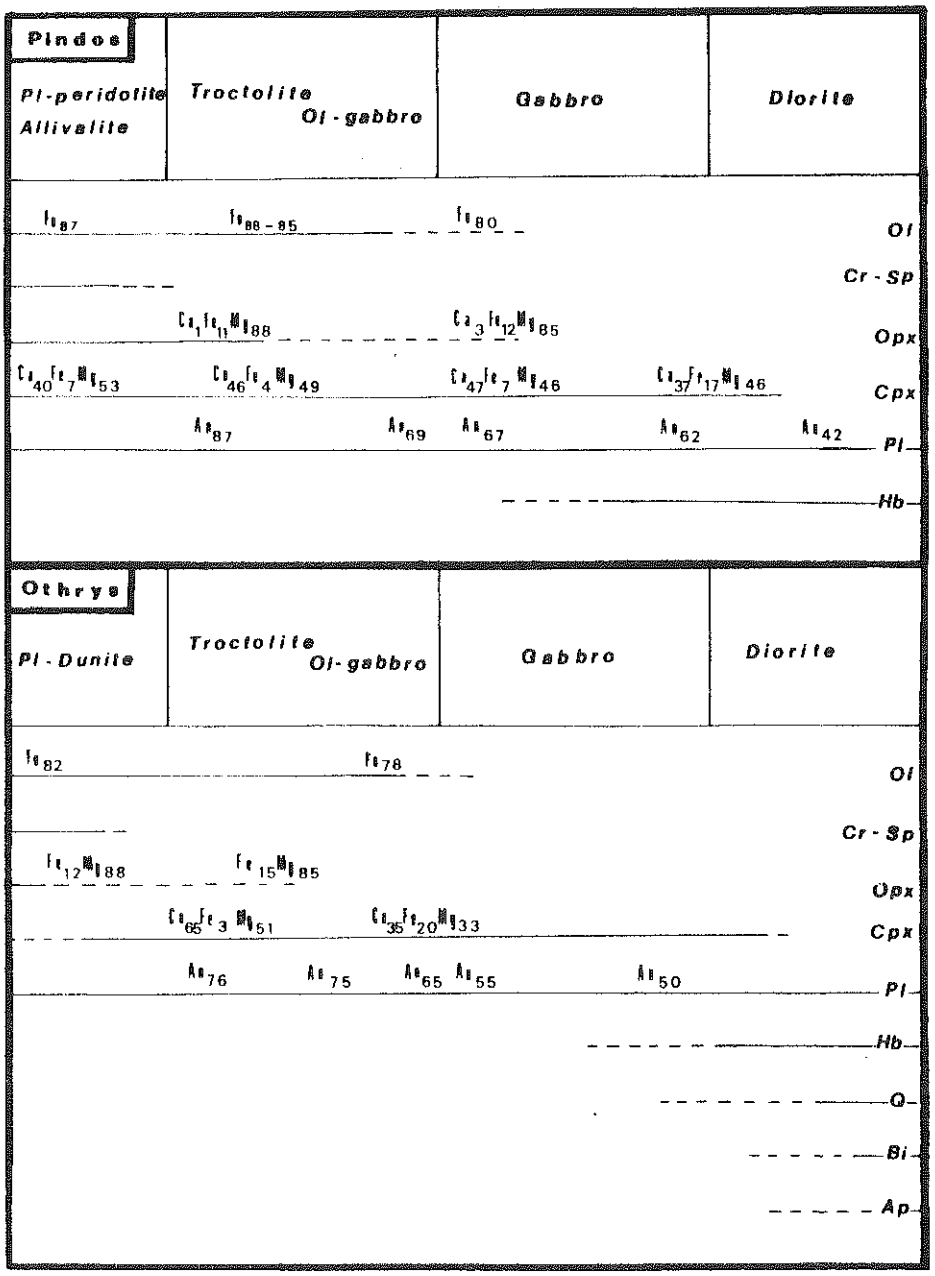


Fig. 8 and 9 - Order of crystallization and mineralogical evolution of the main ophiolitic complexes. For the Vourinos two evolutionary trends exist (see Beccaluva et al., submitted). Keys to the symbols: Ol = olivine; Cr-Sp = chromian spinel; Opx = orthopyroxene; Cpx = clinopyroxene; Pl = plagioclase; Mgt = magnetite (sometimes titaniferous); Hb = hornblende; Ap = apatite; Zr = zircon; Q = quartz; Bi = biotite.

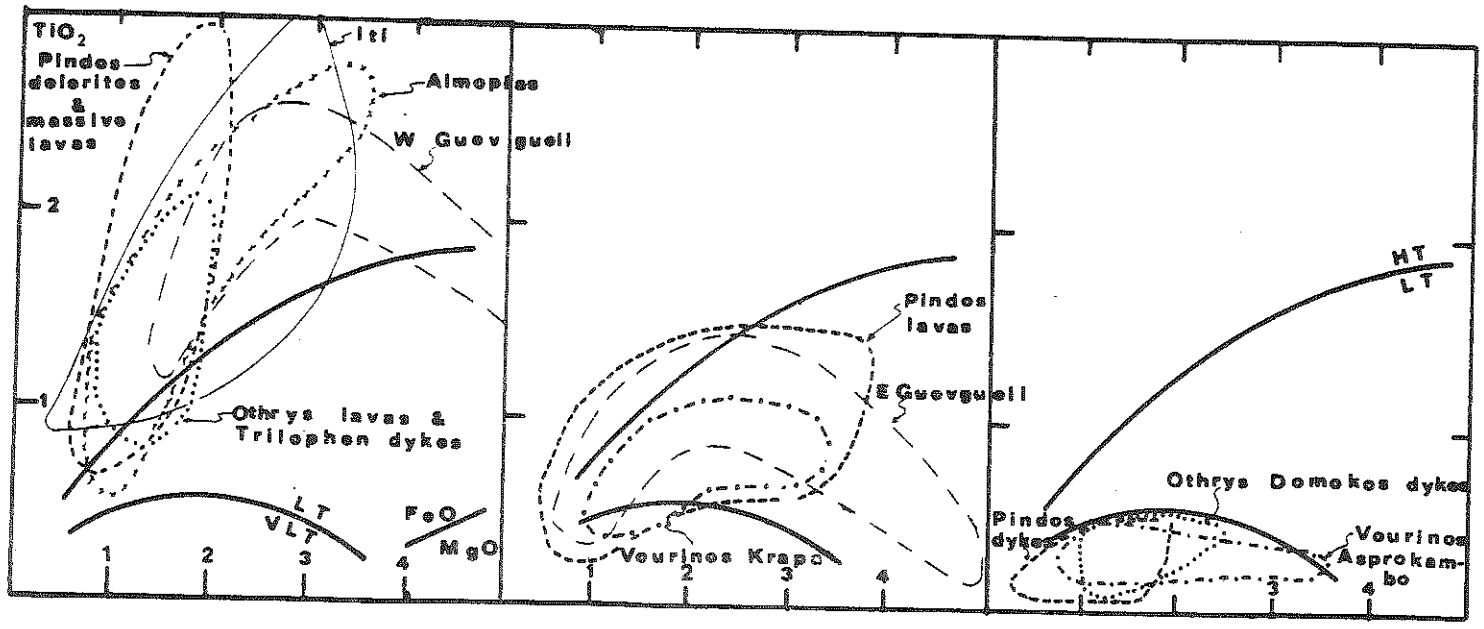


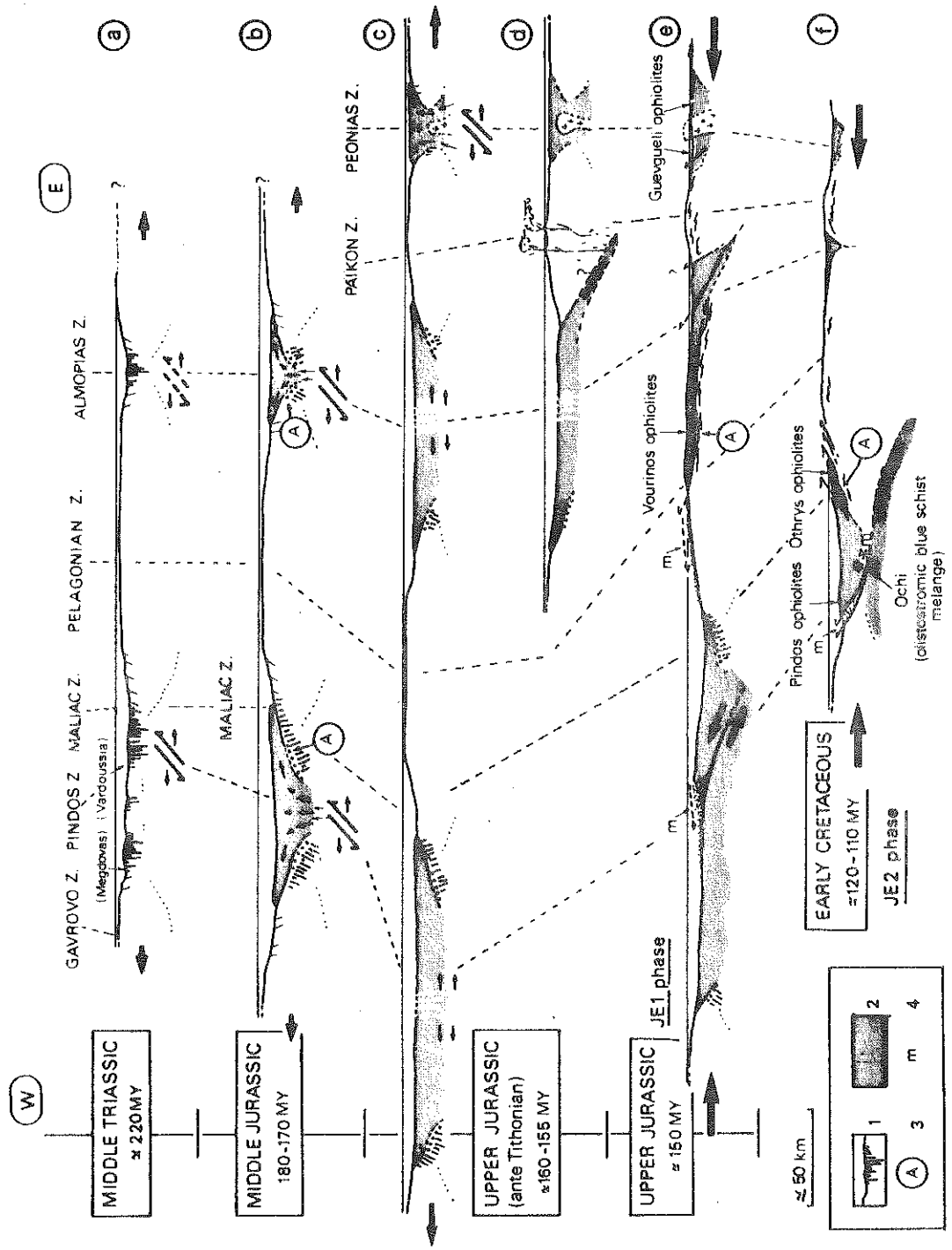
Fig. 10 - Comparison of the different evolutionary trends from the effusive and hypabyssal rocks of the Greek ophiolites in the TiO_2 versus FeO total/MgO diagram (Miyashiro, 1973). The high TiO_2 (HT) and low TiO_2 (LT) boundaries are taken from Bébien (1980) and correspond respectively to his anisotitaniferous and isotitaniferous field. The very low TiO_2 (VLT) field characterizes lavas with affinities to boninites and marianites (see Beccaluva et al., submitted).

The hypothesis of the formation of the Guevgueli ophiolitic complex during the Upper Jurassic in an active continental margin environment suggested by the volcanic rocks of the Paikon Massif - low-K tholeiites, rhyolites accounts well for its geological setting and its petrographic characteristics (Mercier et al., 1975). These ophiolites would correspond to a small basin formed in mobilized continental crust by simultaneous spreading and shearing, as in many presently existing marginal (Lawver and Hawkins, 1978).

GEOTECTONIC EVOLUTION OF THE HELLENIDES RELATED TO THE OPHIOLITE FORMATION

In the Hellenides where several ophiolitic sutures exist, the previous plate tectonic reconstructions have always considered the spatial association of an ocean or an ocean with island arc and marginal basins. The present study points out the lack of evidence for remnants of a true ocean (Tethys ?). On the contrary the geological facts are compatible with an active margin system (Blanchet, 1977). From West to East numerous troughs characterized by rapid subsidence may be represented, for example : Pindos, Maliac, Almopias, and Peonias (see fig. 11) which may be interrupted longitudinally. During their evolution some of them will always keep a sialic base (sialic ocean of Celal Sengör and Monod, 1980) upon which a calc-alkaline pre-ophiolitic volcanism may occur. In some others during the Jurassic the ophiolites are emplaced after a Triassic transitional to alkalic volcanic events. The structural and geochemical features of the ophiolites argue against the evolution of a Mid Oceanic Ridge type but are better explained by diapirism and magmatism related to limited extensional events or to transcurrent movements. The strike-slip system may have been generated in an active margin developed since the Trias during the approach of the edges of Apulian and European continental plates related to punching and rigid plastic squashing (Tapponnier, 1979).

The present-day Aegean model with a sialic marginal basin created during intra-continental collision (Angelier, 1978 ; Le Pichon, 1979 ; Le Pichon and Angelier, 1979 ; Mercier, 1979 ; Mercier et al., 1979) with calcalkaline volcanism (Ninkovich and Hays, 1972 ; Vilminot and Robert, 1974) may best explain the evolution of the Hellenides during the Triassic. The formation of ophiolites may correspond to a more advanced stage of rifting. This model involves narrow and shorted-live oceanic basins (ensialic ophiolites of Ohnenstetter, 1980) locally interrupted and close to their margins. During the diapiric uprise of mantle material, the pre-existing Triassic volcano-sedimentary sequences are metamorphosed and folded (amphibolite sole). In such a context, post-ophiolitic sediments may be pelagic, neritic, recifal and detritic when deposited sometimes after erosional features. Important lineaments help the sliding and/or thrusting of the ophiolitic nappe onto the continental margin. The tectonic events occur earlier (Upper Jurassic) in the Eastern part and migrate


W
E
a
b
c
d
e
f
GAVROVO Z.
PINDOS Z.
(Megdovas) (Vardoussia)
MALIAC Z.
PELAGONIAN Z.
ALMOPIAS Z.
MALIAC Z.
PAIKON Z.
PEONIAS Z.
Guevgueli ophiolites
Vourinos ophiolites
Othrys ophiolites
Pindos ophiolites
Ochi
(distal pelagic blue schist melange)
MIDDLE TRIASSIC
≈ 220 MY
MIDDLE JURASSIC
180-170 MY
UPPER JURASSIC
(ante Tithonian)
≈160-155 MY
UPPER JURASSIC
≈ 150 MY
JE1 phase
EARLY CRETACEOUS
≈120-110 MY
JE2 phase
≈ 50 km
1
2
3
4
m
A

towards the Southwest during the Lower Cretaceous. Therefore, the origin and emplacement of ophiolites are closely related to the development of an orogen (Brunn, 1957 ; 1980).

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Fig. 11 - Evolution of the Hellenides from the Middle Trias to the Early Cretaceous.

a - Middle Trias. Volcanic phenomena are associated with the formation of several troughs. The volcanism is known in the Megdovas, the Vardoussia, and in the Maliac zone. Its existence is only inferred in the Almopias zone : a Triassic almopian volcanism is perhaps represented, as amphibolite soles, beneath the Vourinos ophiolites. There is no evidence for the existence, at that time, of a true ocean in the Hellenic area. However, the occurrence of calc-alkaline volcanism may point out an active margin environment.

b - Middle Jurassic. Emplacement of ophiolites in the Maliac and the Almopias zone. This emplacement is probably related to transcurrent movements (double arrows) ; it induces the transformation of Triassic volcanics into amphibolite soles.

c, d - Upper Jurassic. Emplacement, related to transcurrent movements, of ophiolites (Guevgueli) and granites in the Peonias zone. This magmatism, and the volcanism in the Paikon massif, are probably linked to subduction phenomena, perhaps in the Almopias zone (d).

e - Upper Jurassic. JE_1 phase : ophiolites of the internal areas (Almopias, Peonias) are tectonically thrust toward the West. This phase also provokes the formation of volcano-detritic mélanges (m) in the internal and intermediate areas.

f - Early Cretaceous. JE_2 phase : ophiolites of the intermediate areas (Maliac) are tectonically thrusted towards the East. The previously formed mélanges may be affected by a metamorphism in the blueschist facies (Ochi).

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OPHIOLITES AND RELATED FORMATIONS IN THE CENTRAL PART OF THE SABZEVAR RANGE (IRAN) AND POSSIBLE GEOTECTONIC RECONSTRUCTIONS.

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ABSTRACT.

In the central part of the Sabzevar range, structural and petrological studies have been undertaken on ophiolites, upper Cretaceous volcanosedimentary sequences and metamorphic rocks of the Soltanabad area. During ophiolite formation, several intrusions have been recognized considering dyke relationships and crystallization order in cumulates. Internal consistencies, dating of magmatic amphiboles (Coniacian) and paleomagnetic data are in favour of the genesis of ophiolites in limited extensional areas. These ophiolites are overlain by an upper Cretaceous to Paleocene volcanosedimentary sequence developed from juxtaposed ridges and troughs filled with island arc calc-alkalic and alkalic volcanics and sediments. The metamorphic rocks from the Soltanabad area are of intermediate type between greenschist and blueschist facies. The basic metamorphic rocks may belong to pre-ophiolitic transitional to alkalic series, whereas gneisses, dioritic to tonalitic in composition, are similar to the ophiolitic suite. The culmination of metamorphism occurred during pre Lutetian tectonic phases which led to the building of the Sabzevar range. A model for the structural evolution of the belt is proposed. It emphasizes the role of a strike-slip system during the opening of the ocean and the major part of its closure, the last events being more purely compressive.

INTRODUCTION

In the Mediterranean area, numerous ophiolite belts may be defined from Paleozoic to Mesozoic times. In Iran, according to Lensch and Davoudzadeh (1981), three belts have been recognized, one Paleozoic belt restricted to the Northern part and two Mesozoic belts : the first one running NW-SEcalledthe Zagros - Oman - Baluchestan belt (the outer axial ophiolite belt of Stöcklin, 1977), possibly developed from the Trias to the Turonian, whereas the second one borders the central East-Iran microplate. There, the ophiolites, mainly of upper Cretaceous age, are associated with mélanges (Takin, 1972).

Paleomagnetic data indicate that northern Iran was situated at the rim of Gondwanaland during the Devonian and Carboniferous (Wensink and Varekamp, 1980). Later this continent might have been collided with the Turan plate.

(Sabzehei, 1974 ; Davoudzadeh et al., 1981). The Paleozoic ophiolites may testify to this suture zone (Stampfli, 1978). It is generally assumed that rifting occurred during upper Trias (Sabzehei, 1974 ; Wensink and Varekamp, 1980) and Jurassic times (Haynes and Reynolds, 1980), especially around Central and East Iran, creating the Neotethys which separated Iran from Gondwanaland (Berberian and King, 1981). During Mesozoic times the Lut block might have begun an anticlockwise rotation (Davoudzadeh et al., 1980). In the Cretaceous, convergence of the Arabian plate towards Asia led

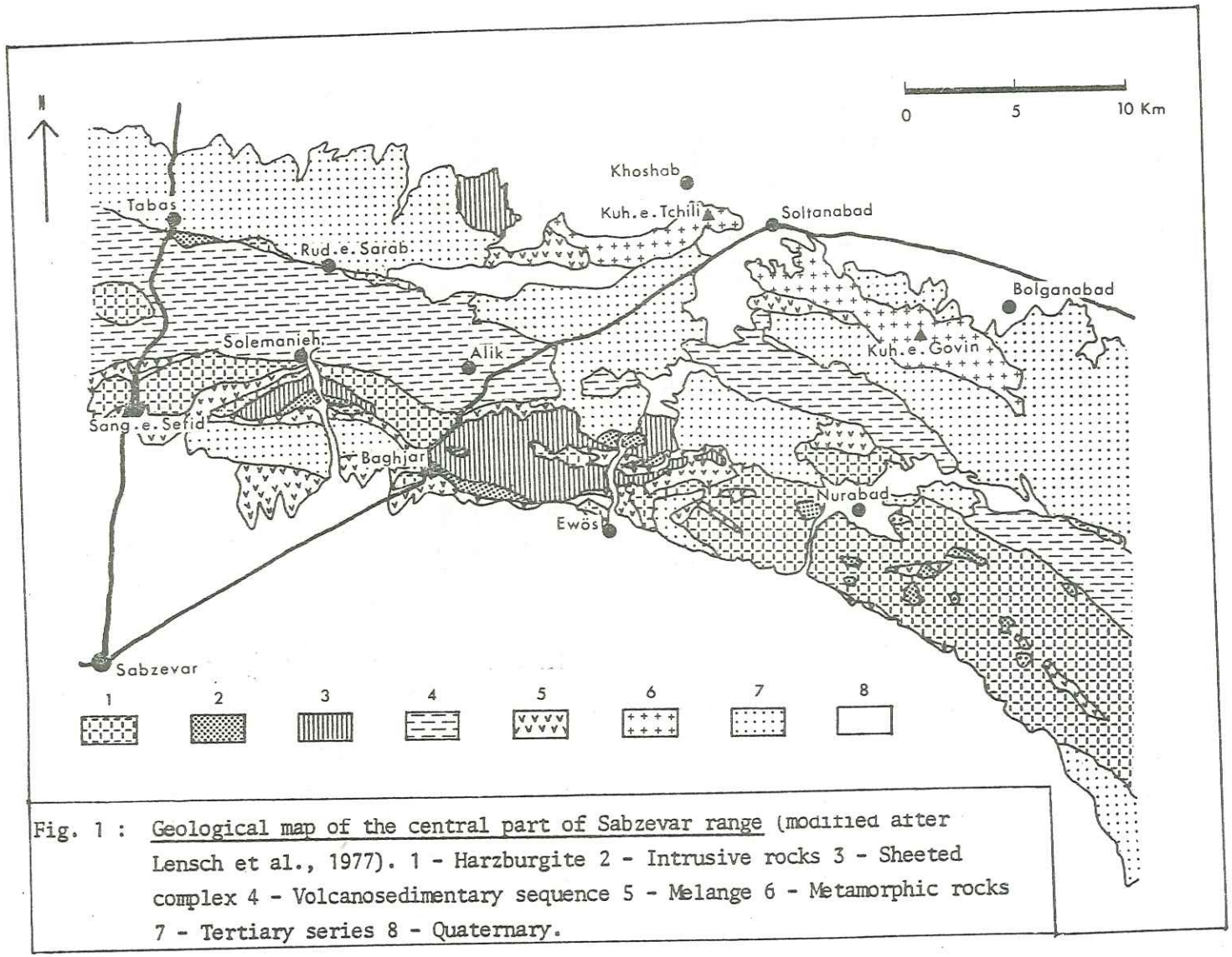

Fig. 1 : Geological map of the central part of Sabzevar range (modified after Lensch et al., 1977). 1 - Harzburgite 2 - Intrusive rocks 3 - Sheeted complex 4 - Volcanosedimentary sequence 5 - Melange 6 - Metamorphic rocks 7 - Tertiary series 8 - Quaternary.

to the disappearance of the ocean, the thrusting of ophiolite onto the northern African continental edge (Ricou et al., 1977) and the shearing of the north-eastern ophiolites around the Lut block (Lensch and Davoudzadeh, 1981). A summary of age data for Mesozoic times indicate generation of oceanic crust from Albian - lower Paleocene, ophiolitic emplacement restricted to upper Cretaceous - Paleocene and mélange formation during upper Cretaceous and possibly during Oligocene times (Delaloye and Desmons, 1980).

The ophiolites from the Sabzevar zone belong to the Mesozoic ophiolites surrounding the Central East Iran microplate (Lensch and Davoudzadeh, 1981) and were generated in small Cretaceous rifts (Stampfli, 1978). There, six main massifs occur between the Lut block, a microcontinent, and the Turan plate, showing more often incomplete series except for a large massif North of Sabzevar (Lensch, 1980). In this range, a Paleocene post-ophiolitic formation composed of flysch-type sediments and island-arc lavas has been outpoured on the rocks of the ophiolite assemblage created in an oceanic extension zone (Lensch et al., 1975 ; 1980). This ophiolitic assemblage includes an upper Cretaceous volcanosedimentary sequence. Detailed studies have been undertaken in the central part of the Sabzevar range. They are focus ed on three main topics :

- the nature and chronology of magmatic events during oceanic crust formation,
- the lithological, petrological and structural characters of the upper Cretaceous volcanosedimentary sequence,
- the structure and type of metamorphism in the Soltanabad area.

The present investigations on ophiolites and related formations were undertaken to determine their time relationships and put some constraints on geotectonic reconstructions.

I - THE MAIN PETROGRAPHICAL AND CHRONOLOGICAL FEATURES IN THE OPHIOLITES

Ophiolitic rocks are exposed in the southern part of the Sabzevar belt, which mainly trends E.W. Some tectonic slices composed of crushed serpentinites and subordinate gabbros may be also found towards the North against the upper Cretaceous volcanosedimentary sequence and Tertiary formations. Field exposures indicate a larger proportion of harzburgite (nearly 80 %) over cumulates (5 %) and effusive and hypabyssal rocks (15 %) mainly represented by a sheeted complex. Three areas of the Sabzevar range were covered with maps revealing major tectonic contacts within and around the ophiolites (Sadredini, 1974 ; Alavi-Tehrani, 1976 ; Vaziri-Tabar, 1976).

Like many other ophiolites, they have been considered as remnants of an oceanic crust created during the upper Cretaceous (Lensch et al., 1977). Later studies have emphasized the close ressemblances of ophiolitic magmatism with some

immature island-arc series (Lensch et al., 1979). Nevertheless, geological constraints lead to the assumption that ophiolites represent a sort of oceanic crust consumed at a converging plate margin since the Paleocene. This mechanism is thought to be responsible for the establishment of an island-arc sequence with calc-alkalic magmatism (Lensch et al., 1979).

Later on, in the ophiolite of the Sabzevar range, structural, petrological and chronological studies were engaged to point out some internal consistencies linked to their primary emplacement. In particular, the compositional evolution of liquids is precised by crystallization order in cumulates and the chronology of liquid emplacement by cross-cutting relationship in dykes.

A - Peridotites

Generally ophiolitic blocks are bounded with faults. Primary contacts, between some peridotites and gabbros and some upper gabbros and dykes have been preserved only at a few places. Those between serpentinized peridotites and gabbroic dykes and sills (which can be traced to the cumulitic pile), indicate that the folded layering in the peridotite was nearly subvertical at the time of gabbroic injection (Nurabad massif). Elsewhere in large peridotitic massifs, the variation in the strike and dip of the layering may be explained by large open folding during diapiric emplacement, later followed by tilting during more superficial thrusting and related folding. The harzburgites are moderatly deformed having mainly coarse-grained protogranular textures with slightly elongated minerals and locally porphyroclastic ones according to the classification of Mercier and Nicolas (1975).

B - Intrusive rocks

There is no evidence that cumulitic rocks once covered all the harzburgitic sole. Cumulates, limited to small, discontinuous massifs are indicative of numerous, isolated magmatic chambers intrusive at different levels in the upper crust. Such evidence results from the studies of four main cumulitic massifs near the villages of Nurabad, Ewös, Solemanieh and Tabas.

In the Nurabad massif, differentiation seems to have occurred in a small magma chamber situated within the peridotites. In the well exposed Solemanieh massif, the basal cumulates have crystallized in peridotites whereas the upper ones cooled near the surface under a carapace of sills and related lavas. Liquids escaping from the magma chamber have contributed to the formation of the sheeted dyke complex. The late plagiogranite plugs were intruded into the carapace. They may also be cut by basaltic dykes which testify to the persistence of a magmatic activity.

The Tabas massif may represent an equivalent of the upper portion of the Solemanieh magma chamber.

In the Ewös area, poorly layered plagioclase-wehrlite is in contact with fragment of sheeted dyke complex. As they present cumulitic textures and crystallization features different from the other well layered massifs, they may belong to a special intrusive event. In some other eastern Mediterranean ophiolites (Rocci et al., 1975), wehrlitic bodies possibly intrusive at a high level are considered to be the last magmatic activity (Smewing, 1980 ; Pallister and Hopson, 1981).

The different orders of crystallization deduced from the successive layered rocks inside and between the different massifs indicate the existence of multiple liquid injections of variable compositions (Church and Riccio, 1977 ; Smewing and Potts, 1976 ; Beccaluva et al., 1979 ; Smewing, 1981).
The usual crystallization order : Ol ± Sp, Cpx, Pl, Opx, Op, Amph is found in every massif. The position of Pl and Opx may alternate. The resulting assemblages are dunite, wehrlite, clinopyroxenite bearing more or less Pl, melagabbro, olivine-gabbro, gabbro norite, opaque-bearing gabbro norite, leucoferrodiorite, tonalite and plagiogranite. In the Tabas and Nurabad massifs, orthopyroxene is an early phase appearing before or at the same time as clinopyroxene so that limited outcrops of pyroxenites occur rather than wehrlite and clinopyroxenite, associated with orthopyroxene-rich gabbro and sometimes norite. Structural features have revealed that the shape of the magma chamber and the layering is mainly controlled by faults (Ohnenstetter, 1980). Magma arrives in radiating open fractures trending mainly about N 120° locally joined by contemporaneous horizontal cracks. In the upper part of the chamber, magma migrates through similar fractures zones subparallel to most of the dykes of the sheeted complexes (Ohnenstetter 1982, submitted).

C - The sheeted complex

The position of the poles of the dykes cutting the upper gabbros or forming the sheeted complex reveals numerous fracture zones which permit the ascension of the magma towards the surface (fig. 2). Such a complex situation prevents us from defining the position of the ridge axis. Dykes occupy radiating fractures possibly linked to conjugate faults or to extension zones varying from place to place. According to their composition and cross-cutting relationship several generations may be pointed out, especially in the Baghjar area. The earlier dykes with asymmetric borders are generally wide and composed of phyric basaltic andesite and andesite with phenocrysts of Pl and Cpx. Later narrower dykes with symmetrical chilled margins possess an intersertal flow texture. They are generally more evolved being richer in quartz and opaque minerals. Dacite representing the most fractionated liquid

follows these basalt and andesite. More basic and highly porphyritic dykes, mostly of basaltic andesite composition with abundant Cpx, Pl and subordinate pseudomorphs of olivine phenocrysts, cut all the previously described dykes. Those cutting the upper cumulates of Solemanieh bear some resemblance to these latest dykes. Some of these successive features are also found in the Troodos sheeted complex (Desmet et al., 1978). In the Sabzevar area, all the liquids analyzed are generally quartz-tholeiite (fig. 2).

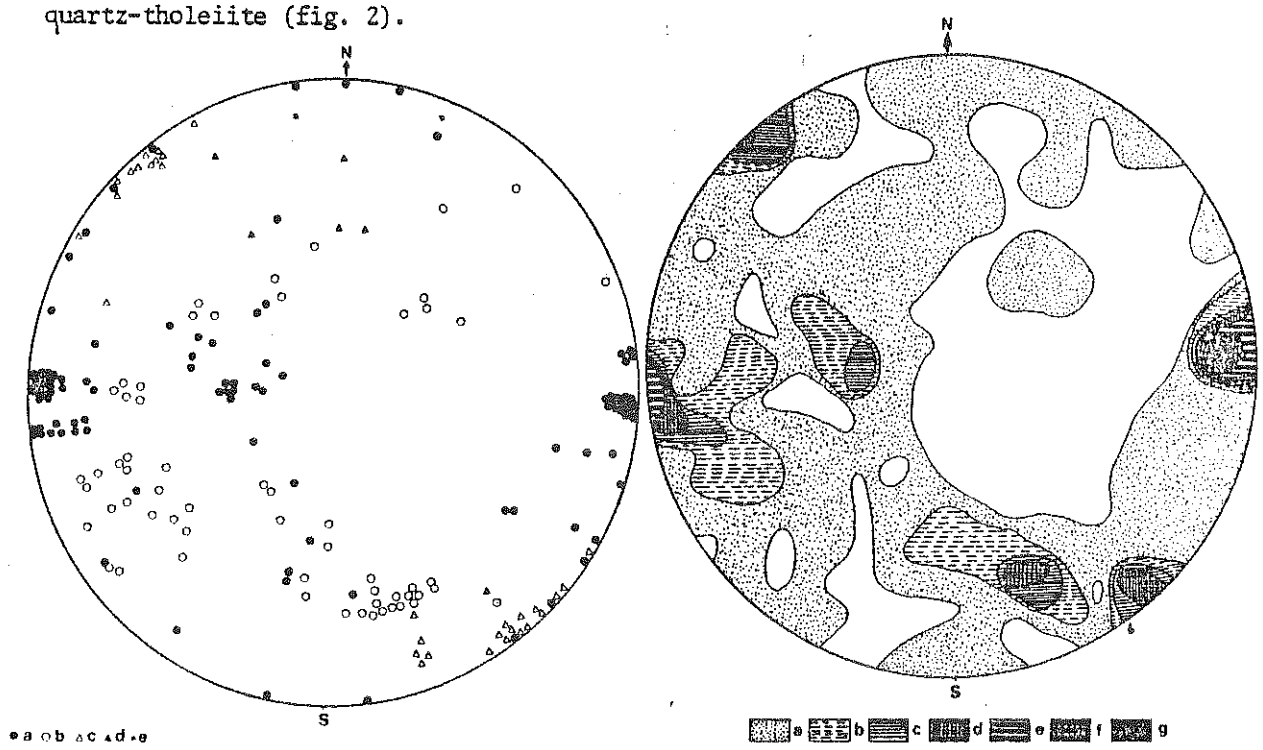

Fig. 2 : Diabase dykes from ophiolites. Left : Schmidt net. a - Dykes cutting gabbro b - Dykes in sheeted complex of Baghjar c - Dykes cutting diorite and plagiogranite Solemanieh d - Dykes from Eastern Nurabad gabbro. Right : Contoured diagram from the same points a - 0.6 % b - 1.2 % c - 1.6 % d - 2.4 % e - 3 % f - 3.6 % g - 4.2 %.

Petrological studies permit to state that two main magmatic events are recorded in the sheeted complex. The earlier dykes (basaltic andesite to dacite) may be related to the development of the main magmatic chamber having Ol-Sp-Cpx-Pl-Opx as crystallization order. Regular MgO and CaO decreases are compatible with an early fractionation of olivine and clinopyroxene (dunite, wehrlite) later followed by plagioclase and orthopyroxene. The highly porphyritic latest dykes may correspond to the intrusion of late wehrlitic bodies. They testify to the arrival of magma poorer in Al (fig. 2) and incompatible elements and richer in CaO (Noghreyan

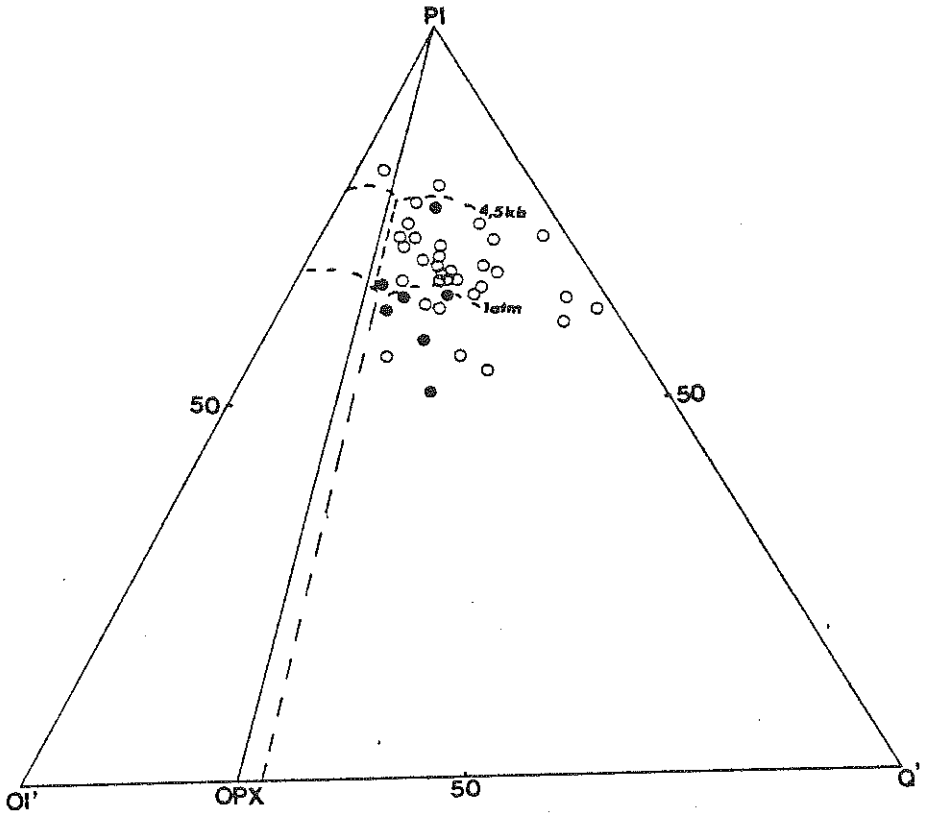

Fig. 3 : Projection from Cpx onto the plane Pl, Ol', Q' (Irvine, 1970). open circle - Dykes from the main ophiolitic sequence ; filled circle - late dykes.

et al., in prep.) issued from a strongly impoverished mantle. Discrimination diagramsbased either on major or trace-element data confirm a close affinity with immature island-arc magmatism (Lensch et al., 1979). The evolution of SiO_2 and TiO_2 during fractional crystallization may be seen in Miyashiro's diagram (fig. 3 and 4) and reveals a strong SiO_2 and a gentle TiO_2 increase. In the TiO_2 versus FeO/MgO diagram liquids fall in the low TiO_2 or isotitaniferous field (Bébien, 1980), above the very low TiO_2 (VLT) field mainly represented by the latest dykes (Bébien et al., 1980). Since Miyashiro's hypothesis (1973), this geochemical evolution of liquids, similar to those of most of the Eastern Tethyan ophiolites (Rocci et al., 1975), implies an origin of ophiolites in the vicinity of subduction zones (Ohnenstetter and Ohnenstetter, 1976 ; Beccaluva et al., 1979 ; 1980 ; Pearce, 1980).

II - THE UPPER CRETACEOUS VOLCANOSEDIMENTARY SEQUENCE OF SABZEVAR

The upper Cretaceous volcanosedimentary sequence of the Sabzevar range was interpreted as part of the ophiolite complex in North-Eastern Iran (Lensch and al., 1977 ; Lensch, 1979). A stratigraphic order has not yet been established due to the tectonized state of this sequence composed of about 90 % in volume of pyroclastic agglomerates or lava breccias and scarce pillowed or massive lavas. Pinkish pelagic limestones and greenish litharenites are slumped and associated to these volcanic deposits. The volcanosedimentary sequence outcrops around the ophiolitic members but it is always faulted against them and has never been found as a normal succession.

For the first time, four lithostratigraphic units were mapped (scale 1/50 000), between the Sabzevar-Tabas road and the Sabzevar-Soltanabad road, 15 km to the East. Because of poor stratigraphic control within the major rock units, numerous sections were obtained at different localities throughout this area . They are considered as representative of the succession. The following summary provides a stratigraphic framework to the facies description.

A - Proposed nomenclature and description of the lithostratigraphic units

1°/ - Alik formation

The Alik formation is folded in anticline and occupies a large part of the described area. The lower part of the formation is not exposed but from field evidence it can be considered as the oldest unit of the volcanosedimentary sequence. The apparent thickness of the formation is at least 100 m.

In an upward sequence the formation comprises :

- A main lava flow. The flow is dissected into huge masses surrounded by vitric, crystal and lithic lapilli tuffs or coarse ashes (Schmid, 1981). The pillow lavas of microlitic porphyritic or finely intersertal basalt, in places vesicular, are cut by some mineralogically similar dykes of doleritic diabase, with small needle-shaped phenocrysts of albite or albitized labradorite, dendritic Ti-augite, decomposed olivine phenocrysts and subhedral skeletal opaque oxide (Ti-magnetite). The groundmass is transformed into calcite, celadonite, epidote, chlorite, zeolite.

- The white breccia. This level (less 2 m to more 6 m), forms a sharp based unit of lava fragments up to 1 m, pillow segments and rarely isolated pillow-shaped bodies. These blocks are scattered in a hyaloclastite which consists of unsorted angular glassy lava fragments up to 1 m in diameter and broken crystals (plagioclase, olivine, Ti-augite) cemented by abundant white sparite. Bedding or grading

is absent. The homogeneous composition and texture of the clasts, absence of scoriae and relationship with pillow-lava flows suggest that this facies results from auto-brecciation of a surface lava flow (Carlisle, 1963) in a marine environment below the zone of wave action.

- Tuffs and sedimentary rocks. The highest levels of the Alik formation include several sequences of interbedded volcanic tuffs, greenish litharenites and limestones in beds from 1 m to 10 m. The volcaniclastic beds (Schmid, 1981 ; Fisher, 1961) mostly consist of spalled-off hyaloclastic material more or less reworked by currents. They are crystal and vitric ashes containing pebble-sized fragments somewhere graded or reverse graded which are usually amphibole-bearing andesitic tuffs with strongly zoned andesine phenocrysts, green hornblende with surrounding reaction rims, bastite and rare augite. All green-grey litharenites contain some small-scale graded beds showing weakly developed micro-cross and parallel laminations and load, flame and slump structures, but current sole marks and convolute laminations have never been seen. Commonly angular detrital material consists of volcanic detritus in a chloritic and calcitic matrix. The limestone beds vary from impure pelagic limestone with volcanic detritus to pinkish pelagic limestone with abundant planktonic foraminifera (Campanian - Maastrichtian) and absence of benthos. Slump structures are common. The presence of pelagic limestone shows that the depth was less than the C.C.D. The rarity of both sharp bases to the beds and Bouma (1962) sequences suggest that the facies accumulated by gravitational settling of pyroclastic material in an open marine environment.

2°/ - Sang-e-Sefid formation

Sang-e-Sefid formation rests in unconformity on the Alik formation and occupies a larger area and volume than the other formations, but this formation is restricted to the northern side of the Alik anticline. Its thickness, at least 300 m, can reach 600 m.

- The lower agglomerates. They are interbedded with yellowish or greenish lapilli tuffs and coarse ashes. Facies are up to 250 m thick, poorly polygenic, consisting of angular porphyritic and commonly vesicular lava fragments and crystal tuffs in a deeply weathered tuffaceous matrix. The coarser grained beds are unsorted although in some places the larger blocks are concentrated along one level. Fine grained beds sometimes show poorly developed grading or reverse grading with scattered larger clasts. They are commonly : - basaltic andesite (Nicholls and Whitford, 1976) with major phenocrysts of zoned labradorite, augite, partly resorbed olivine in a microfelsitic groundmass dominated by plagioclase, augite and secondary minerals.

- crystal tuffs with major phenocrysts of zoned labradorite, augite, bastite and rare small needles of basaltic hornblende. Vitric or palagonitic clasts are present. The matrix is usually finely calcitic.

The lapilli tuffs and coarse ashes appear as a palagonitic assemblage (Carozzi, 1960) with crystal or lithic fragments and abundant chlorite.

The agglomerates show no evidence of pillow structure and rounded clasts and although highly scoriaceous or bomb-shaped fragments are absent, it seems most likely that the facies resulted from pyroclastic eruptions and could represent pyroclastic fall out deposits. Absence of scouring or traction current structures in the finer beds suggests accumulation in a marine environment below the zone of wave action.

- The upper agglomerates. A chaotic structure is the feature of this unit the thickness of which grades from 0 to about 350 m. Masses more than 20 m thick are common in the coarsest units. Clasts are unsorted subangular to subrounded although some units contain rounded boulders. Volcanic clasts are : - porphyritic clinopyroxene-bearing andesitic basalt or basalt with major phenocrysts of albitized plagioclase (An 50-60) strongly zoned, rounded augite and scarce orthopyroxene (bastite), in a chloritic microfelsitic groundmass. - andesitic crystal tuffs with phenoclasts of zoned plagioclase (An 40-60), clinopyroxene, bastite, green hornblende and vitric fragments in a groundmass of secondary minerals (chlorite, zeolite, celadonite, calcite, quartz). Sedimentary clasts are : brecciated pelagic limestones showing abundant slump structures, which can determine huge lenticular masses at several stratigraphic levels. Zeolitic veinlets can be observed. - biomicrites with Rudiste shells occur as boulders in rare localities. The structureless agglomerates are interbedded with and rest with a sharp and erosional contact on graded greenish litharenites and mudstones. The litharenites contain mainly volcanic detritus but also some benthonic and pelagic foraminifera and algal limestones.

Subangular and angular clasts, poor sorting and absence of internal structure in the litharenites indicate that strong currents were absent in the depositional area. The lack of structure within agglomerate, sharp base to these masses and the absence of cobbles in the interbedded litharenites indicate that each agglomeratic pile was emplaced as a turbulent flow. The angularity of the clasts and lack of chilled margins can be explained if the agglomerate originated as subaerial lahars (Schmincke, 1967) which flow downslope in the deeper part of the basin. Rounded clasts and shelf limestone fragments could be reef deposits of plateform incorporated in moving lahars.

3°/ - The Solemanieh formation

The Solemanieh formation (250 m) rests in unconformity on the Sang-e-Sefid formation. It is mainly composed by regular sequences of agglomerates and palagonitic tuffs (Rittmann, 1958). No chaotic units have been observed and thin lava flows are interbedded at several levels. Intrusive rocks also occur. Lithologically the agglomerates are similar to the Sang-e-Sefid agglomerates but richer in andesites and andesitic tuffs. The agglomerates consist of : - porphyric amphibole-bearing andesites with strongly zoned and albitized phenocrysts of plagioclase (An 50-30), Mg-hastingsite surrounded by reaction rims, rare augite in a microfelsitic groundmass. - andesitic tuffs with major phenoclasts of zoned oligoclase, Mg-hastingsite augite and bastite in a chloritic groundmass. - porphyritic clinopyroxene-bearing andesitic basalt with the same mineral assemblage as those of the Sang-e-Sefid agglomerates. Four types of lava flows occur : - olivine-bearing basalts and diabases with albitized needle-shaped labradorite, partly resorbed and totaly altered olivine, Ti-augite with arborescent structures, skeletal opaque oxide (Ti-magnetite) in a hematitic groundmass. These lava flows (1 to 5 m in thickness) usually have a pillow structure. - clinopyroxene-bearing basalts. They have the same mineral assemblage as those observed in the pyroclasts in the Sang-e-Sefid agglomerate. The flows (1 to 10 m) have pillow structure. - Porphyric amphibole-bearing andesites which have the same mineral assemblage as those observed in the agglomerates. These lavas, always in massive flows, occur in the upper part of the Solemanieh formation. - - Dacitic and rhyodacitic lavas with major phenocrysts of zoned andesine, green hornblende and rare augite in a light microfelsitic groundmass. These massive lavas are often injected in fault zones.

Intrusive rocks occur in the upper part of Solemanieh formation. They are ophitic gabbros composed principally of labradorite, Ti-augite and skeletal opaque oxide. Some primary brown and green hornblende is present. Gabbros are cut by doleritic dykes which show the same mineral assemblage.

The Solemanieh formation ends with a pile of slumped pelagic limestones which have a fairly good lateral extension. Planktonic foraminifera indicate a Maastrichtian age.

4°/ - The Rud-e-Sarab formation

This formation rests in unconformity on the Solemanieh formation and lies throughout the length of the mapped area in three narrow E.W. zones, one in the median part of the area, the two others on its northern and southern borders where

the formation is faulted again harzburgite massifs. Its thickness is about 100 m. The formation consists mainly of beds of red to chocolate-brown calcareous mudstones interbedded with graded greenish litharenites and coarse-grained volcanoclastic rocks. Pelagic limestones and lava are also present.

The calcareous mudstones vary from clayed siltstones to impure pelagic limestones containing volcanic detritus and red clay both disseminated through the rock and as laminae or microstylolites. Some siltstones show ripple cross-lamination and alternating graded volcanic laminae occur. They may be either turbidites or ash falls. The pelagic limestones consist largely of partly recrystallized planktonic foraminifera (Maastrichtian and Paleocene) in a micritic matrix.

Graded greenish litharenite beds with an erosional base contain some units of the vertical sequence of Bouma's turbidites but convolute laminations are very rare. Load and flame structures are common but current sole marks have been seen only in some localities. Detrital material consists of subangular volcanic detritus including quartz, pelagic intraclasts, but also less common fragments of coralline algae and benthonic foraminifera (Miliolid). These bioclasts indicate an allochtonous character and a shallow water derivation. The internal structure and geometry of the beds allow one to interpret them as turbidity-current deposits (Kuenen, 1967).

The coarse grained volcanoclastic rocks are tuffaceous breccias or conglomerates, poorly sorted but sometimes weakly graded. Clasts are subangular to subrounded and consist of all the lavas types observed as flows in the volcanosedimentary sequences of Sabzevar. The varied textures of the clasts in the same bed suggest that they are epiclastic fragments derived from lavas with differing textures and laid down as series of debris flows.

Some thin (1 to 5 m) lava flows with a pillow structure occur ; they can reach locally more than 50 m in thickness. They are basaltic andesites strongly porphyritic with major phenocrysts of albitized and zoned labradorite with abundant fine glassy inclusions, augite and rare transformed olivine in a strongly chloritized or devitrified glassy groundmass.

B - Geochemistry

Most of the lavas observed in the volcanosedimentary sequence of Sabzevar both as flows and agglomerates are altered and/or spilitized. Thus analyses (70) without H_2O and $CaCO_3$ recalculated to 100 % were used in the present work. % Fe_2O_3 is adjusted according to the equation of TN. Irvine et W.R.A. Baragar (1971). In spite of these adjustments considerations on the alkalies shall be avoided. Some analyses of the more typical volcanic rocks are listed (Appendix I). Variation patterns of TiO_2 with FeO^*/MgO (Miyashiro, 1974) show that the lavas fall into three groups (fig. 4).

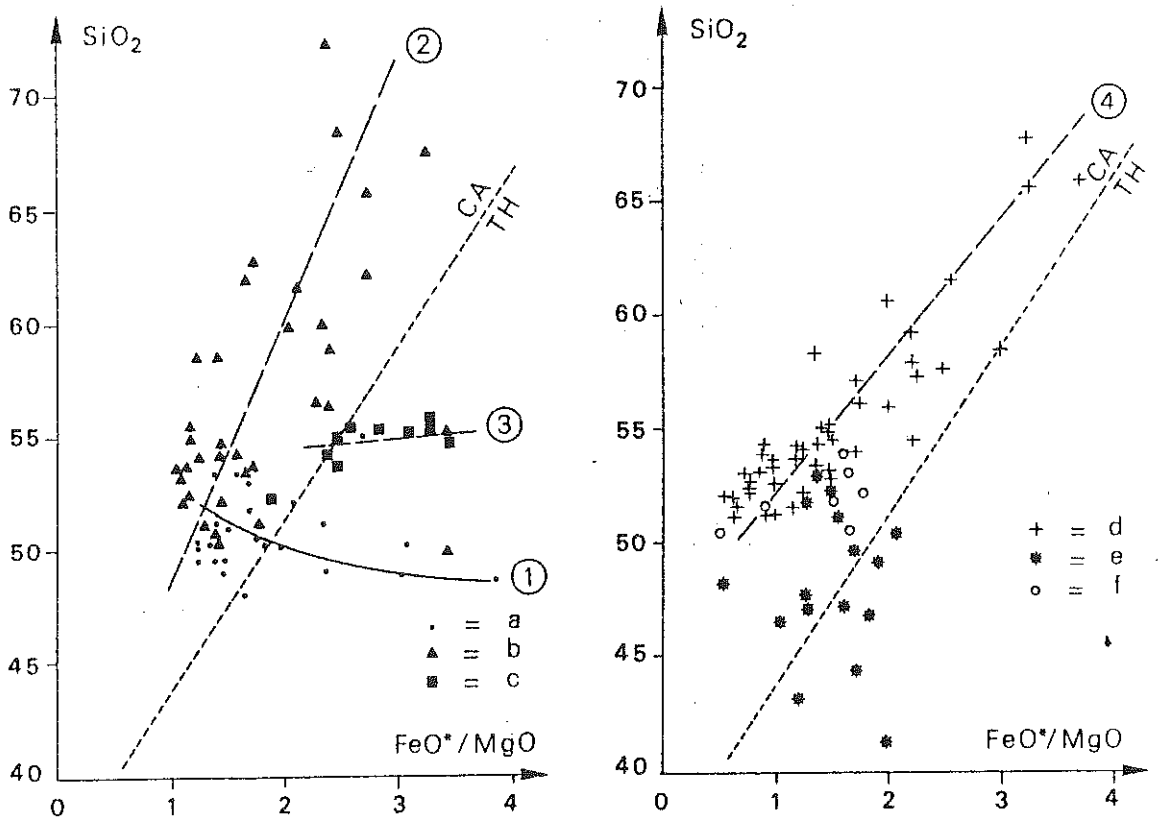

Fig. 4 : Miyashiro's diagram SiO_2 - FeO^*/MgO. Left : formations of the volcano-sedimentary sequence 1) a - Rocks from Alik and upper part of Solemanieh 2) b - Lavas from Sang-e-Sefid and Solemanieh 3) c - lavas from Rud-e-Sarah. Right : 4) d - Hypabyssal rocks from ophiolites. Metamorphic rocks : e - Amphibolites and granoblastic series f - banded schists.

In the first one, the rocks, dominantly basaltic andesites and andesites with fewer basalts and dacites, belong to the Sang-e-Sefid and Solemanieh formation. As in CA series, the TiO_2 contents, always low (0,4 to 0,8 %), decrease with increasing FeO^*/MgO. The trend of FeO^* contents is very similar (fig. 5) and in an A F M diagram they do not exhibit an Fe-enrichment (fig. 6). The slope of SiO_2 versus FeO^*/MgO is very steep. All these chemical characters and the mineralogical assemblages allow us to consider this group to be calc-alkalic.

In the second one, lavas are always basaltic andesites and richer in TiO_2 (about 1 %) and only located in the Rud-e-Sarab formation. The variations of TiO_2 contents with FeO^*/MgO show very slight increasing (fig. 4) and the FeO^* contents have a similar trend but a steep slope (fig. 5). Thus the first and second groups

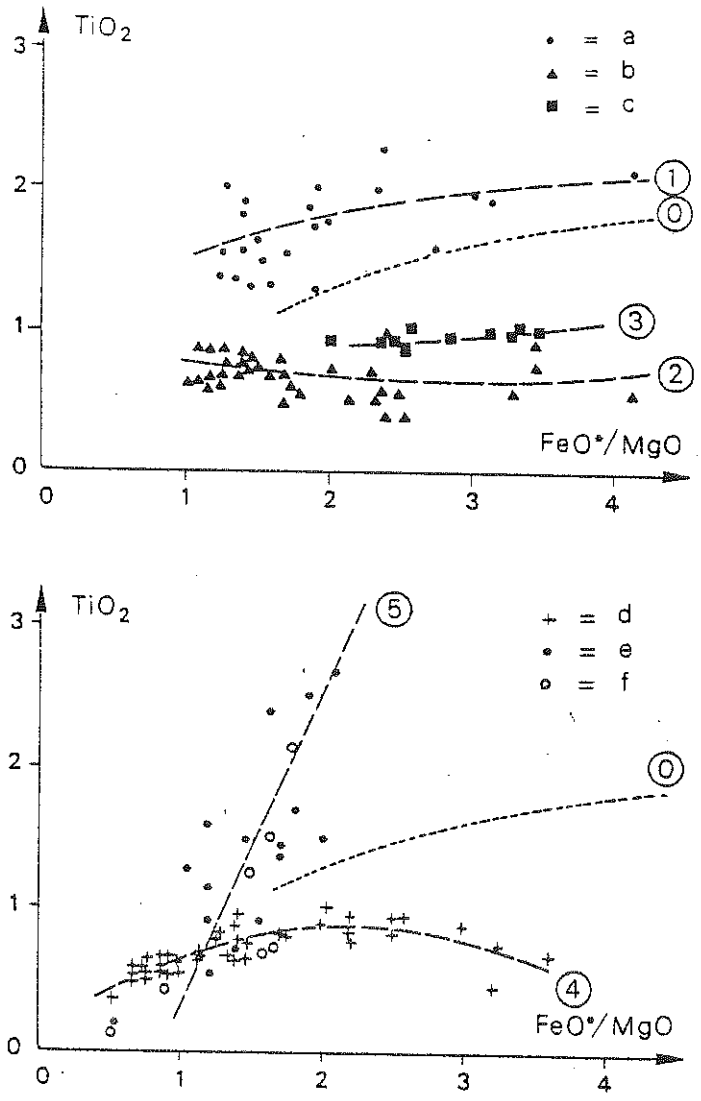

Fig. 5 : Miyashiro's diagram TiO_2 - FeO^*/MgO. Upper : 0) Limit of anisotitanian/isotitanian fields (Bébien, 1980).1) a - 2) b - 3) c - Same symbols as in Fig. 4. Lower : 5) Metamorphic rocks 4) d - e - f - Same symbols as in Fig. 4.

are well individualized even though they are both in the isotitanian field of the orogenic lavas (Bébien, 1980). In an A F M diagram (fig. 6), these rocks are the most Fe-rich but they do not exibit a trend of Fe-enrichment as TH series. The trend of this group and its location in an SiO_2 - FeO^*/MgO diagram allow to envisage for it tholeiitic affinities. The third group, basalts and gabbros with alkaline-type

mineralogical assemblages, is mainly located in the Alik formation but also as thin flows or intrusions in the upper part of the Solemanieh formation. These rocks, the richest in TiO_2 (1.70 to 2.40) fall in the anisotitanian field (fig. 4) and according to Bébien (1980) could indicate important wrench motions. Due to albitization, Na_2O/K_2O ratio is very low (0.10 to 0.45) but these rocks differ from the calk-alkaline trend by having a high iron content with a marked Fe-enrichment. They differ

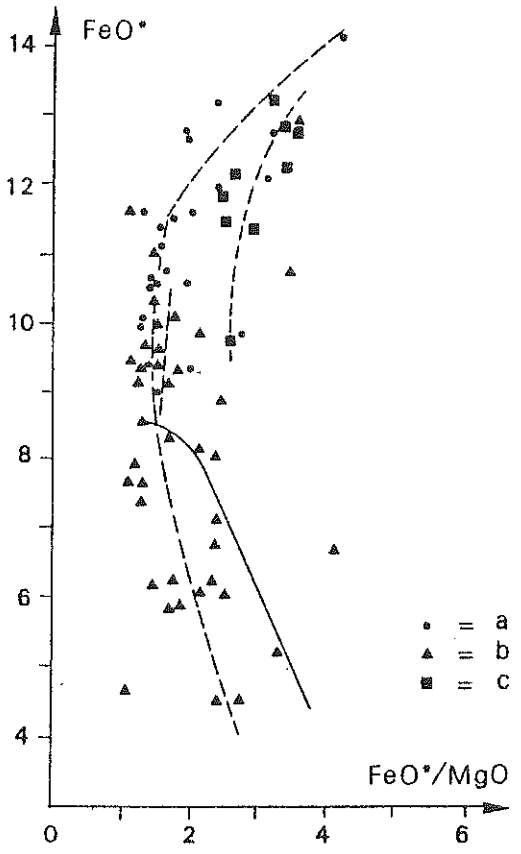

Fig. 6 : Miyashiro's diagram FeO^* - FeO^*/MgO. Volcanosedimentary sequence. Same symbols as in Fig. 4.

also from lavas of tholeiitic affinities by an increasing FeO^*/MgO when SiO_2 decreases (fig. 3). Most of these lavas have ne in the norm. For mineralogical and geochemical reasons, rocks of the third group are considered to be alkali basalts.

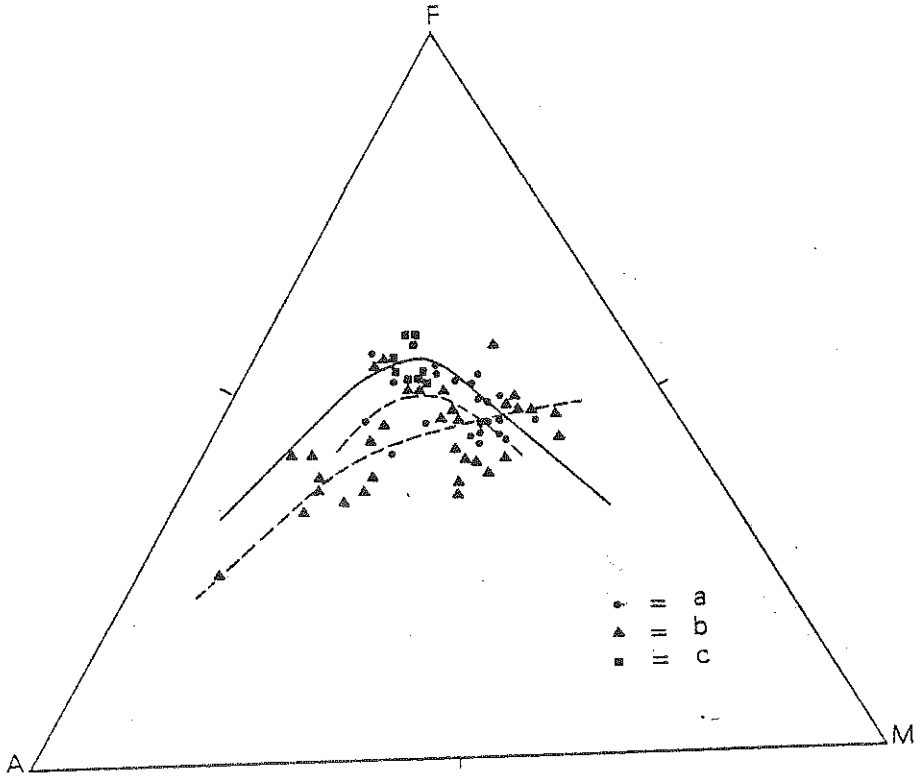

Fig. 7 : A F M diagram. Same symbols as in Fig. 4.

All the lavas of the upper Cretaceous volcano-sedimentary sequence are high in Al_2O_3 irrespective of SiO_2 contents (15 to 20 %), as generally are the rocks associated with island-arcs (Jakes and White, 1972). The lava of tholeiitic affinities in the Rud-e-Sarab formation can be compared with island-arc tholeiites because their iron enrichment is less than that of abyssal tholeiites (Gill, 1970 ; Jakes and White, 1972). According to A. Miyashiro (1974) alkaline series are not scarce in the island-arcs characterized by a slow rate of plate convergence and CA and TH series can be erupted without definite order during the different stages of island-arc formation.

The predominance of the calk-alkaline andesitic volcanism, the explosive character of the eruptions which eject mainly fragmental material (Garcia, 1978), the imbrication of the major CA series with subordinate lavas of tholeiitic and alkalic transitional characters, are typical of all associations of island-arc areas. The volcanoclastic sedimentary rocks derived both from deep sea-water lavas and from hypothetical plateforms are more often described in island-arc rather than in oceanic areas. Because of the rarity of terrigenous materials and SiO_2-rich lavas, a continental margin is unlikly.

C - Structural framework

The volcanosedimentary area of Sang-e-Sefid - Alik is bounded on its south side by a main tectonic contact which separates it from harzburgitic massifs. This contact is steep and generally dips southward (Solemanieh - Kuh-e-Siab). The area is divided into smaller structural units by reverse longitudinal faults dipping either southward or northward. The fault plane is steep, shows a wide crushed zone and is made conspicuous by serpentinite outcrops. The flattening of the series along the fault and in the fault zone is marked by well developed boudinage with y axis nearly vertical.

The internal structure of each block is marked by folding and fracturing : - The main fold system has a period broader than a kilometre ; it is weakly assymetric and its axial plane dips northward (Alik or Kuh-e-Sefid anticlines). The series are strongly folded by decametric assymetric folds (area north of Solemanieh) with axial planes of different vergences. Their axes lie between N 70°E and N 120°E and rather plunge toward the West. - Fracturing is strong and confused ; one can note two main events - 1) close sets of wrench shears with a dominant NS strike and dextral or sinistral movement, displacing EW reverse shears. 2) flat lying late shears (strike : EW - dip : 10° - 30° S) cutting across fold limits, the upper block moving northward.

The whole area is cut by two conjugate sets of regional wrench faults (Main : N 130° - 140° dextral ; subordinate : N 30° - 50°, sinistral). These faults seem contemporaneous with the main tectonic events, but they acted again later because magmatic bodies of Miocene age seem to be affected

III - METAMORPHIC AREA OF SOLTANABAD

A - Presentation

The metamorphic area of Soltanabad occurs on either side of the Neyshapour-Sabzevar road (15 km long by 2 to 4 km wide). To the West of the town the mountain mass of Kuh-e-Tchily (1 350 m) dominates the region of the Khoshab path followed by a hilly zone showing the western end of the area. The median part, at the level of Soltanabad, is hidden mostly by recent sediments while to the East can be seen the mighty mountain mass of Kuh-e-Gowin (1 780 m). The area has a main E.W. axis, slightly curved to the ESE in its western part. It is bordered :

- on the South by sedimentary series of the same kind as these studied at the part I of this article or by series reworking ophiolitic material. (Ophiolitiche resediment - Vaziri Tabar, 1976) ;

- to the North by the andesitic volcanodetritics of Chezg-Garmab of Eocene age (Lensch and Alavi Tehrani, 1977).

These contacts are mainly faulted and extremely steepy ; generally they dip strongly southward. To the North, the uncomformity of the Chezg-Garmab series on metamorphic rocks is sometimes perceptible (Khoshab road - main talweg to the West of Bolghanabad).

The metamorphic area of Soltanabad was located by Vaziri Tabar (1976) who described the main rock types. A panorama of the metamorphism of the chain, mainly directed towards the region of Kuh-e-Joghatay (70 km to the West of Soltanabad), wad given by Alavi Tehrani (1975 ; 1976) and by Lensch and Alavi Tehrani (1977), who consider these formations as being linked to ophiolitic material suggesting that they may be "tectonic wedges at the base of allochthonal ophiolite complexes". They point out traces of polycyclic metamorphism.

B - Petrographical and chemical characteristics of the metamorphic series

A detailed study of the mineral associations, mineral and whole-rock chemistry, has been carried out. The principal results of this study are given here. A detailed analytical study will be published elsewhere (in preparation by Macaudière).

1°/ - Main petrographic types

- Banded schists and associated rocks. One common aspect displayed by the rocks is a very obvious metamorphic schistosity S_1 and a compositional banding that seems primary. They are mutually interbedded sometimes on the scale of a centimetre. The original material is variable but it seems predominantly orthoderived. The principal mass is composed of medium to coarse-grained greenschists in which occur porphyroblasts of albite and epidote. Muscovite quartzites (rare), quartz-muscovite schists, albitic schists, epidote-amphibole schists, banded amphibolites and metamorphic limestones are the main types.

- Chlorite-tremolite schists. This is a heterogenous aggregate of soft schists, bright green or whitish, dotted with egg-shaped balls of varying size and composition. These schists can be found in important masses and are the cause of numerous land-slides or play a structural role by marking out the limits of tectonic lenses.

- Granoblastic rocks. These rocks are associated with the two previous types and characterized by both a granoblastic texture and a round or elliptical form. These bodies are either very tapered or in large masses. S_o/S_1 moulds itself on these bodies at every scale.

- Serpentine rocks. Serpentinite scheets, lenses or blocks are tectonically intercalated in the metamorphic series. These rocks are tenacious, massive or sometimes coarsely cleaved. Their orientation is weak or non-existent. They are composed of antigorite retromorphosed along fractures into chrysotile-lizardite.

- Kuh-e-Tchily epidote gneisses. This formation crops out in a large body measuring 3 x 1 km to the West of Soltanabad. Its contacts with other rocks are tectonic and under lined on its southern fringe by a serpentinite sheet. Variations in these quartz-bearing gneisses are a function of the size of the grain (0.2 to 2-3 mm), the structure (massive, banded, with a flaggy aspect), and the colour (bright gneiss with epidote-quartz ; dark gneiss rich in garnet - chlorite - amphibole with a tendency to be poorer in modal quartz).

- Nodular gneisses. They form a small foliated structure and possess structural and geochemical affinities with the gneisses of Kuh-e-Tchily.

- Granitoids and associated rocks. About a dozen small massifs or sills, varying in size from about 100 m to 1.5 km, are dispersed along the metamorphic area. Their composition varies from Tonalites, Quartz Diorites and Diorites but, nevertheless, their mineral associations are in equilibrium in the greenschists facies. In some places Bolghanabad, Macaudière in press), their emplacement is late in the metamorphic history, being post phase 2 (see below).

- Muscovite amphibolites. These are very particular dark facies which were only encountered close to the granitoids. They can be seen as amphibolites of variable grain but with local pegmatitic tendencies. Their most striking character is the presence of phengite flakes or rosettes slicing through the dark groundmass of the rock.

2°/ - Mineral associations

The most ordinary mineral associations are formed from four or five minerals : (± Q) - Pl. - Epid. - (± Chlor.) - Pheng. to which are often added calcic or sodiocalcic amphibole, sodic amphibole, garnet which often contains much almandine, some calcite and, occasionally few biotite. Everywhere the plagio-

clase is extremely pure albite although locally, in certain kinds of rock in the western part, a rim richer in An (up to An 15) may be seen . White mica is phengite whose degree of substitution is low to middle (Si^{IV} = 6.2 to 6.9). Although the conditions for Velde's barometer (1965) are not fulfilled ideally, it is possible to assign to the formation of the most substituted ones (in banded schists) minimal pressures of 6 kb for temperatures of 400° C. The amphiboles are very different. The most significant facts seem to be : - actinolite associated with remnants of augite in the middle of granoblastic balls, - a winchite-crossite or barroisite-crossite pair widely distributed in rocks with Ab. 99, - tchermackite or ferroan pargasite in rocks with oligoclase rims. - Biotite always highly scarce, with a late appearance and a composition not far from that of phlogopite (30 % to 35 % annite).

Two other kinds of associations can be met in rocks where the chemical composition is close to that of peridotites : - Chlor.-Trem.-Talc is a characteristic of series with basic balls. - Antig.-Trem.-Diops.-Calc.-Hemat. corresponds to serpentinites. Some pectolite prehnite fissures occur which are early or contemporary in relation to the events which sliced up the series.

3°/ - Type of metamorphism

The main metamorphic event in the Soltanabad area is a syntectonic metamorphism and after the data given by Liou et al. (1974), Brown (1976), Ernst (1978), Turner (1980), we can safely put forward temperatures between 400° and 500° and pressures between 6 and 7 kb. Two characteristics must be noted :

- metamorphic conditions are close to the transition between greenschist and blueschist facies. This transition is indicated by the crossite-barroisite association. Liknesses in paragenesis as well as in the chemistry of the amphiboles should be noted with the Sambagawa series (Banno, 1964) ;

- a final temperature elevation (550°C ?) brought these association closer to the paragenese of the epidote amphibolite facies (an increase in the percentage of An, the presence of hornblende, the rarity of chlorite). However, the amphibolite facies is not reached.

4°/ - Chemical study. A search for the protoliths

- Banded schists and granoclastic rocks. The diagram by Moine and de La Roche (1968) in fig. 8 shows the relationship between these rocks and basaltic rocks. Two tendencies are noticeable : a higher Na content (trend towards the left of the diagram) in rocks with a spilitic tendency and, a weak sedimentary contamination for

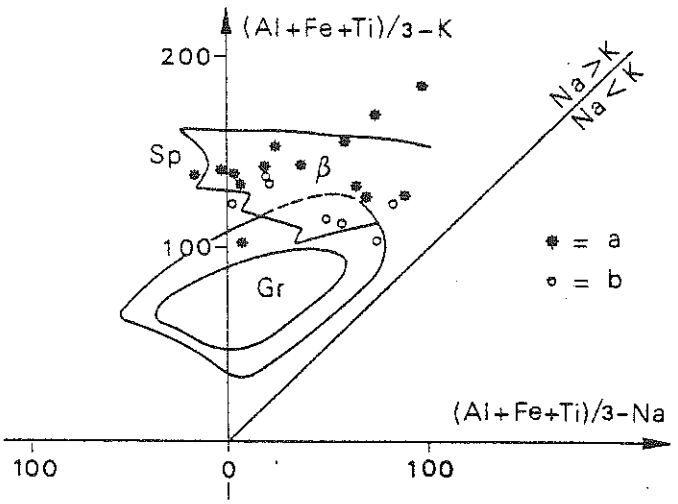

Fig. 8 : (Al+Fe+Ti) / 3 - K - (Al+Fe+Ti) / 3 - Na diagram. a - Amphibolites and granoblastic series b - Banded schists.

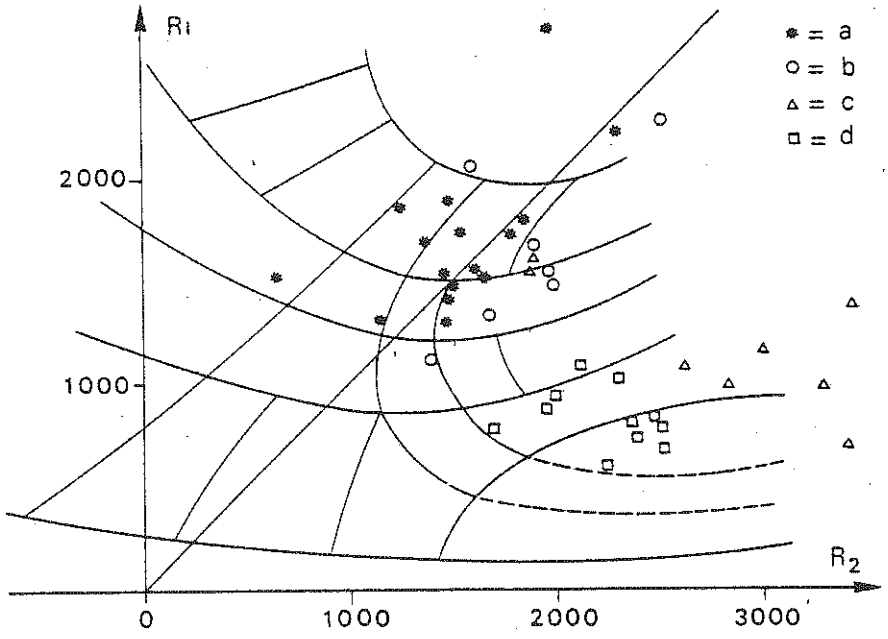

Fig. 9 : R_1 R_2 diagram. a - Amphibolites and granoblastic series.b - Banded schists c - Kuh-e-Tchily gneisses d - Granitoids.

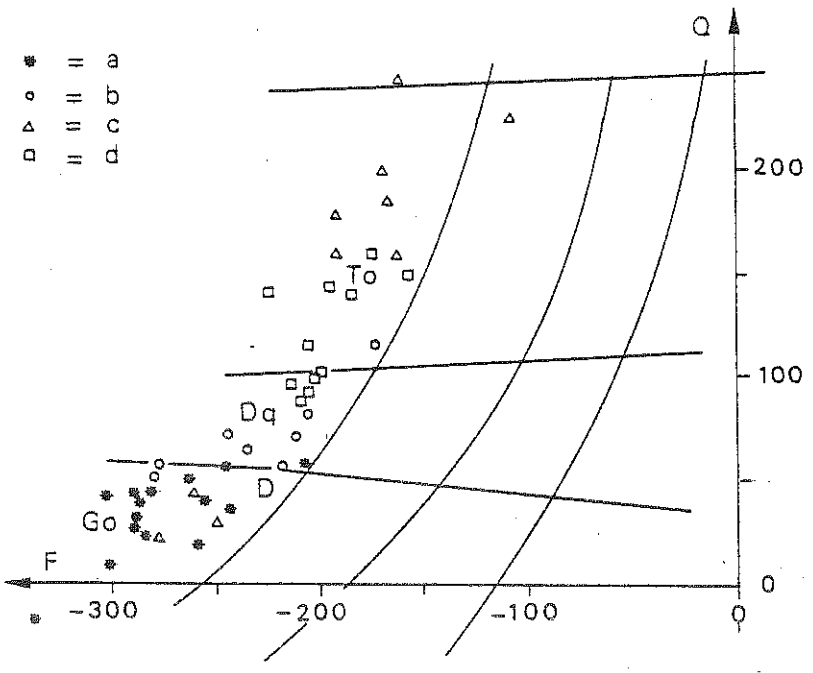

Fig. 10 : Q F diagram. Same symbols as in Fig. 9.

the banded schists (dots pointing toward the bottom and toward the right). Miyashiro's diagrams (1973) shown in figures 4 and 5 clearly distinguish these rocks from the volcanic and hypabyssal series already studied especially by their extremely "anisotitaniferous" character (Bébien, 1980). The rocks can probably be assimilated to some transitional opening series.

- Serpentine rocks and tremolite-chlorite schists. The former possess the same composition as the harzburgite mentioned by Lensch et al. (1979). They may have undergone the same metamorphic cycle but it is difficult to explain their apparently isotropic structure. The antigorite may also be the result of a low pressure metamorphism in the oceanic crust and the incorporation of these rocks into the Soltanabad area would be a purely mechanical process. The latter have global compositions close to those of the pyroxenites. Field relations, and, anterior to the metamorphism, the incorporation of basic material make us think of a mixture of material coming from serpentized harzburgite and hypabyssal basic rocks.

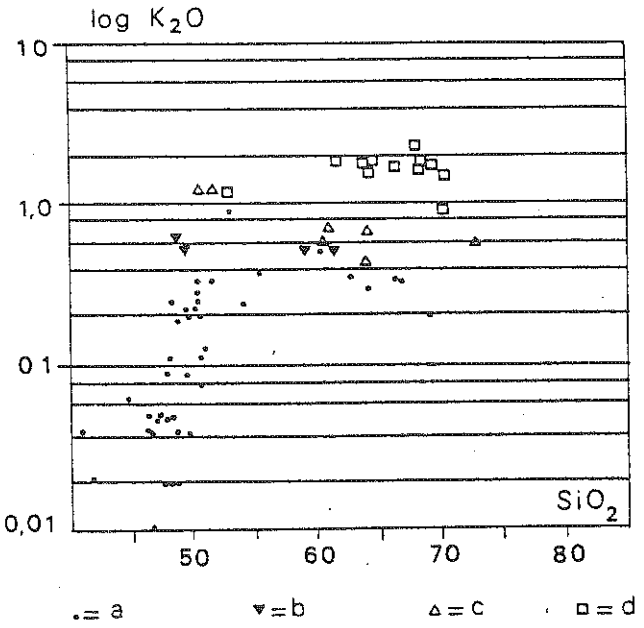

Fig. 11 : K_2O content of plutonic series. a - ophiolitic intrusive rocks b - ophiolitic gabbros from F. Vaziri Tabar (1976) c - Kuh-e-Tchily gneisses d - granitoids.

- The Kuh-e-Tchily gneisses. The characteristics of their occurrence indicate a plutonic origin ; whether we use the R_1 R_2 diagrams (de La Roche et al., 1980) or QF (de La Roche, 1964) their representative points fall into the field of tonalites (or gabbros for some very dark types) (fig. 9 and 10). Sometimes they have a large quantity of normative quartz, Q' = 40 for ANOR = 85 according to Streckeisen's and Le Maître's parameters (1979) Their percentage of K_2O is in the neighbourhood of 0.60 % and come close to the most differentiated plutonic rocks of the ophiolitic series (fig. 11) with which they have other affinities (fig. 12). This leads us to consider them as witnesses of the plutonic rocks in the ophiolitic series, although the existence of types very rich in SiO_2 poses a problem.

- Granitoid rocks. They have a well-balanced mineralogy in the greenschist facies. They possess a leucocratic character and are relatively Na rich but have a K_2O percentage in the neighbourhood of 1.8 % and sometimes higher which clearly distinguishes them from oceanic plagiogranites. In the diagrams of fig. 9 and 10 they are

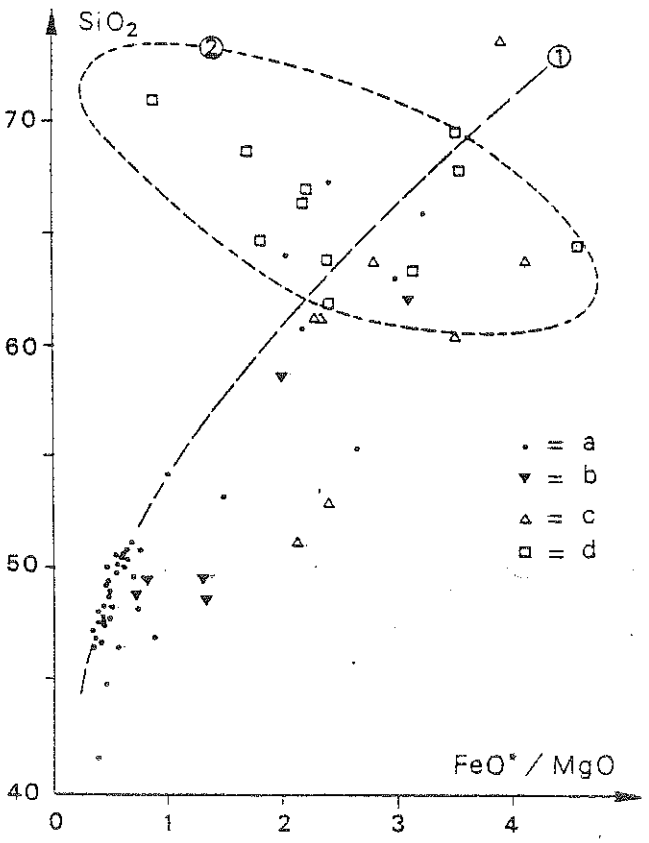

Fig. 12 : Miyashiro's diagram SiO_2 - FeO^*/MgO. Same symbols as in Fig. 11.

dispersed between diorites, tonalites and granodiorites. The granitoids are linked to the metamorphism and do not represent the final stage of the ophiolitic differentiation. Nevertheless, they represent almost mantle derived material as supported by $^{87}Sr/^{86}Sr$ = 0.7046 (A. Michard - C.R.P.G. Nancy).

C - Structural analysis of the metamorphic area - Chronology of deformation

1°/ - Structural boundaries

The southern boundary of the metamorphic area follows the outside edge of the Kuh-e-Gowin and disappears beneath a recent cover to the South of Soltanabad. A crushed, altered and highly disorganized zone brings the tremolite-chlorite schists into contact with the tectonic melanges. The movements correspond to a reverse dip slip on surfaces dipping from 50° to 80° towards the South.

The northern boundary is rarely visible since it is often hidden under alluvial fans. Sometimes it is formed by small, longitudinal faults. However, we notice that : - 2 km to the West of the Khoshab road Lutetian Nummulite limestones

(Diagnosis Brun - Société Nationale Elf Aquitaine) rest on the schists and granitoids ; - to the South of Bolghanabad the andesitic tuffs and agglomerates lie unconformably on the albite schists with a basal conglomerate dipping to 30° North.

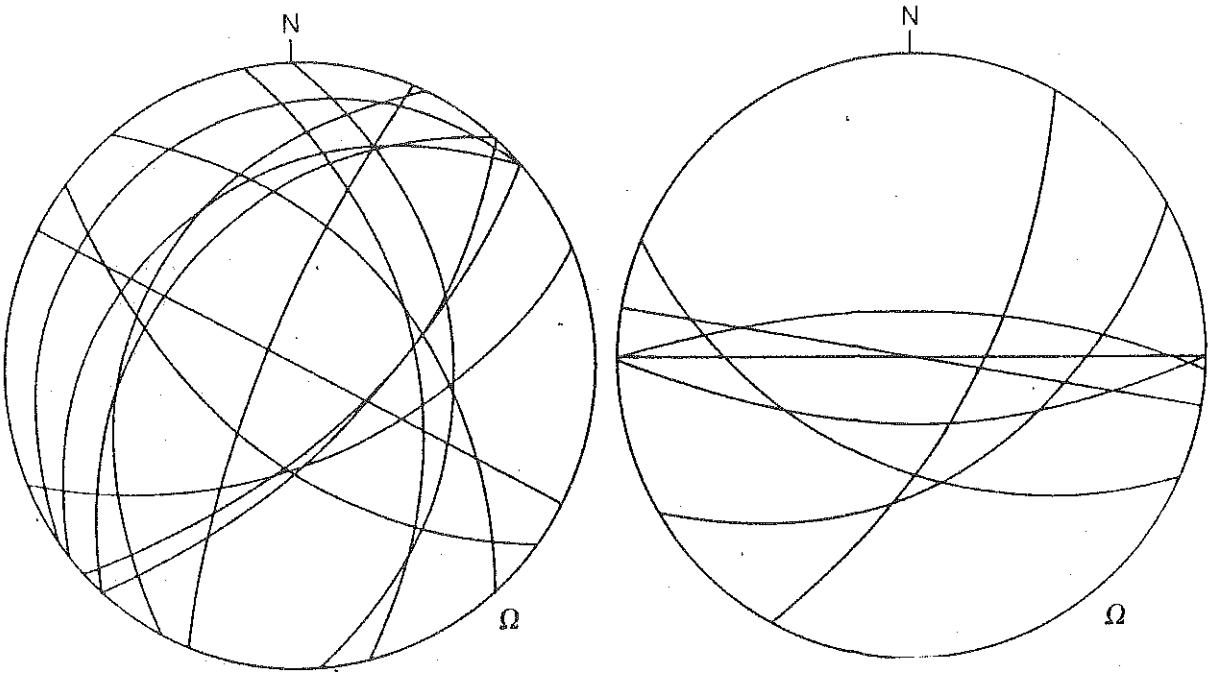

Fig. 13 : Stereographic projection of bounding surfaces of tectonic slices in the metamorphic area. Left : west and central part.
Right : eastern part (Kuh-e-Govin).

2°/ - Chronology of structural events

The detailed structural analysis of the Soltanabad area has been carried out on different scales (Macaudière, in press). The chronology of the structural events in relation to the metamorphism can be summed up in the following way :

- Prematamorphic events. Except for sedimentary phenomena corresponding to the deposition of the series which were to be metamorphized, the earliest tectonic event, which is clearly visible, concerns the creation of important mixtures of serpentinite and basic materials. We must distinguish between this event and that of post-metamorphic tectonic melanges which can, among other things, affect these same series (tremolite - chlorite schists).

- Development of the main schistosity. Throughout the whole expanse of the metamorphic area there exists a surface S_1 clearly defined, shown by the preferential orientation of the minerals. Folds P_1 are rare but always very sharp, with some boudinage and some pinch and swell structures affecting granoblastic material caught in the banded schists.

- Development of folding. One phase (2) of folding deforms S_1 by creating a family of folds P_2 which are very polymorphous according to the lithology and their localisation on megascopic structures. S_1 is locally transposed into S_2 which is a strain slip schistosity in a rather high-pressure metamorphic environment (crossite + borroisite following S_1 or S_2). Phases (1) and (2) do not represent isolated events but rather the progressive evolution of an area, in which the well-formed schistosity is deformed continuously but not to an important degree (there are no sheath folds), before metamorphic conditions come to an end.

- Development of folding P_3. Numerous folds P_3 with an axial plane mostly steep have a marked triclinic symmetry. Their axes b_3 are curved and exhibit important variations in direction and in dip. Their interferences with (2) are clear. S_3 is not very obvious and correspond to a strain slip cleavage at the hinges of the folds.

- Post metamorphic slicing . The metamorphic area is cut up into slices with retromorphism : disorganization in the soft chlorite-tremolite schists and antigorite converted into chrysotile and lizardite in the serpentine.

The bounding surface of the tectonic slices undergo important torsions and dip in such a way that the lenses seem, at the same time, folded both on N-S (western part and the southern edge) and on E-W axes (the NE part) following a geometrical pattern shown in fig. 13. In the very steep EW striking parts, we can see a boudinage with highly symmetrical pressure shadows and an Y axis following the dip of the flattening plane. This is a general characteristic of the Sabzevar mountain range and, at this stage, explains a final obvious right-angled compression with the length of the chain which steepens the tectonic limits.

- Longitudinal faulting. The faults have a strike of N 80° E to N 120° E. They clearly cut across the folded limits of the lenses and are parallel to the extension of the chain. They also affect the unconformable Eocene and can be interpreted as reverse faults linked to a moderate compression. They are displaced by the following family.

- Wrench faulting. Three major wrench faults close to N 140° E affect the metamorphic area. They displace the Lutetian sediments (Khoshab road) and shear some dacitic pods related to Miocene volcanism by Lensch et al. (1977). Their movement is typically dextral but the East compartment is generally raised (Kuh-e-Tchily, Kuh-e-Gowin).

IV - GEOCHRONOLOGICAL RESULTS

In order to clarify palinspastic reconstructions, some geochronological data have been obtained from ophiolitic diorites, granitoids and metamorphic rocks. K/Ar radiometric ages determined on ophiolitic magmatic amphiboles gave a Coniacian

age (tabl. I), which is close to the end of the expansion processes as the deposition of the following volcanosedimentary sequence took place during Campanian - Maastrichtian and Paleocene (Lensch et al., 1977 ; this study). These results agree with those of Lensch and Davoudzadeh (1981) obtained from lavas and diabases using K/Ar methods (81.2 ± 4.1 Ma ; 76.8 ± 3.8 Ma). A narrow range of ages (48-53 Ma, Ypresian) has been obtained either from amphibole or from muscovite in the metamorphic rocks and granitoids. In this last case K/Ar and Rb/Sr methods lead to similar results (tabl. I). Metamorphic events and granitoid intrusions are nearly contemporaneous as shown also by structural evidence (see above).

V - INTERPRETATION

According to structural relationships, the Sabzevar ophiolites are thought to have been generated in areas of limited extension possibly related to wrench faulting (Ohnenstetter M., 1980). The existence of short-lived ocean is also shown by the age of magmatic amphibole close to those of the volcano-sedimentary formations (Lensch et al., 1977 ; Lensch and Davoudzadeh, 1981 ; this study). According to paleomagnetic data, the present-day position of the Northern part of Iran with respect to Eurasia, similar to that in Cretaceous time (Wensink and Varekamp, 1980) precludes the existence of a large Tethys in this area during Mesozoic times. As a result, it may be proposed that the anticlockwise rotation of the Lut block (Davoudzadeh et al., 1980) might have induced, at first, shearing and thinning of the crust at the margins of microcontinents along wrench faults. These steep faults might have favoured the development of short-lived basins, some of them bearing ophiolites. These hypotheses are in agreement with paleogeographic reconstructions which imply the creation of Cretaceous rifts in Iran and Afghanistan during the approach of Eurasia and Africa, especially during the drift of India (Stampfli, 1978). In the Sabzevar range, strike-slip faults seem to have occurred on an active margin, when a change in microplate motion induced lateral movements. The compositions of ophiolitic liquids similar to those of immature island arcs (Lensch et al., 1979) are in agreement with a genesis of ophiolites at an active margin. Moreover the different intrusive events occurring during ophiolite formation show that the composition and volume of magma have changed with time, possibly in response to structural evolution.

In Maastrichtian times, during convergence, accretion processes reduced and ceased, whereas limited subduction zones were initiated along the previous lineaments. An island arc was progressively built up, characterized by explosive volcanisms. The alternating CA and A magmatisms resulted from partial melting of new mantle wedges possibly above a Benioff zone (Ringwood, 1974 ; Green, 1980). Previous faults controlled the moving submarine topography creating troughs and ridges. In troughs, the pelagic sedimentation was locally interrupted by the deposition of volcanoclastic

K/Ar dating on Amphiboles ^{87}Sr ^{87}Rb ^{40}Ar

Rock Type	K_2O (%)	^{40}Ar radio. 10^{-11} mol/g	$\frac{^{40}Ar \text{ radio.}}{^{40}A \text{ total}}$	Age ± 1σ Ma ①	Age ± 1σ Ma ②
Ophiolitic series Diorite IRN.78.154	0,114	1,471	6,2	85,5 ± 11	87,5 ± 11,1
Ophiolitic series Diorite IRN.78.172	0,052	0,6597	10,4	84 ± 22	86 ± 23
Metamorphic series Amphibolite	0,329	2,596	41	52,5 ± 2,5	54 ± 2,5
Metamorphic series nodular gneisses	0,869	6,545	54	50,5 ± 1,6	51,5 ± 1,6
Metamorphic series granoblastic lens	0,049	0,3532	11,2	48 ± 12,3	49,5 ± 12,7

Dpt : R. MONTIGNY

① $\lambda\varepsilon = 0,585 \times 10^{-10} an^{-1}$; $\lambda\beta = 4,72 \times 10^{-10} an^{-1}$; $^{40}K/Total_K = 1,19 \times 10^{-4}$ mol/mol

② $\lambda\varepsilon = 0,581 \times 10^{-10} an^{-1}$; $\lambda\beta = 4,962 \times 10^{-10} an^{-1}$; $^{40}K/Total_K = 1,167 \times 10^{-4}$ mol/mol

K/Ar dating on Muscovite

Rock Type	K_2O %	^{40}Ar radio 10^{-11} mol/g	$\frac{^{40}Ar \text{ radio}}{^{40}Ar \text{ total}}$	Age Ma ± 2
Granitoid of Bolghanabad	7,98	72,3	75 %	51,5 ± 2
idem	7,98	72,6	77,5 %	51,7 ± 2

Dpt : J.L. ZIMMERMANN - Laboratoire de Géochronologie
C.R.P.G. Vandoeuvre-lès-Nancy.

$^{Rb}/Sr$ dating - Dpt : A. MICHARD - Laboratoire de Spectroscopie isotopique
C.R.P.G. Vandoeuvre-lès-Nancy ($\lambda Rb = 1,42\ 10^{-11}$)

Rock Type	Rb ppm	Sr ppm	$^{87}Sr/^{86}Sr$	$^{87}Rb/^{86}Sr$	$(^{87}Sr/^{86}Sr)_0$	Age_{Ma}
Granitoid Bolgh Muscovite	220	52,2	0.71360	12,0	0,7046	53 ± 1
Granitoid Bolgh Whole rock	37	460	0,70480	0,232		

Table I - Geochronological data -

sedimentary rocks transported in turbidity currents. On ridges, shallow water carbonates were deposited. The tectonic instability of the sea floor, due to vertical movementsisattested by the abundance of slumps, synsedimentary faults and by the large variability in thickness of the formation. Similarly, in the Naïn area, surrounding the Central East Iran microplate, sedimentary basins, sometimes with an ophiolitic substratum, were organized in horsts and grabens limited by faults which controlled erosion, lack of sedimentation, rapid subsidence and late uplift which ended the trough (Tehrani, 1978). No evidence of emergence has been pointed out in the Rud-e-Sarab formation during Paleocene, but on the South of the Sabzevar zone, in the area of Oryan, pelagic sediments were followed by neritic formations (Lensch and Davoudzadeh, 1981), indicative of rapid filling of the trough during upper Cretaceous. Soon after the deposition of the Rud-e-Sarab formation the Sabzevar area underwent major E-W folding. As a consequence, Lutetian flysch-type sediments rest with an angular unconformity on the volcanosedimentary sequence.

In the northern part of the range, the study of the metamorphic rocks reveals that the thermal culmination of the more important metamorphic event occurred just after the deposition of most of the upper Cretaceous volcanodetritic series. This metamorphism of homogeneous grade across the whole area was contemporaneous with the main structural event leading to the building of the Sabzevar range. These facts and the absence of previous relictual deformation indicate that the metamorphic rocks did not result from the breakdown of a possible dynamothermal sole (Malpas, 1979 ; Jamieson, 1981) beneath a mantle slice as in some ophiolites. However, the geochemistry of basic metamorphic rocks, strongly different either from ophiolitic material or volcanosedimentary series but similar to amphibolites from some other ophiolites, may infer a derivation from a common source. It could be a transitional to alkalic series linked to the opening of the oceanic basin (Bébien et al., 1980). In the Sabzevar range, these formations as well as some ophiolitic fragments (Kuh-e-Tchily gneisses)were metamorphosed during the accretion of the ophiolitic crust against its margin, possibly when the upper Cretaceous - Paleocene volcanosedimentary series were deposited. The emplacement of K_2O-rich granitoids in the banded schists at the end of the metamorphic processes testifies to an evolution towards a more mature island arc. During the late stage of this structural evolution, pervasive fracturing led to the uprise of the chain. These main tectonometamorphic events as well as the folding of the Cretaceous volcanosedimentary series took place at the beginning of the collision between India and Asia from Paleocene to Miocene. During this time rotation of the Lut block persisted (Conrad et al., 1981). Since the Lutetian, the uplifted ophiolitic belt was eroded. At the same time magmatism caracteristic of mature island arc went on (Lensch et al., 1977). Basalt and andesite, dacite and microdiorite along stretching faults and alkali basalt and shoshonite were

intruded in and outpoured on the ophiolites respectively during Eocene, Oligo-Miocene and Miocene (Lensch et al., 1980).

A - Concluding remarks

The reconstruction of the evolution of the Sabzevar belt has led to a single model which emphasizes the role of a strike-slip system from the opening to the closure of the ophiolitic basins (Ohnenstetter M., 1980). The creation of their margins was linked to the splintering of microplates triggered during lateral movements of shifted large continents. At the onset of strike-slip regime transtension superseded transpression (Robertson and Woodcock, 1980) favouring the thinning of continental crust and the filling of volcanosedimentary basins. The alteration of tensile and compressive zones along the chain and with time was responsible for large lateral and vertical variations during the ophiolite and later volcanosedimentary formation. The shearing and shortening of the blocks forming accretionary wedges result from the progressive predominance of transpression as the bringing of continents proceeded. No obduction would be infered if strike-slip motion occur during the tectonic emplacement of ophiolites (Brookfield, 1977). CA magmatism and metamorphism might develop along local subduction zones. Protrusion of serpentinites and upheaval of deep parts of the mountain roots were favoured in temporary tensile areas. All these successive events might be linked to an oblique-collision model (Badham and Halls, 1975). The last stages of purely compressive pre Lutetian tectonic phases reinforced the vertical tectonic fabrics of the Sabzevar belt.

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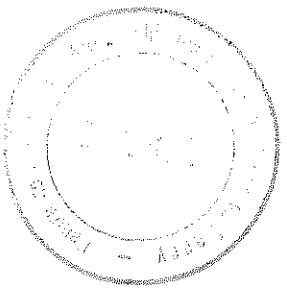

A N N E X E I

	1	2	3	4	5	6	7	8	9	10
SiO_2	53,34	52,12	44,33	43,54	48,64	57,08	60,23	45,32	46,33	53,03
TiO_2	0,81	0,37	1,51	0,59	0,72	0,54	0,52	1,21	2,26	1,03
Al_2O_3*	15,49	12,87	15,73	14,47	14,69	17,53	17,27	15,81	16,40	15,37
Fe_2O_3	7,23	6,47	9,39	8,98	8,40	6,95	6,01	8,39	11,34	12,29
CaO	7,54	12,52	10,24	14,48	11,65	3,86	2,44	14,54	9,51	6,01
MgO	5,08	11,81	4,89	6,39	5,84	2,92	2,82	5,65	4,79	3,69
MnO	0,11	0,09	0,19	0,15	0,20	0,08	0,12	0,12	0,17	0,13
Na_2O	3,44	1,73	4,43	2,71	2,83	7,67	7,79	2,97	4,62	3,98
K_2O	0,60	0,10	1,30	0,50	0,25	0,94	0,88	0,87	0,46	1,09
L.I.	5,91	1,76	7,84	7,96	6,74	2,26	1,94	5,16	4,03	3,15
Total	99,55	99,84	99,85	99,77	99,96	99,83	100,02	100,04	99,91	99,77

Ophiolitic series : 1 - basaltic andesite from the main intrusion event 2 - basalt from the latest dyke. Upper Cretaceous post ophiolitic sequence. 3 - Basalt from the Alik formation 4 - basalt 5 - basaltic andesite 6 - andesite 7 - dacite from the Sang -e-Sefid and Solemanieh formation 10 - basaltic andesite from the Rud-e-Sarab formation.

Analyst R. Montanari, Laboratory of Petrology, Nancy I.

A N N E X E II

Analyses from metamorphic area

The whole rock analyses are available from "Fichier Artemise"
C.R.P.G. Vandoeuvre-lès-Nancy, France. Main Types are only given here :

	1	2	3	4	5	6	7	8	9	10
SiO_2	53,31	51,76	50,37	49,63	51,63	39,09	73,60	61,29	70,84	63,54
TiO_2	0,73	1,24	2,68	1,47	0,06	0,01	0,22	0,52	0,09	0,41
Al_2O_3*	17,13	13,69	13,36	14,45	3,92	0,78	12,86	14,37	15,63	17,56
Fe_2O_3	8,21	9,94	13,31	11,72	6,83	10,56	3,59	8,61	0,74	3,65
CaO	5,45	9,49	8,83	8,82	10,76	0,06	3,58	6,72	2,43	5,15
MgO	4,46	5,98	5,76	6,21	22,34	35,21	0,83	3,39	0,79	1,05
MnO	0,14	0,17	0,26	0,18	0,08	0,16	0,03	0,16	0,01	0,03
Na_2O	4,59	2,74	4,06	3,53	0,19	0,12	3,45	1,91	5,82	4,92
K_2O	1,30	1,08	0,16	0,31	0,06	0,02	0,60	0,65	1,57	1,73
P_2O_5	-	-	0,38	0,14	-	-	-	-	-	-
L.I.	4,46	3,70	0,82	2,32	3,90	12,44	1,01	2,33	1,61	1,52
Total	99,78	99,79	99,99	98,78	99,77	98,45	99,77	99,95	99,53	99,56

(1) JM8084 - Banded schists (2) JM8098 - Banded schists (3) JM7029 - Banded amphibolite (4) JM7097 - Granoblastic basic body (5) JM 7106 - Tremolite - chlorite schist (6) JM 8051 - Antigorite serpentinite (7) JM 8059 - Kuh-e-Tchily gneiss (8) JM8057 - Kuh-e-Tchily gneiss (9) JM8034 - Granitoid W Soltanabad (10) JM8111 - Granitoid S. Bolghanabad.

(3) (4) From Laboratoire d'analyse quantométrique - C.R.P.G. Vandoeuvre - The others from Laboratoire Nancy I

TROISIÈME PARTIE

LES FRACTIONNEMENTS DANS LES OPHIOLITES. RÉFÉRENCE PARTICULIÈRE À LA CORSE ET AU VOURINOS

"Theory or experiment whithout reference to the rocks is like Gauguin without reference to colour".

H.J. Greenwood (1979)

TROISIÈME PARTIE

LES FRACTIONNEMENTS DANS LES OPHIOLITES

RÉFÉRENCE PARTICULIÈRE À LA CORSE ET AU VOURINOS

I - DE LA NECESSITE DU FRACTIONNEMENT

II - ETUDE QUANTITATIVE DU FRACTIONNEMENT

III - IMPORTANCE DE LA CONNAISSANCE DES MASSIFS DE CUMULATS

IV - LE CHEMINEMENT A TRAVERS LES ARTICLES REGROUPES

TROISIÈME PARTIE

LES FRACTIONNEMENTS DANS LES OPHIOLITES

RÉFÉRENCE PARTICULIÈRE À LA CORSE ET AU VOURINOS.

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Les deux premières parties du mémoire ont permis de faire apparaître de façon empirique, les discontinuités géochimiques, structurales et stratigraphiques des ceintures ophiolitiques. En replaçant celles-ci dans leur contexte géologique plus globale, la durée d'accrétion a pu être estimée et le lieu des premières émissions supposé. Des sites géotectoniques possibles ont aussi été proposés à partir des comparaisons entre les ophiolites et les équivalents magmatiques et tectoniques actuels. Cependant, la nature des discontinuités géochimiques mérite d'être précisée ainsi que d'une manière générale les mécanismes se déroulant en profondeur. En particulier le lien unissant péridotites, cumulats et laves dans chaque assemblage doit être démontré d'un point de vue géochimique, de même que les variations de la composition des magmas dans le temps et l'espace. C'est seulement alors que des contraintes supplémentaires seront appliquées pour apprécier l'environnement géotectonique initial.

Cette partie s'attache plus spécialement à la mise en évidence de fractionnements distincts dans les zones de basses pressions (chambres magmatiques) et quelquefois dans le manteau. Le choix des massifs a été fait en fonction de l'importance des discontinuités géochimiques observées que reflète la classification pétrologique des ophiolites. Pour réaliser ce travail, plusieurs méthodes ont été employées, basées soit sur le comportement des éléments majeurs, soit sur celui des traces. Ce sujet s'intègre donc aussi dans un cadre plus général portant sur la genèse des basaltes qui sont, à l'opposé des liquides ophiolitiques, rarement associés à des cumulats, témoins des fractionnements.

I - DE LA NECESSITE DU FRACTIONNEMENT

La surface du globe, surtout celle des océans est recouverte de basaltes dont les moins évolués présentent une certaine homogénéité quant à leur concentration en MgO. La rareté des picrites, excepté dans l'Archéen, a conduit à la suite des hypothèses d'O'HARA (1965) à de nombreuses controverses sur l'origine primaire ou dérivée des liquides observés en surface. Si un certain consensus se dessine en faveur de la seconde hypothèse -on expliquerait aussi les teneurs plus ou moins constantes en MgO par la faible densité des liquides (fig. III.1) pour 9 % <MgO <11 % (HUPPERT et SPARKS, 1980 ; SPARKS et al., 1980 ; STOLPER et WALKER, 1980)- la nature des fractionnements et fusions partielles conduisant à la composition des liquides observés est encore très hypothétique. A priori,

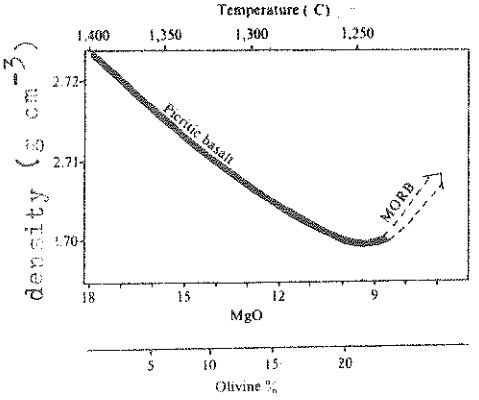


Fig.III.1 - Variation de la densité du magma en fonction de sa teneur en MgO ; tiré de HUPPERT et SPARKS (1980).

il est possible d'envisager des fractionnements de basses pressions liés à l'existence de cumulats et des fractionnements de hautes pressions que l'on peut supposer à cause de la longue route entre zone de fusion et chambre crustale lors de la montée de liquide à P et T décroissant (BOTTINGA, 1974). Chaque assemblage ophiolitique qui comporte un résidu, des cumulats et des liquides (laves et roches hypabyssales) constitue alors un excellent champ d'investigation pour comprendre l'évolution des magmas. Dans ce cadre, la composition du magma parental ne peut pas seulement être estimé à partir de la "croûte" ophiolitique au-dessus du manteau pétrologique (fig. III.2) mais il faut aussi tenir compte des extractions péridotitiques effectuées dans les magmas ascendants comme en Corse (OHNENSTETTER M ; IV.4).

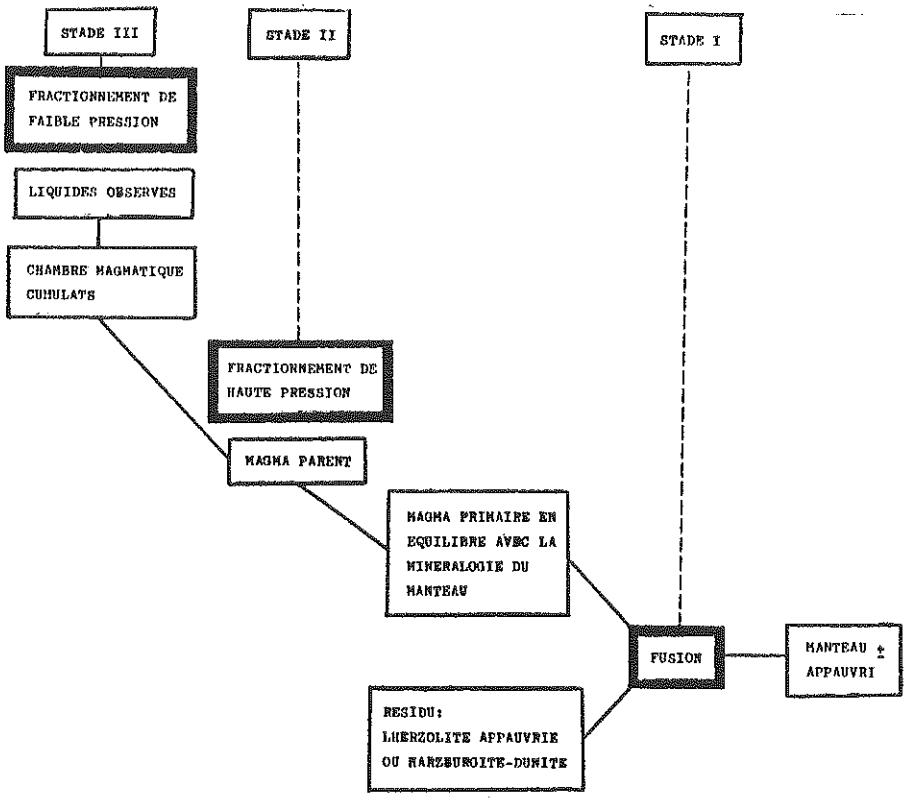


Fig. III.2 - Cheminement possible des magmas depuis la fusion jusqu'aux liquides observés. I : fusion partielle ; II : fractionnement de haute pression ; III : fractionnement de basse pression. Le stade II a été mis en évidence en Corse. On peut passer aisément du stade I au III et calculer les extraits nécessaires à la composition du magma parental en équilibre avec le manteau à partir des cumulats inférieurs des séquences ophiolitiques.

Notre démarche dans les travaux sur le fractionnement a toujours été de considérer le dernier produit obtenu pour progressivement rechercher les liquides parents et enfin les magmas primaires. Nous avons évité de spéculer directement sur la nature des magmas primaires en équilibre avec le manteau bien que la simple observation du nombre des phases dans les résidus de Méditerranée occidentale (4-5 phases) et de Méditerranée orientale (3-4 phases) implique, dans chacun des cas des liquides aux concentrations et aux rapports d'éléments majeurs très variés. Etant donné la diversité pétrologique des ophiolites, on peut choisir pour démontrer les types de cheminements possibles plusieurs séquences cumulatives parmi les deux types principaux d'ophiolites. On considérera d'une part le magmatisme ophiolitique corse voisin de celui des dorsales actuelles qui a pu durerpendant 40 MA et où un certain équilibre a été réalisé

depuis la fusion partielle jusqu'au fonctionnement des chambres magmatiques, d'autre part, on étudiera certaines évolutions du magmatisme des assemblages de Méditerranée orientale, qui sont plus variées dans le détail bien que de durée plus brève (2-10 MA). En particulier, on examinera au Vourinos, deux types d'émissions décalées dans le temps, la seconde (Asprokambo) étant plus sursaturée que la première (Krapa).

II - ETUDE QUANTITATIVE DU FRACTIONNEMENT

La quantification du fractionnement qui résulte d'extraction de solide à partir de liquide peut être abordée de plusieurs façons. Toutes conduisent à des imprécisions en partie liées à la multiplicité des facteurs intervenant lors de l'évolution d'une chambre magmatique ; signalons principalement l'état d'équilibre entre les minéraux et la nature du système, ouvert ou fermé (O'HARA, 1977) (fig. III.3) . Pour calculer les extraits responsables de l'évolution des liqui-

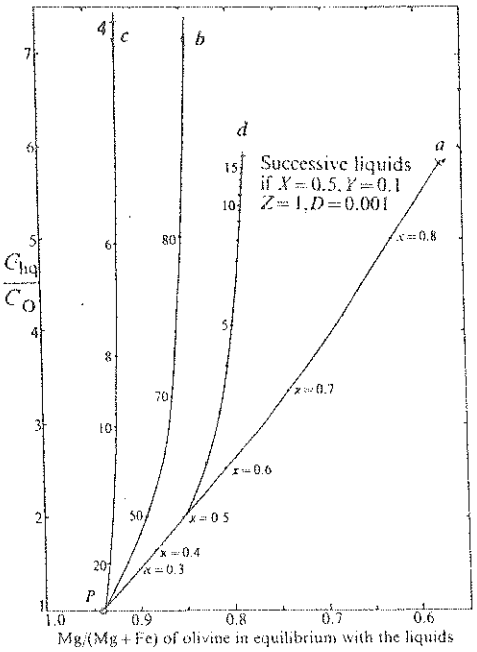


Fig. III.3 - Variation de la concentration relative d'un élément incompatible C_{liq}/C_o, en fonction de Mg/Mg+Fe dans l'olivine en équilibre avec le liquide assumant $K_D = 0.33$. Le rapport C_{liq}/C_o correspond à (1-F) avec F, fraction de liquide résiduel selon la loi de RALEIGH. La courbe a montre l'évolution d'un liquide en système fermé selon une cristallisation fractionnée parfaite, x étant la fraction solide cristallisée. La courbe b correspond à une cristallisation à l'équilibre parfaite ; la courbe c montre la variation de C_{liq}/C_o lorsque le taux de fusion varie. Des courbes intermédiaires entre a et b peuvent être obtenues si la chambre est périodiquement remplie (d'après O'HARA, 1977).

des, on peut comparer le comportement des incompatibles à celui du rapport Zr_o/Zr qui représente la fraction de liquide résiduel selon le fractionnement de Raleigh (TREUIL et JORON, 1975 ; MINSTER et al., 1977 ; BECCALUVA et al., 1977, cette partie). Il implique une cristallisation fractionnée parfaite. Dans ce cas, choisir le basalte le moins évolué pour calculer $Zr_o/Zr = 1$, c'est, si l'on ignore avec quel solide il a été en équilibre, admettre qu'il représente la composition du liquide entrant dans la chambre, c.a.d. le magma parental (fig. III.4). Cependant, comparer l'évolution de ce rapport à celui de FeO/MgO dans les liquides

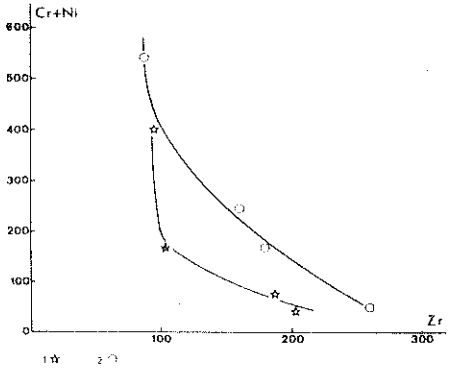
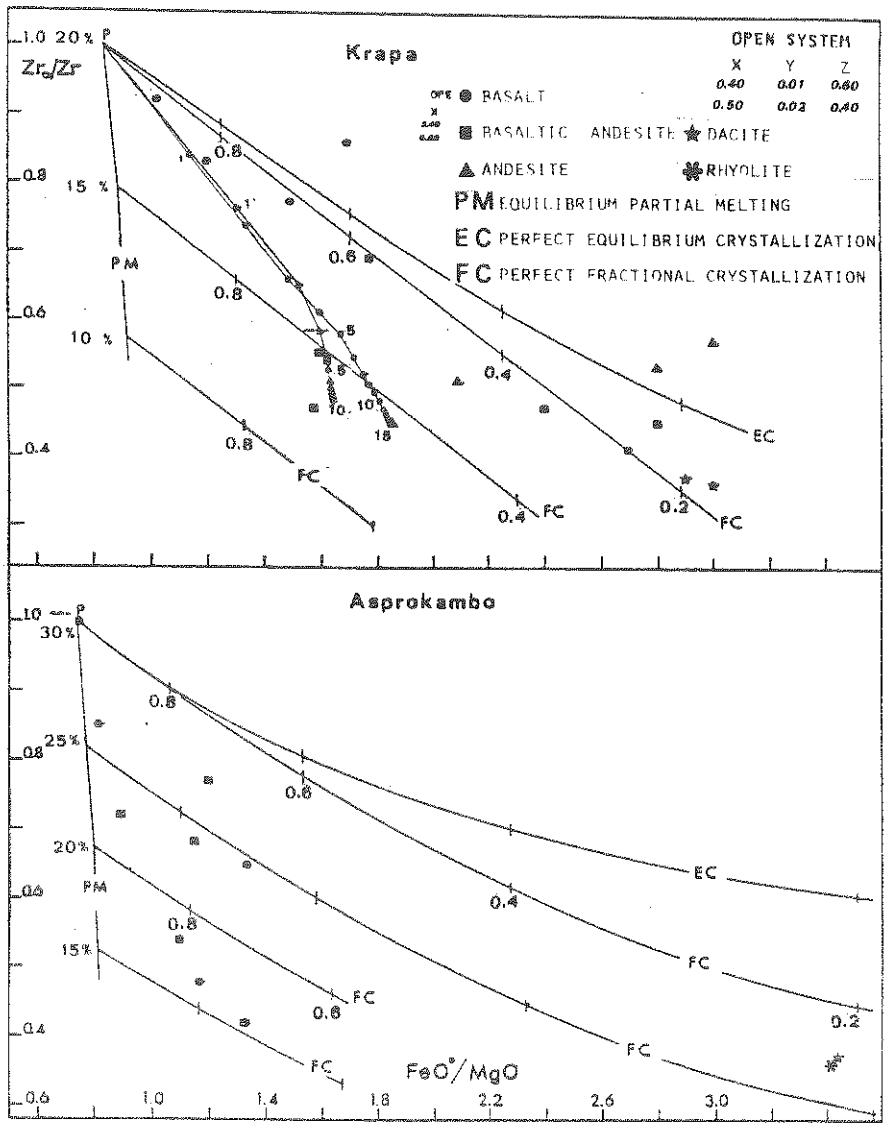
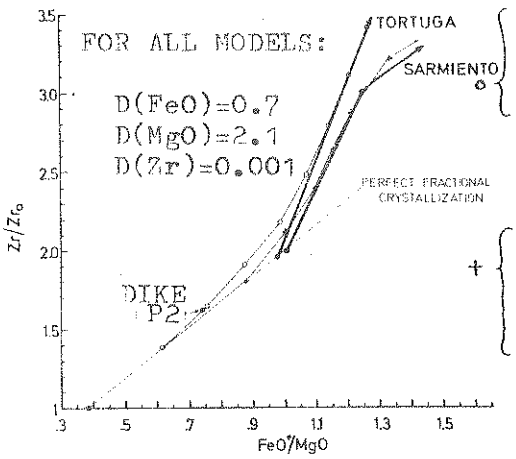


Fig. III.4 - Détermination de la concentration minimum de Zr dans les liquides observés afin de choisir Zr_o, la concentration initiale, pour appliquer le fractionnement de Raleigh à une série ; tiré de BECCALUVA et al.,(1977 ; III.4).

ou les solides associés permet d'obtenir des indications supplémentaires sur l'état du système et les cheminements possibles des magmas. En particulier, cette confrontation met en évidence le découplage des éléments majeurs par rapport aux traces (O'HARA, 1977 ; STERN, 1979) qui selon O'HARA et MATTHEWS (1981) serait principalement dû à un système ouvert (fig.III.5).Si cette technique permet de juger quel solide a pu être en équilibre avec le liquide d'après la répartition de Fe et Mg entre les minéraux (ROEDER et EMSLIE, 1970), une certaine approximation persiste quant à la nature et la proportion des cumulats extraits (STERN, 1979), et quant au rôle du fractionnement, par rapport à la fusion à l'équilibre.

Il est aussi possible de calculer des extraits à partir de la composition des phénocristaux, des minéraux cumulus ou des roches totales et de confronter la composition des liquides observés et calculés (COX et al., 1979). Si ces calculs sont traités de façon linéaire, ils peuvent donner une approximation comparable à celle obtenue par la méthode des moindres carrés (BANKS, 1979). Le





29% CRYSTALLIZATION (X)
1% ERUPTION (Y)
STEADY STATE MAGMA INPUT (Z)
PRIMARY MAGMA
FeO=7.85 wt %
MgO=20.15 wt %
Zr_o=21 ppm

\+

45% CRYSTALLIZATION (X)
30% ERUPTION
20% DECREASE OF MAGMA INPUT (Z)
PRIMARY MAGMA:
FeO=6.85 wt %
MgO=18.11 wt %
Zr_o=25.5 ppm

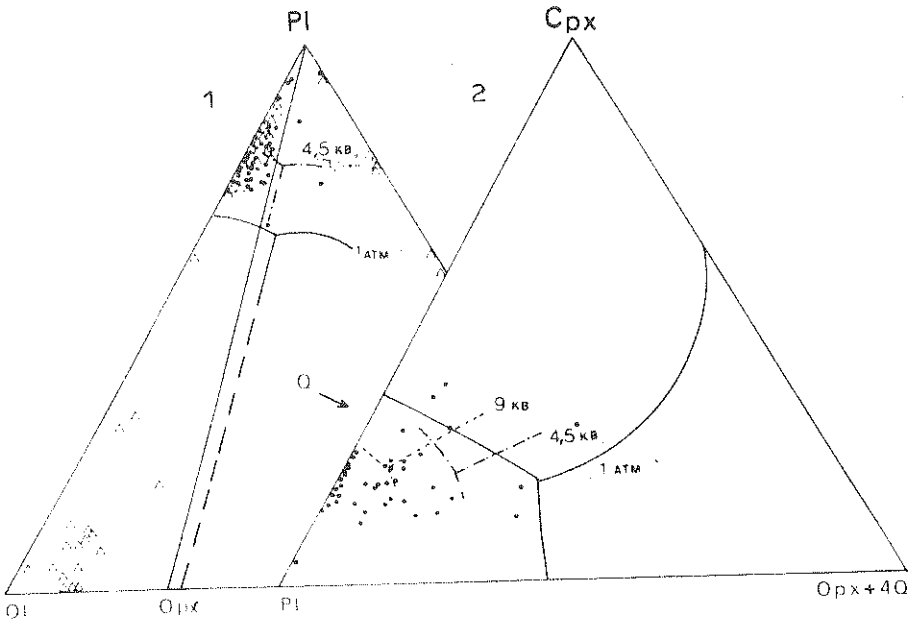
Fig. III.5 - Variation de la concentration relative de Zr/Zr_o en fonction de FeO/MgO dans les liquides de deux séries ophiolitiques du Chili(A) confrontée aux calculs effectués pour un système ouvert ; tiré de STERN (1979). B : même diagramme pour les séries de Krapa et Asprokambo, tiré de BECCALUVA et al. ; (III.4) . La série de Krapa est compatible avec un système ouvert et des cristallisations plus ou moins à l'équilibre. Celle d'Asprokambo est compatible avec un système fermé et une cristallisation fractionnée plus parfaite.

calcul des bilans de masse à partir de la composition des phénocristaux conduit à des résultats erronés par suite soit du retrait sélectif de certains minéraux comme dans la série de Krapa (OHNENSTETTER M., cette partie), soit par suite de cristallisation plus tardive mais effective en position intercumulus comme en Corse (BECCALUVA et al., 1977). Par ailleurs ce type de calcul ne tient pas compte du liquide interstitiel piégé dans les cumulats (MAALOE, 1976).

Bien qu'en système ouvert, le pourcentage d'extraction calculé à partir des incompatibles ou des compatibles s'écarte du fractionnement réel (O'HARA et MATTHEWS, 1981), il peut apparaître intéressant d'effectuer un bilan entre solide et liquide afin de mieux juger quel degré de fractionnement est nécessaire avant l'éruption des liquides . Cette approche est d'autant plus justifiée lors d'une étude comparée des ophiolites que la composition et l'évolution des éléments majeurs montrent une très grande variabilité qu'un système ouvert seul ne peut expliquer : en première approximation, les liquides sont contrôlés par les phases au liquidus appropriées (OHNENSTETTER et al., 1979 ; III.1 ; OHNENSTETTER M., III.5). De plus, si l'emploi de ces calculs est fait en se référant à la nature et à la texture des roches, les bilans de masse peuvent alors représenter avec une bonne probabilité les réactions se déroulant dans la série (GREENWOOD, 1979).

III - IMPORTANCE DE LA CONNAISSANCE DES MASSIFS DE CUMULATS

En effet, quelles que soient les approches employées, les fractionnements proposés sont d'autant plus valables que les suites cumulatives sont mieux connues. La mise en évidence des types de fractionnement, au moins dans les zones de faibles pressions, peut se faire de façon empirique en décrivant la nature, l'ordre de cristallisation et la proportion des minéraux cumulus au sein des principales séquences. Le rôle des phases au liquidus sur l'évolution des liquides peut être constaté sur des diagrammes de projection établis à partir du tétraèdre Ol-Cpx-Pl-Q (IRVINE, 1970). Dans le cas de la Corse (OHNENSTETTER et OHNENSTETTER ; 1976), les projections à partir de l'olivine et du clinopyroxène mettent en évidence l'importance de la cristallisation cotectique du plagioclase et de l'olivine avant d'être rejoint par le clinopyroxène et surtout l'orthopyroxene. Les liquides de deux massifs de Méditerranée orientale ont été également projetés : Vourinos (PAUPY, 1976) et Sabzevar (NOGHREYAN, 1982). Le caractère sursaturé des liquides est net, la plupart d'entre eux ayant quitté le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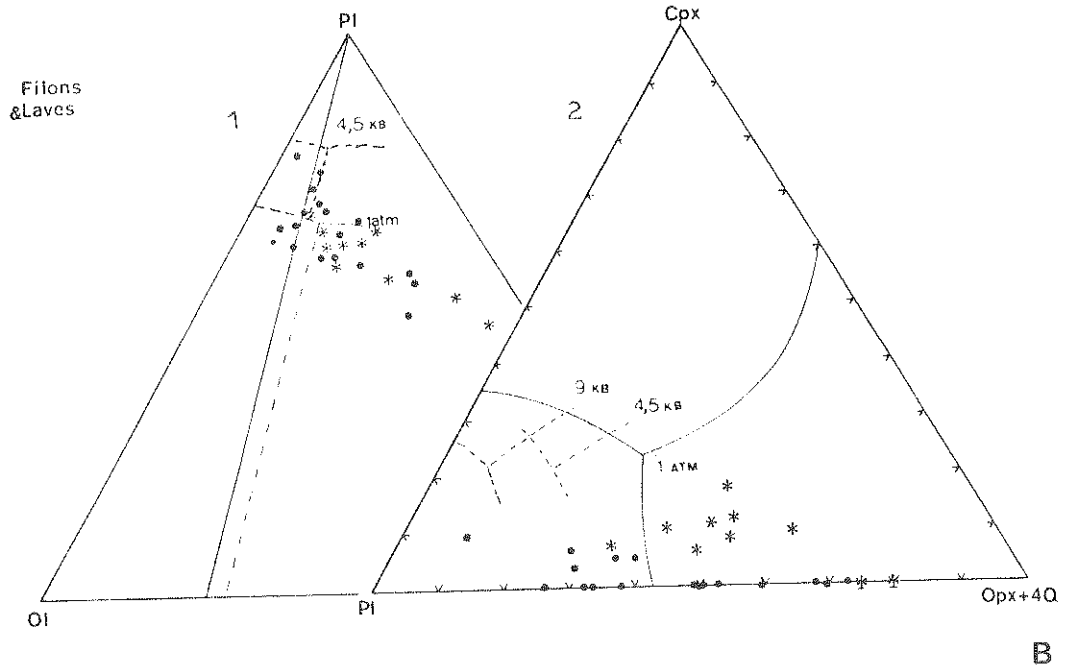
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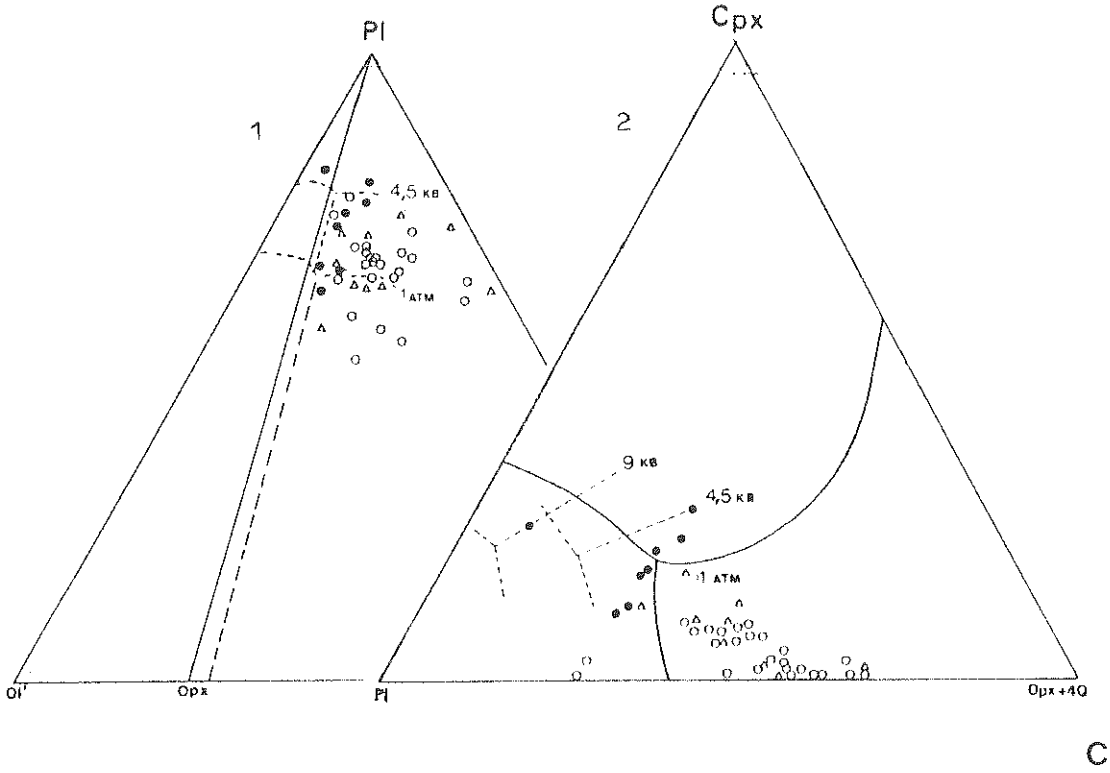


Fig. III.6 - Projection du tétraèdre Ol-Cpx-Pl-Q à partir du clinopyroxène (1) et de l'olivine (2) sur les plans Ol-Pl-Q et Pl-Cpx-Q : A pour les séries de Corse (tiré de OHNENSTETTER et OHNENSTETTER, 1976), B, du Vourinos, tiré de PAUPY, 1976 et C de Sabzevar, tiré de NOGHREYAN (1982).A : ronds,roches effusives ; triangle , plutonites. B : Etoiles à six branches, série d'Asprokambo ; autre , laves et sills de Krapa. C : rond blanc, complexe filonien de Baghdjar ; rond noir, filons recoupant les gabbros ; triangle, filons sécant sur les plagiogranites.

péritectique pour rejoindre l'eutectique Opx-Cpx-Pl-Q. Comme les liquides primaires doivent être en équilibre avec l'olivine et l'orthopyroxène, la position des liquides observés souligne l'importance du fractionnement, en particulier de l'olivine entre la zone de fusion et la surface. En considérant les projections à partir du clinopyroxène, la rareté des liquides dans le volume primaire du plagioclase tendrait à indiquer que ce sont les pyroxènes qui fractionnent après l'olivine, soit l'orthopyroxène, soit le clinopyroxène en fonction de la sursaturation du magma parental. Par ailleurs, le rapport CaO/Al_2O_3 varie selon les différentes pulsions de magma.

D'autres observations permettront de constater l'agencement des minéraux, la proportion des phases interstitielles de basses températures, les répétitions cycliques afin de juger de l'état du système, ouvert ou fermé, et des conditions d'équilibre entre minéraux et liquides. Toutes ces données, jointes aux conditions

thermométriques, permettent de préciser le bilan chimique du fractionnement. Par ailleurs, les travaux portant sur le gisement des cumulats, la taille des chambres magmatiques, l'importance des phénomènes gravitaires et des courants comparée aux cristallisations statiques à proximité des parois donnent des renseignements sur les processus physiques affectant les magmas dans les zones à faibles pressions qui doivent être compatibles avec les données chimiques.

On pourra se reporter à la première partie pour rechercher la nature des relations des cumulats entre eux et avec les autres membres de l'ophiolite. Dans le cas de la Corse, la nature des cumulats et le style d'accumulation décrits antérieurement (OHNENSTETTER et al., 1975) ont été repris et précisés plus récemment (OHNENSTETTER et al., 1979; III.1). D'autres travaux montrent que dans la série de Rospigliani, la chambre en fin d'évolution a pu fonctionner sous une faible carapace de laves : 100 à 150 m (OHNENSTETTER M., 1979; I.5 et 1ère partie). Les rubanements peu marqués, de même que l'abondance des cristallisations squelettiques constitueraient des arguments en faveur de chambres petites (1 km) et multiples, localisées au voisinage de la ride. Dans le cas des ophiolites de type II, les différents ordres de cristallisation observés sont en accord avec une variation de la composition des magmas avec le temps, ce qui se traduit aussi par une multitude de chambres magmatiques juxtaposées, quelquefois emboîtées. La morphologie de ces chambres, soulignée par le rubanement des cumulats apparaît guidée par le jeu des accidents majeurs qui dépendent eux-mêmes des conditions d'ouverture (OHNENSTETTER, 1982, I.3, et 1ère partie). Quelques vastes chambres existeraient dans certains massifs ophiolitiques tels en Oman (PALLISTER et HOPSON, 1981) qui appartiendraient à un type structural particulier (OHNENSTETTER M., 1980, I.2).

IV - LE CHEMINEMENT A TRAVERS LES ARTICLES REGROUPES

Une bonne image des séquences ophiolitiques de type I et II (Corse et Vourinos), en particulier de leurs cumulats sera obtenue en lisant le premier article (OHNENSTETTER et al., 1979, III.1) qui souligne les divergences quant à l'évolution chimico-minéralogique des cumulats et leur rôle possible sur la nature des fractionnements successifs. En tenant compte du comportement de MgO et CaO dans les deux suites cumulatives (fig. 5 de cet article) il a été supposé que l'évolution des liquides était respectivement caractéristique de cristallisa-

tion à l'équilibre (Corse) et de cristallisation fractionnée plus ou moins parfaite (Vourinos). Les fractionnements sont accompagnés, uniquement dans le cas du Vourinos, de dépôt de chromite. Dans cette note, lors d'une compilation sur la métallogénie des ophiolites, les paragenèses sulfurées sous-soufrées de Corse ont été opposées aux concentrations plus abondantes des amas sulfurés sur-soufrés des ophiolites de type II.

Les deux articles suivants (OHNENSTETTER et OHNENSTETTER, 1980 ; III.2 ; OHNENSTETTER et al., 1981 ; III.3) portent sur la partie supérieure des cumulats de Corse, en particulier les différenciats acides bien développés dans la série de Rospigliani (OHNENSTETTER M., I.5). Ces roches montrent une grande analogie avec celles des océans qui dériveraient pour la plupart d'un processus d'immiscibilité (OHNENSTETTER et al., 1979 ; III.1). Les enrichissements extrêmes en Zr, Y et TR légères de certaines albitites seraient liés à la conjonction de plusieurs facteurs : cristallisation fractionnée, immiscibilité auquel s'ajoute le rôle d'une phase vapeur riche en Na qui contrôlerait la distribution des éléments traces comme cela a été proposé pour des granites peralkalins.(TAYLOR et al. 1981). Il est à noter que des roches analogues ont été décrites dans des terrains

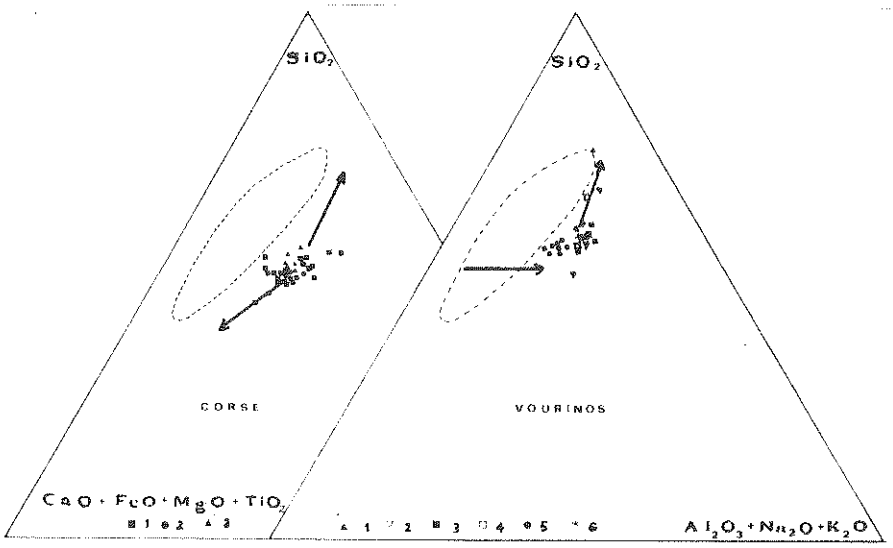


FIG. 7 — Diagramme SiO_2, $Al_2O_3 + Na_2O + K_2O$, CaO + FeO total + MgO + TiO_2 pour les liquides.
En pointillé domaine d'immiscibilité de deux liquides d'après J. W. GREIG (1927) et E. ROEDDER (1951).
Corse : 1 laves, 2 dolérites, 3 filons diabasiques. Vourinos : 1 dacites, 2 sills, 3 = laves massives, 4 = laves en coussins, 5 basaltes, 6 filons de basaltes, d'andésites basaltiques et de dacites.

Fig. III.7 - Diagramme de Greig (1927) soulignant la possibilité d'immiscibilité en fin de certaines différenciations ophiolitiques. Les flèches indiquent l'évolution de la composition des cumulats, d'après OHNENSTETTER et al. (1979 ; III.1).

plus anciens (MENOT, 1981 ; PIBOULE et al., 1982). Dans la seconde publication

qui fait aussi le point sur la nomenclature des roches acides des ophiolites, les textures des différenciats sont considérées en fonction de leur gisement ; les indications obtenues ont été alors employées pour proposer un fonctionnement possible des hauts de chambre à proximité d'une zone transformante. La troisième publication relative à la Corse renferme des données sur la typologie et l'âge (161 ± 3 MA) des zircons de même que sur la nature des inclusions qu'ils contiennent. Ces dernières de composition granitique témoignent du fractionnement du potassium, au moins dans certains magmas ophiolitiques, et océaniques, (fig.III.8) et de sa libération probable dans une phase vapeur (SINTON et BYERLY, 1980). La typologie des zircons correspond dans le diagramme employé à de hautes températures de cristallisation et à un indice d'agpaïcité fort qui sont en accord avec les processus magmatiques envisagés plus haut.

	A	B	C	D	E
SiO_2	32.54	32.40	75.85	71.9	69.68
Al_2O_3		-	15.15	16.3	12.41
FeO^{total}		0.18	0.06	0.59	5.09
MnO		0.09	0.03	n.d.	n.d.
MgO		0.01	-	0.21	0.40
CaO		-	0.63	2.92	2.85
Na_2O		-	6.30	6.94	4.29
K_2O		-	1.07	0.28	1.47
TiO_2		-	0.01	0.06	0.61
P_2O_5	0.28	0.20	n.d.		0.07
ZrO_2	67.30	67.30	n.d.		
Total	100.12	100.18	99.10	99.20	96.87

Fig.III.8 - Compositions des inclusions granitiques dans les zircons de Corse tiré de OHNENSTETTER et al. (1981 ; III.3) . A et B, analyses de zircon ; C, analyse d'une inclusion ; D, tache granophyrique, Océan Atlantique, tiré de SINTON et BYERLY (1980) ; E, rhyodacite, Galapagos, tiré de BYERLY et al. (1976).

Les trois derniers articles traitent du fractionnement dans les chambres magmatiques de Corse et du Vourinos dont la quantification a été abordée par les méthodes signalées précédemment, employées ensemble ou séparément (BECCALUVA et al., 1977 ; III.4 ; BECCALUVA et al., III.6 ; OHNENSTETTER M., III.5).

En Corse, l'emploi du fractionnement de Raleigh a conduit à la reconnaissance de trois stades majeurs dans l'évolution des liquides qui est compatible avec la cristallisation de troctolite (Ol-Pl-Sp) pour $0.85 < F < 1$ (F étant la fraction de liquide résiduel) puis d'euphotide (Pl-Cpx$\pm$Ol) pour $0.85 < F < 0.45$

F = FRACTION DE LIQUIDE

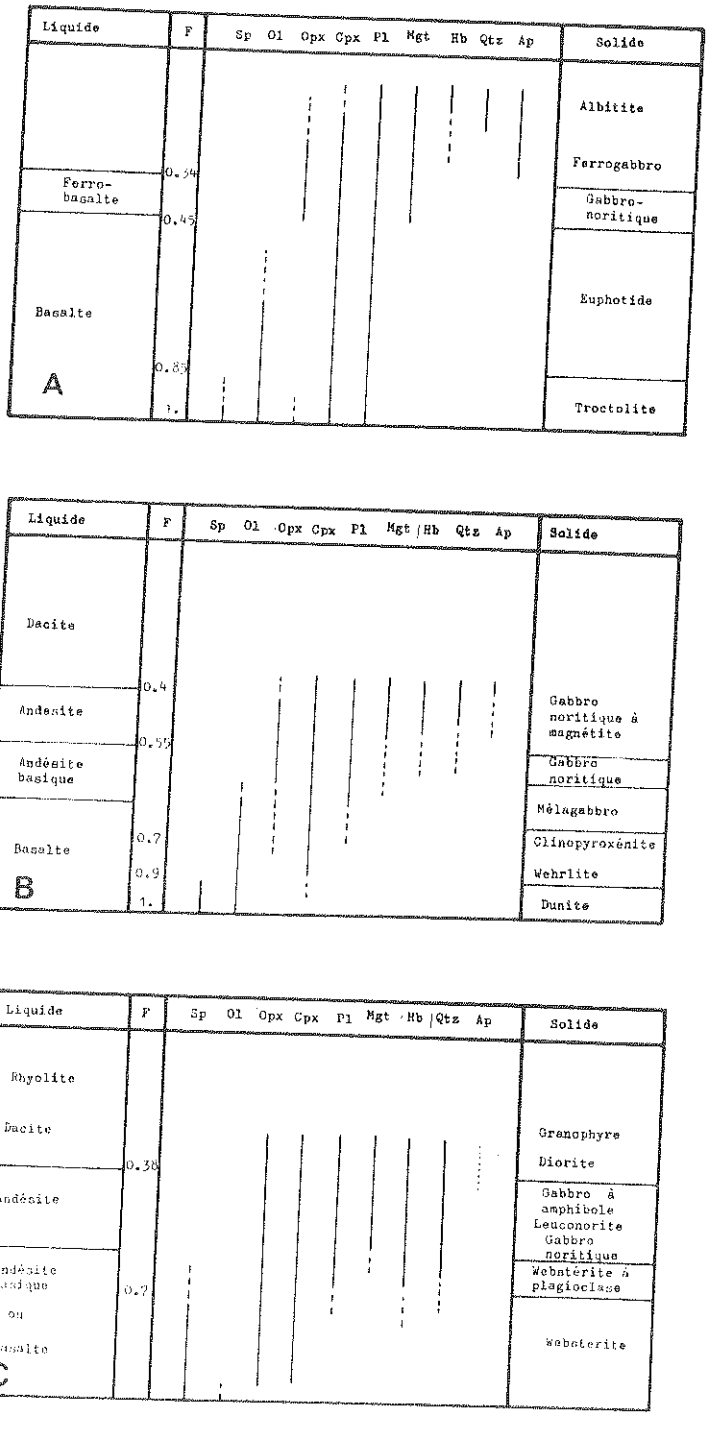


Fig. III.9 - Nature des liquides et cumulats associés en fonction de la fraction de liquide résiduel calculé selon la loi de Raleigh. A, Corse ; B, Krapa ; C, Asprokambo. L'évolution de Corse aurait lieu dans des chambres petites, fermées avec des cristallisations proches de l'équilibre. Pour les caractères des séries de Krapa et d'Asprokambo, se reporter à la légende de la Fig.III.5.

pendant lequel la tendance tholéitique typique se forme, et enfin de ferrogabbro $0.45 < F < 0.35$ entraînant la diminution de Ti, Fe et V dans les liquides. Les ferrogabbros riches en apatite et zircon et les leucoferrodiorites dérivent d'une fraction de liquide $F < 0.35$.

Le manuscrit sur le Vourinos met en évidence l'existence de deux séquences magmatiques ayant chacune un cortège de cumulats et de liquides associés qui ont été définis en couplant des études minéralogiques et géochimiques. Dans la série principale ou série de Krapa, la plus ancienne, la cristallisation de wehrlite, clinopyroxenite et mélagabbro pour $0.9 < F < 0.7$ a lieu à partir de basaltes. Les gabbros noritiques sont en équilibre avec des basaltes andésitiques ($0.7 < F < 0.4$) tandis que la précipitation des opaques donnant des gabbros noritiques riches en magnétite a lieu au niveau des andésites. Ces indications jointes aux données thermométriques et à la nature des textures indiquent que la chambre de Krapa était vaste, ouverte au départ et que les cristallisations ont eu lieu à l'équilibre. Progressivement, la taille de la chambre s'est réduite tandis que le nombre et le volume des apports régressaient conduisant à une cristallisation fractionnée plus parfaite. Au contraire les chambres suivantes de type Asprokambo étaient plus petites, probablement juxtaposées et fermées. Les textures et évolutions chimiques sont compatibles avec une cristallisation fractionnée plus parfaite. Des webstérite , webstérite plagifère, gabbro norite, leuconorite plus ou moins riche en amphiboles sont responsables de l'évolution des liquides depuis des basaltes ou basaltes andésitiques jusqu'à des rhyolites. Une fois définis les fractionnements des deux principales séries, des hypothèses quant aux conditions de fusion ont pu être avancées ; elles auraient eu lieu au voisinage d'une zone de subduction à partir d'un manteau appauvri de type océanique qui s'appauvrirait au fur et à mesure des fusions successives pour laisser après le magmatisme d'Asprokambo un résidu dunitique à harzburgitique.

Enfin, les bilans de masse effectués à partir des suites cumulatives de Krapa, Asprokambo et de Corse ont permis de préciser le degré et la nature du fractionnement dans les zones de basses et hautes pressions.

Les magmas parentaux calculés sont différents dans les trois séries selon que l'on se base sur les éléments majeurs ou traces. Dans le cas de la Corse comme dans celui des océans, cette étude a montré que le fractionnement de Opx, Sp et Ol est possible à haute pression (15- Kb) antérieurement à l'extraction de Ol-Pl et Cpx dans les chambres magmatiques à faibles profondeurs. L'extrusion des li-

COMPARAISON DES MAGMAS PARENTAUX

	Krapa L2	Aspro. L4	Aspro L'4	Troodos 4	Tortuga NT23	Boninite C2	Corse
SiO_2	50.66	53.04	54.37	52.40	47.25	54.19	48.49
TiO_2	0.30	0.13	0.15	0.30	0.79	0.18	1.05
Al_2O_3	10.73	9.19	7.64	11.7	13.64	9.57	15.35
FeO	7.39	7.90	8.01	8.4	9.77	8.40	6.33
MgO	14.41	15.04	15.35	15.80	17.61	12.79	11.44
CaO	9.27	9.56	9.19	10.7	9.58	8.54	7.96

Fig. III.10 - Composition des magmas parentaux de Krapa, Asprokambo et Corse en équilibre avec une olivine mantellaire (Fo_{90}). L4 et L'4 ont été calculés respectivement à partir d'un basalte et d'une andésite basique. La composition des magmas parentaux du Troodos, de Tortuga sont reportés d'après DUNCAN et GREEN (1980) et ELTHON (1979). La composition de la boninite est tirée de KOMATSU (1980).

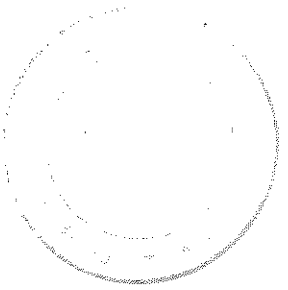
quides observés pourrait se faire rapidement après l'entrée du magma dans la chambre. Dans les ophiolites de Méditerranée orientale, les basaltes qui sortent de la chambre magmatique de type Krapa peuvent dériver de 40 % de fractionnement réalisé à basses pressions qui engagent surtout Ol+Cpx. Les calculs effectués sur les laves d'Asprokambo soulignent la diversité du magmatisme en fin d'évolution de l'ophiolite et l'importance du fractionnement pyroxénitique. Comparés au liquide de Krapa, ils sont plus sursaturés en SiO_2 mais plus pauvres en Al_2O_3 et autres éléments fusibles et incompatibles. En conséquence , certains magmas parentaux peuvent être constitués d'andésites magnésiennes se rapprochant alors de la composition des boninites.

Ces bilans sur les trois séries choisies soulignent que les fractionnements de basses et hautes pressions ont été accomplis à partir de magmas primaires de compositions différentes. Ceci implique que les processus de fusion à l'origine des magmas ophiolitiques soient également distincts, une hypothèse en accord avec le comportement des incompatibles et la nature et la minéralogie du résidu.

V - CONCLUSION

Dans cette troisième partie, il a été démontré qu'au sein des principaux assemblages ophiolitiques tels que ceux de Corse et du Vourinos, un lien magmatique unit les laves aux cumulats. Les calculs sur le fractionnement montrent, en particulier dans le cas du magmatisme de Krapa (Vourinos) que les liquides ne viennent en surface qu'après 40 % d'extraction de cumulats principalement de roches ultramafiques. En Corse, des fractionnements de basses et hautes pressions existeraient. La confrontation des compositions des magmas parentaux en équilibre avec le manteau calculées à partir des deux principaux types d'ophiolites, témoigne des divergences pétrologiques entre les sutures ophiolitiques. En ce qui concerne les ophiolites de Méditerranée orientale, comme le Vourinos, la composition des magmas évoluerait avec le temps au sein d'une même suture.

Ces études qui précisent les modalités de fractionnement rendent bien compte des discontinuités observées entre les sutures et au sein de celles-ci. (1ère et 2ème partie). En cela, elles justifient des classifications avancées et permettent de proposer des modèles évolutifs propres à chaque suture (2ème partie et chapitre introductif) selon que le magmatisme évolue avec le temps (ophiolites du type II) ou qu'un certain état d'équilibre est atteint. Le rôle du manteau et de l'asthénosphère sur la nature et le comportement des liquides sera envisagé lors de la montée des diapirs dans une zone d'accrétion au cours de la 4ème partie.



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in Facteurs contrôlant les minéralisations sulfurées de nickel

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IV — La diversité des ophiolites : importance de la nature du fractionnement et conséquences métallogéniques,

par D. OHNENSTETTER, M. OHNENSTETTER, A. PAUPY et G. ROCCI *

Résumé

Les séquences ophiolitiques de Corse et du Vourinos, respectivement caractéristiques des types I et II de la classification proposée pour les ophiolites de Méditerranée sont comparées. Une attention particulière est portée au rôle joué par les cumulats. Il apparaît que le comportement des liquides en surface est conditionné par la succession des cumulats et aussi par le mode de fractionnement dans les zones de basses pressions.

Les roches grenues de Corse résultent de cristallisations proches de l'équilibre ce qui est en accord avec les textures d'adcumulats tandis que celles du Vourinos résultent de cristallisations fractionnées plus ou moins parfaites. Par ailleurs, en Corse, des phénomènes d'immiscibilité ont pu modifier la nature des liquides en fin d'évolution pour leur faire franchir le plan critique de sursaturation.

Ces études montrent également que les magmas arrivant dans les chambres magmatiques situées à faible profondeur présentent des caractères géochimiques distincts qui ne peuvent avoir été acquis que dans des zones de hautes pressions. En particulier, un modèle à plusieurs stades est nécessaire dans le cas du Vourinos pour expliquer les faibles concentrations d'éléments incompatibles, d'alcalins, de chrome et de nickel, dans des liquides sursaturés. Des environnements géotectoniques différents : ride médio-océanique, arc ou mer marginale caractérisés par des conditions thermodynamiques particulières sont à l'origine de la dualité des ophiolites.

La potentialité métallogénique des ophiolites est liée à leur classification. Dans les ophiolites de type I (Corse), il n'y a pas de gisement de chromite et les amas sulfurés sont sous-soufrés et de faible extension. Ces caractères stériles sont liés aux propriétés physico-chimiques du magma tholéïtique à olivine : forte activité de FeO, et faibles fugacités d'oxygène et de soufre. Dans les ophiolites de type II (Chypre), les gisements de chromite et les amas sulfurés sur-soufrés sont liés à l'existence d'un magma sursaturé en silice (tholéïte à quartz) présentant une faible activité de FeO, et de fortes fugacités de soufre et d'oxygène.

Étant donné l'étroite analogie entre les ophiolites de type I et les roches des rides médio-océaniques, la probabilité de trouver des gisements importants dans la croûte océanique est faible. A l'inverse, la métallogénie des ophiolites de type II s'apparenterait à celle d'un environnement d'arc insulaire.

Abstract

The ophiolites of Corsica and Vourinos, respectively representative of types I and II of the classification of the Mediterranean ophiolites are compared. A special attention has been brought on the cumulates. The evolution of the basalts may be explained by the cumulate sequence and the nature of the fractionation under near surface conditions.

In Corsica, equilibrium crystallization may be involved to explain major element variations and the concentration of incompatible elements in the residual liquids. In the opposite, for the Vourinos ophiolites, more perfect fractional crystallization may be envisaged. Moreover, in Corsica, immiscibility phenomenon may have occurred at the end of the magmatic evolution to permit the liquid to cross over the thermal barrier of the low pressure divide.

Besides these low pressure processes, some of the chemical features of the basaltic liquids have been got at higher pressures. For the Vourinos, several stages of magma formation are needed to explain the low concentration of incompatible elements, alkalis, nickel and chromium in a sursaturated magma. Different geotectonical environments, caracterized by peculiar thermodynamic conditions, may be envisaged for generating the two types of ophiolites.

The metallogeny of the ophiolite complexes may be easily related to the duality of ophiolites.

In the type I (Corsica), there is a lack of chromite deposit and the volcanogenic sulphides are sulfur poor. These characteristics are linked to the physico-chemical constraints of the olivine-tholeiite magma : high FeO activity, low oxygen and sulphur fugacities.

In the type II (Cyprus), chromitites and sulphides deposits of economic interest may appear in a silica saturated magma (quartz tholeiite) showing low iron activity and high oxygen and sulphur fugacities.

As type I ophiolites and Mid Ocean ridges exhibit a close resemblance, it appears that important ore deposits are probably lacking in the oceanic crust. In the opposite, the metallogeny of type II ophiolite may be generated in an island arc setting.

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Introduction

On considère volontiers que les ophiolites représentent des fragments de croûte océanique incorporés au continent lors de la collision des plaques. Cependant, en dépit des mobilisations secondaires, les divergences portant sur la distribution des éléments majeurs et traces ont été jugées suffisamment importantes pour proposer l'hypothèse de la création d'ophiolites dans d'autres contextes géotectoniques : arcs, bassins marginaux. Ainsi en est-il du Troodos qui a d'abord été considéré comme preuve évidente d'une zone d'accrétion médio-océanique (GASS I. G., 1968; MOORES E. M. et VINE J. F., 1971) et ensuite imaginé dans un contexte d'arc insulaire (MIYASHIRO A., 1973). Les variations dans la composition chimique des basaltes permettent de classer les ophiolites et de les situer dans des environnements géotectoniques particuliers (MIYASHIRO A., 1975; OHNENSTETTER D. et OHNENSTETTER M., 1975, 1976; PEARCE J. A., 1975).

En Méditerranée, il a été distingué deux types d'ophiolites dont l'un, le type I, ressemble étroitement aux rides médio-océaniques : Corse (1), Alpes et Apennins et l'autre, le type II, regroupe des ophiolites plus orientales telles Chypre ou le Vourinos (ROCCI G. et *al.*, 1975).

La dualité des ophiolites doit être expliquée par le concours d'un ou de plusieurs facteurs : variation dans les mécanismes de la cristallisation fractionnée à partir d'un même magma ou non, variations dans la profondeur et le pourcentage de fusion partielle à partir d'une source identique ou non. La géochimie des basaltes étant en grande partie contrôlée par la nature des solides associés (BECCALUVA L. et *al.*, 1977), les variations minéralogiques et géochimiques des cumulats seront en particulier étudiées pour définir le mode de fractionnement et son incidence sur le comportement des liquides résiduels (O'HARA M. J., 1977). Parmi les ophiolites, celles de Corse et du Vourinos, respectivement caractéristiques des types I et II ont été choisies pour mener à bien cette étude.

Cette dualité se retrouve au niveau des potentialités métallogéniques. Les gisements de chromite et les amas sulfurés sont importants dans les ophiolites de type II alors que le type I est relativement stérile.

Rappel de la constitution des deux séquences ophiolitiques

Les ophiolites de Corse et du Vourinos sont antétithoniques d'après l'âge des sédiments associés aux laves et d'après des datations géochronologiques. En Corse, dans la zone du Piémont des Alpes occidentales, les fragments des ophiolites (ROUTHIER P., 1946; NETELBEEK T. A. F., 1951; GUILLOU J. J. et *al.*, 1962; FRANCONI A., 1967; PETERLONGO J. M., 1968; OHNENSTETTER D. et OHNENSTETTER M., 1975) sont tectoniquement inclus dans les schistes lustrés. Ces écailles ont été diversement affectées par de nombreux événements métamorphiques que l'on peut rattacher successivement au métamorphisme océanique, à une zone de subduction et à l'orogenèse alpine (OHNENSTETTER D. et *al.*, 1976).

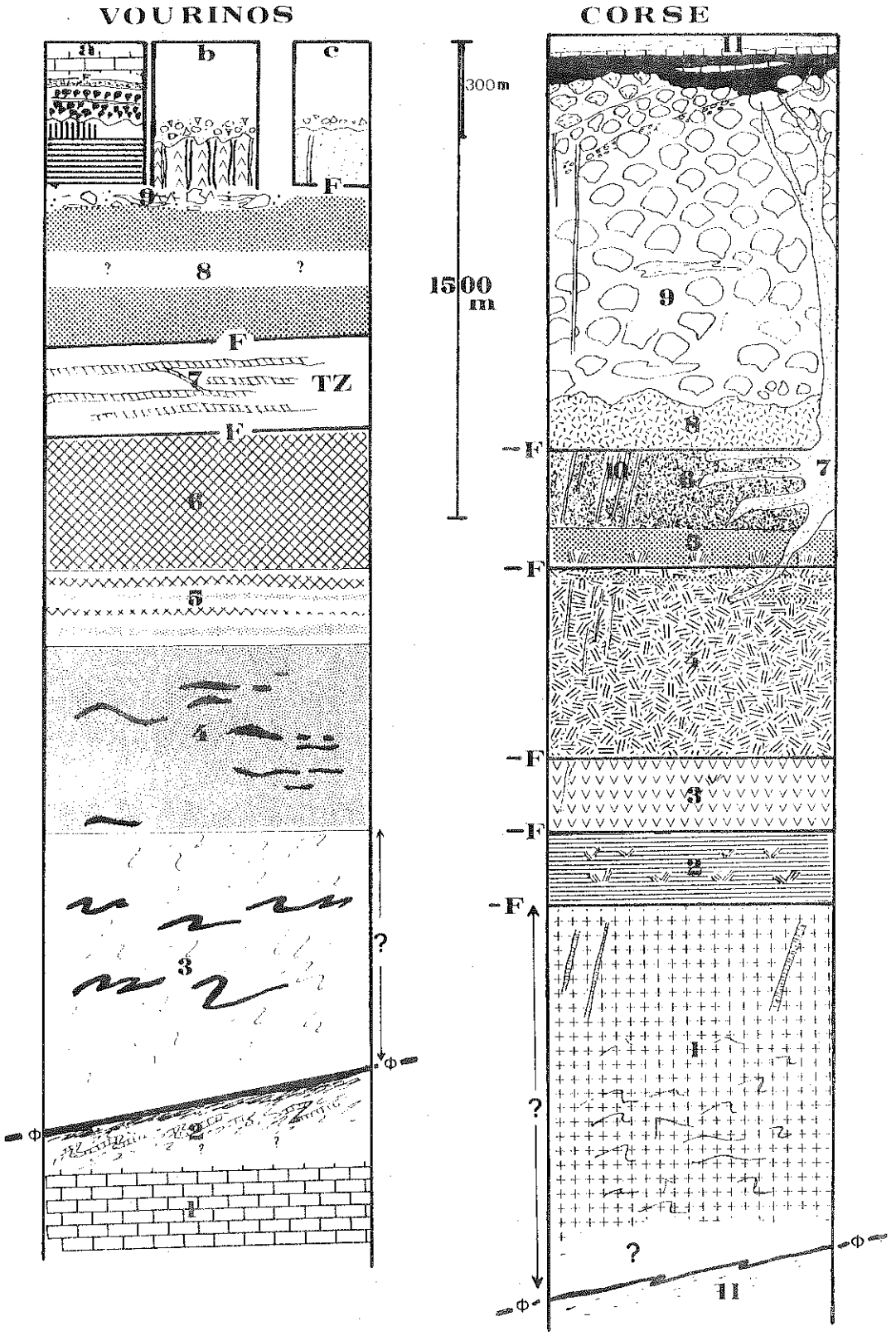
Dans le système hellénique – dinarique, le complexe ophiolitique du Vourinos (BRUNN J. H., 1955; MOORES E. M., 1969; PAUPY A., 1976) chevauche la bordure

(1) La séquence ophiolitique corse a été reconstituée à partir des massifs de l'Inzecca, de la Casaluna, de Farinole et de Monte Piano Maggiore (cf. OHNENSTETTER D. et M. 1975).

FIG. 1 Logs schématiques des ophiolites du Vourinos et de Corse.

Les successions de roches, de bas en haut, sont constituées dans le cas du Vourinos de : 1. Carbonates du Trias au Jurassique. 2. « Mélange » : serpentinites, schistes amphiboliques et niveaux marmoréens plissés; 3. Harzburgites « tectonites » et chromitites. 4. Dunites et chromitites. 5. Unités cycliques de dunites, wehrlites, clinopyroxénites avec ou sans olivine. 6. Unités cycliques de clinopyroxénites à olivine ou à plagioclase; 7. Zone de transition (TZ). Dykes de webstérites dans des dunites serpentinisées. 8. Gabbros inférieurs et supérieurs; 9. Brèches magmatiques de gabbrodiorites; 10. *a*) Colline de Krapa : sills accompagnés de quelques dykes, laves en coulées massives ou en coussins, radiolarites et calcaires éocrétacés; *b*) Asprokambo : complexe filonien d'Asprokambo intrusif dans des diorites et des diorites quartziques, conglomérat néogène; *c*) Plagiogranites de Mandania recouverts du conglomérat néogène.

Dans le cas de la Corse : 1. Péridotites évoluant de harzburgites à des lherzolites à spinelle et/ou à plagioclase. Ces péridotites sont parfois faiblement tectonisées. Des filons de gabbros, souvent rodinginitisés, sont présents dans l'unité ultramafique; 2. Zone de transition composée de troctolites, d'allivalites, d'anorthosites, de gabbros à olivine et de péridotites plagifères; 3. Allivalites et troctolites. 4. Euphotides. 5. Gabbros intermédiaires. 6. Ferrogabbros. 7. Albitites; 8. Dolérites massives; 9. Laves en coussins variolitiques, aphanitiques, porphyriques associées à des hyaloclastites et à des brèches volcaniques; 10. Filons diabasiques. 11. Série sédimentaire de l'Inzecca débutant par des radiolarites.


VOURINOS
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CORSE
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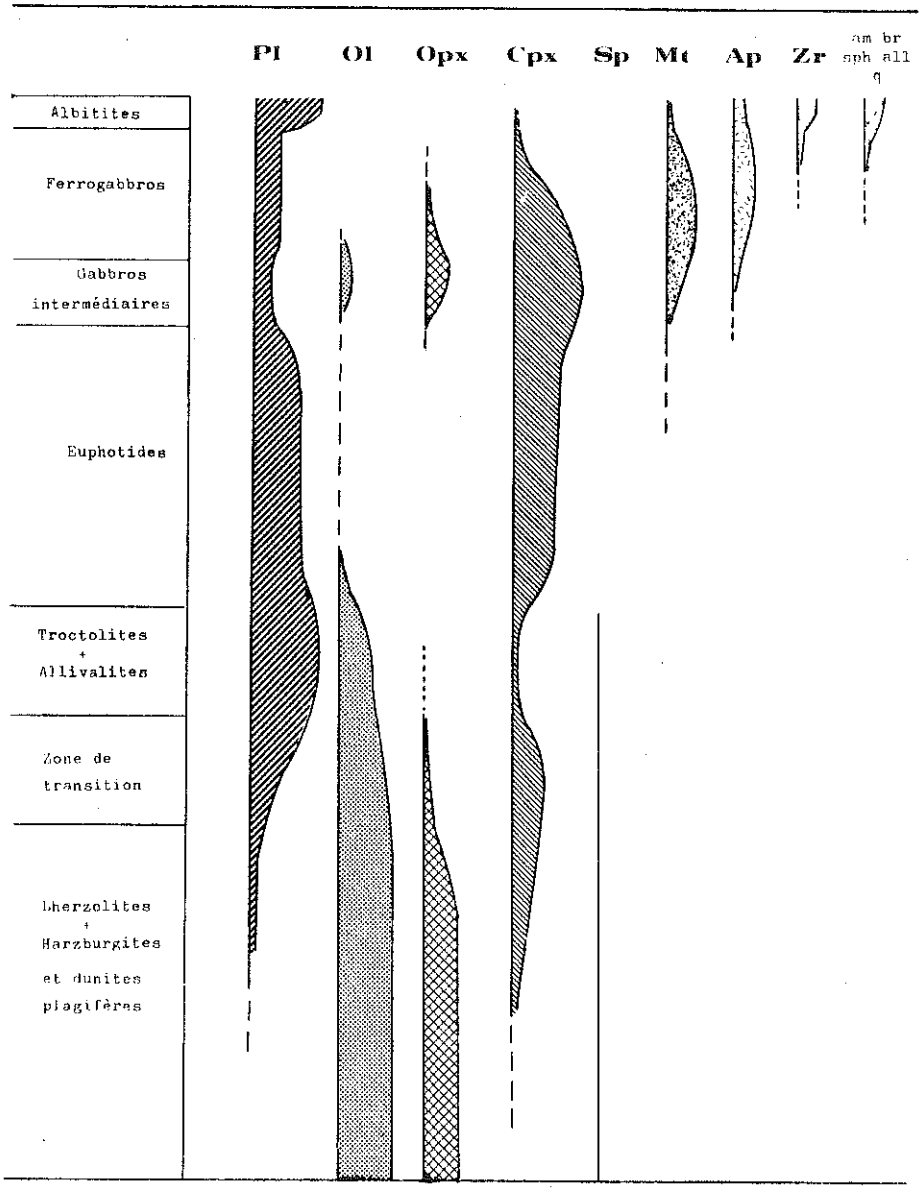

FIG. 2 Ordre de cristallisation des cumulats de Corse.

Abréviations des principaux minéraux : pl. : plagioclases, ol. : olivines, opx. : orthopyroxènes, cpx : clinopyroxènes, sp : spinelles, cr : chromites, mt : magnétites, am : amphiboles, am.br. : amphiboles brunes, ap. : apatites, bi. : biotites, q. : quartz, sph. : sphènes, all : allanites.

occidentale de la zone pélagonienne. Les termes de base, c'est-à-dire les péridotites, reposent sur des amphibolites déformées et sur un « mélange » tectonique. Un métamorphisme de faciès schiste vert affecte ces ophiolites.

Ces deux complexes correspondent à la définition de STEINMANN en cela qu'ils comprennent de bas en haut des péridotites, des gabbros et des laves. Néanmoins si l'on juxtapose les deux logs lithologiques reconstitués, des différences apparaissent dans la composition des principaux termes (fig. 1).

1 — Les péridotites

Elles offrent des caractères profondément différents quant à leur nature et à leur déformation. Si l'on se réfère à la composition de ces roches on peut classer les péridotites de Corse et celles du Vourinos respectivement dans les sous-types lherzolitiques et harzburgitiques de T. P. THAYER (1972).

Dans le Vourinos, comme c'est le cas dans la plupart des

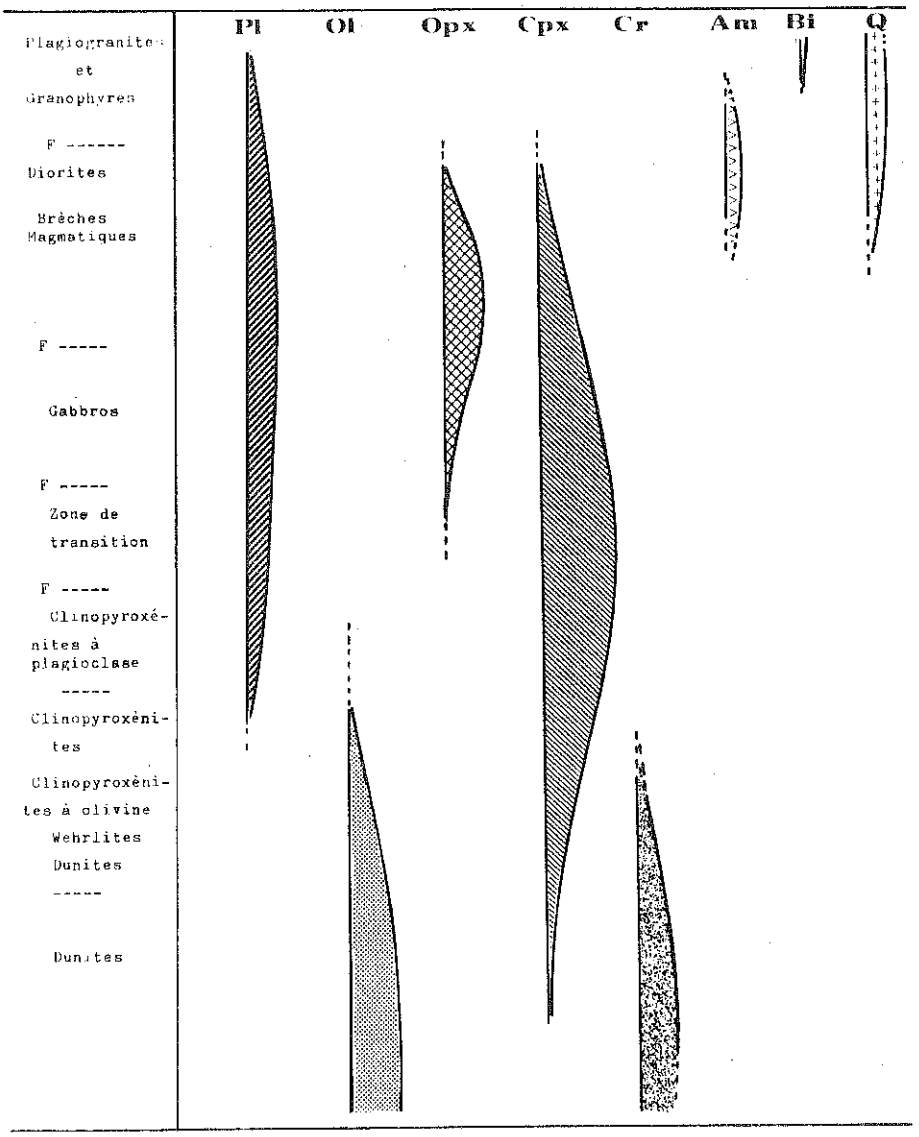

FIG. 3. — Ordre de cristallisation des cumulats du Vourinos.
Abréviations des principaux minéraux : (voir fig. 2).

ophiolites, les harzburgites possèdent des textures protoclastiques caractérisées par une forte fabrique (MOORES E. M., 1969). Ces harzburgites et les roches qui les accompagnent : dunites et chromitites, constitueraient le matériel réfractaire issu de la fusion d'un manteau pyrolitique. Une section composite de cumulats péridotitiques les recouvre. Dans les 24 unités cycliques reconnues (JACKSON D. E. et *al.*, 1975), le plan du rubanement et de la lamination ignée des olivines est parallèle au plan de foliation des tectonites ou forme un angle très faible avec celui-ci.

En Corse, les péridotites essentiellement lherzolitiques sont alumineuses. Des variations graduelles dans la proportion du clinopyroxène rendent compte de la transition entre harzburgite et lherzolite. Spinelle et plagioclase sont souvent présents ensemble ou séparément. Quelques péridotites possèdent des textures porphyroclastiques et granuloblastiques caractéristiques des tectonites. Péridotites à texture de cumulats et péridotites tectonisées qui sont étroitement associées sur le terrain, comme dans le massif de Monte Maggiore, ont fait suggérer que les secondes étaient issues de la déformation des premières et qu'il

s'agissait donc de cumulats déformés. Des péridotites plagifères composées essentiellement d'olivine, de plagioclase et de clinopyroxène en proportion variable, ce qui confère à la roche un excellent rubanement, assurent la transition vers les roches grenues gabbroïques.

2 — Les roches grenues basiques

En Corse, la série de roches basiques commence par des troctolites riches en plagioclase dans lesquelles le clinopyroxène est accessoire. De la variation dans la proportion des deux principaux constituants : l'olivine et le plagioclase résultent des niveaux dunitiques et anorthositiques. Lorsque le clinopyroxène (diallage) augmente aux dépens de l'olivine, on obtient des gabbros à olivine et des euphotides (gabbro *S. S.*) qui sont les plus répandus dans la chaîne alpine. Les gabbros intermédiaires où apparaissent l'orthopyroxène et la titanomagnétite assurent la transition entre euphotide et ferrogabbro. L'augite est souvent branchue dans ces roches riches en pyroxène. Les ferrogabbros possèdent, en plus, apatite, sulfures et hornblende brune. Les résidus acides qui recoupent les gabbros supérieurs et les laves sont composés de plagioclases acides zonés, de pyroxène ou d'amphibole sodiques issus de la transformation des pyroxènes et amphiboles magmatiques, de zircon, de quartz et d'allanite.

Dans le Vourinos, les pyroxénites, souvent des clinopyroxénites, constituent la transition entre les péridotites et les gabbros. Dans ces derniers, essentiellement noritiques, le clinopyroxène en intercumulus est transformé en hornblende verte. Dans la zone de brèches magmatiques qui séparent les gabbros des diorites, le quartz et les hornblendes primaires apparaissent tandis que l'orthopyroxène disparaît. Le plagioclase y est en cumulus et en intercumulus. Les différentiats acides, importants en volume, sont composés de diorites quartziques et de granophyres. Dans les diorites, l'interpénétration de quartz et feldspath caractéristique de la texture granophyrique est visible en position interstitielle.

Dans chaque séquence, on peut déduire de la succession des cumulats un ordre de cristallisation (fig. 2 et 3).

En Corse, (fig. 2) les fractionnements du plagioclase et de l'olivine sont très importants à la base des gabbros. Celui du clinopyroxène, d'abord accessoire, devient prépondérant lorsque l'olivine diminue. Le fractionnement des deux pyroxènes atteint un maximum dans les gabbros intermédiaires auquel s'ajoutent dans les ferrogabbros, les fractionnements des apatites et des titanomagnétites. Le quartz est très peu exprimé même dans les différentiats acides.

Bien que dans le Vourinos (fig. 3) il y ait à la base des unités cycliques, on peut néanmoins déduire approximativement le domaine de stabilité des principales phases en cumulus. Olivine et chromite sont les minéraux importants dans les péridotites à texture de cumulat. Dans la zone de transition, les pyroxènes, en particulier le clinopyroxène, dominent tandis que le plagioclase rare apparaît en intercumulus. L'orthopyroxène prépondérant dans les gabbros et à la base des diorites disparaît lorsque l'amphibole et le quartz précipitent. La proportion de quartz augmente ensuite régulièrement jusque dans les termes les plus évolués.

3 — Les termes effusifs

En Corse, les laves très épaisses sont constituées de basaltes en coussins le plus souvent variolitiques, aphyriques et porphyriques, dans lesquels on peut reconnaître des reliquats de plagioclase, d'olivine et de spinelle. Des dolérites massives apparemment intrusives dans les laves en coussins séparent ces dernières des cumulats. Les filons sont rares et présentent les mêmes caractéristiques pétrographiques et géochimiques que les dolérites massives.

Dans le Vourinos, les laves en coussins ou en coulées, les sills et les dykes varient des basaltes aux dacites en passant par des andésites. Les laves et les roches hypabyssales représentent des liquides issus de chambres magmatiques dans lesquelles se sont déposés les cumulats. Les différents stades de fractionnement peuvent être suivis par l'évolution géochimique des liquides ponctionnés et par les assemblages successifs des cumulats. L'évolution des liquides ophiolitiques se fait toujours soit dans le domaine des tholéites à olivine, soit dans le domaine des tholéites à quartz d'après le diagramme de H. S. YODER et C. E. TILLEY (fig. 4). Dans ce tétraèdre, tous les points sont déplacés vers le pôle sous-saturé en raison de la spilitisation. Néanmoins les points représentatifs des cumulats et des

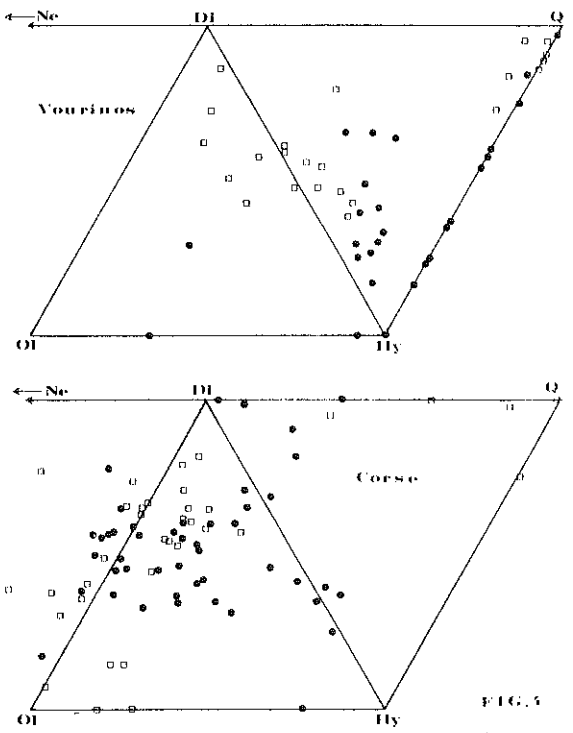

FIG. 4. — Projection d'après le diagramme de H. S. YODER et C. E. TILLEY (1962). Composition de roches exprimée en pourcentage de minéraux normatifs. Carrés : cumulats; points : liquides.

laves de Corse tombent dans le champ des tholéïtes à olivine à l'exception de quelques roches différenciées, ferrodolérites et albitites, qui sont dans le domaine des tholéïtes à quartz. Les liquides et les cumulats gabbroïques du Vourinos se placent essentiellement dans le champ des tholéïtes à quartz. Le magma du Vourinos, à ce stade de son évolution, présentait donc un caractère sursaturé en SiO_2 par rapport à celui de Corse.

Géochimie des cumulats

1 — Évolution chimico-minéralogique des cumulats

a) Variation des éléments

L'évolution des éléments majeurs dans les cumulats peut être suivie sur la figure 5. Certains de ces éléments ont un comportement voisin dans les deux ophiolites : Mg et Ca décroissent au cours de la différenciation dans les niveaux gabbroïques tandis que les alcalins et le titane augmentent. Au contraire la proportion de silice augmente régulièrement dans les cumulats du Vourinos tandis que la teneur en silice des gabbros de Corse est constante sauf au niveau des ferrogabbros où elle montre un minimum et dans les résidus acides où elle est maximum. La courbe de la silice du Vourinos, située pour toutes les valeurs de 100 × FeO/MgO + FeO au-dessus de celle de la Corse, souligne le caractère sursaturé du magma du Vourinos.

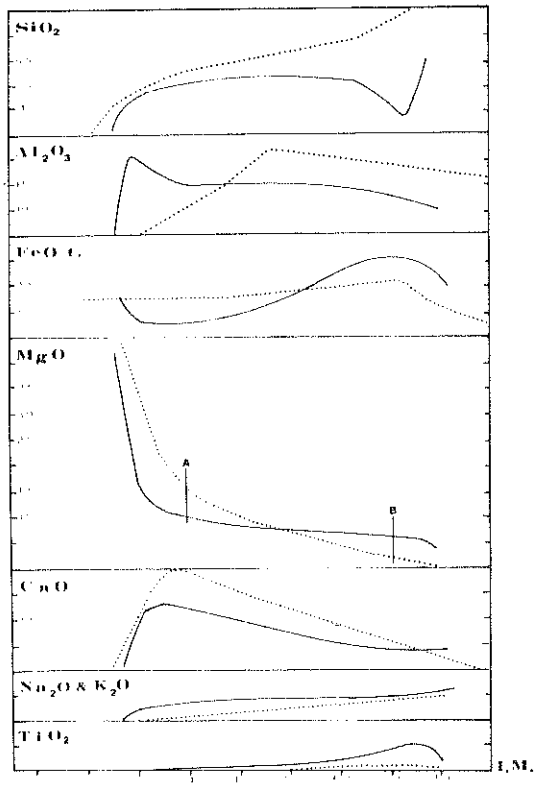

FIG 5. Variation des éléments majeurs en fonction de l'indice mafique. IM = (100 · FeO total/FeO total + MgO).
En pointillé : Vourinos; en trait plein continu : Corse. A et B voir le texte.

L'alumine augmente brutalement au niveau des troctolites de Corse (20% Al_2O_3), puis décroît régulièrement jusqu'en fin de différenciation. Dans le Vourinos, les variations de l'alumine, plus régulières, montrent un maximum à 100 FeO/FeO + MgO = 48 de telle sorte que les cumulats supérieurs apparaissent relativement enrichis en cet élément. La distribution du calcium dans le Vourinos montre un pic important pour 100 FeO/FeO + MgO = 25. La décroissance du calcium est ensuite régulière jusqu'en fin de différenciation. La courbe du calcium des gabbros de Corse, régulière sur sa plus grande partie, montre néanmoins un fléchissement au niveau des ferrogabbros. Le pourcentage de sodium augmente régulièrement dans les cumulats du Vourinos, de façon plus sensible après 100 FeO/FeO + MgO = 40 lorsque l'alumine commence à décroître. En Corse, les teneurs de sodium apparaissent sensiblement constantes sauf dans les résidus acides où les concentrations peuvent être importantes. Les positions respectives des courbes du calcium et des alcalins pour les deux massifs ou les surfaces délimitées par ces courbes indiquent que les cumulats du Vourinos sont plus riches en calcium et plus pauvres en alcalins que ceux de Corse.

Après les faibles teneurs en fer enregistrées dans les cumulats de base de Corse, l'enrichissement en cet élément au niveau des ferrogabbros est très prononcé et contraste avec l'augmentation régulière du fer, puis sa décroissance dans les gabbros du Vourinos. Dans la série corse, le titane, comme le fer, montre un net maximum au niveau des ferrogabbros. Les concentrations de titane dans le Vourinos sont toujours inférieures à celles de la Corse.

b) Contrôle minéralogique de l'évolution

Si l'on considère l'ordre de cristallisation et la proportion des phases en cumulus dans les deux séquences ophiolitiques ainsi que l'évolution des éléments majeurs, on constate que les divergences observées quant à l'amplitude et la position des maxima s'expliquent par la précipitation de certains minéraux. Ainsi dans la série corse, le pic de l'alumine au début de la différenciation est dû à la précipitation massive des plagioclases abondants dans les troctolites et les anorthosites. De même, le maximum de calcium dans les roches peu différenciées du Vourinos s'explique par la cristallisation importante des clinopyroxènes. Des teneurs plus importantes de calcium dans le magma du Vourinos, entraînant un rapport

$CaO/CaO + Al_2O_3$ plus élevé que dans celui de Corse auraient favorisé la cristallisation du clinopyroxène aux dépens du plagioclase dans le Vourinos (IRVINE T. N., 1970).

Le comportement de ces deux éléments souligne les caractères respectifs des zones de transition troctolitique de Corse et clinopyroxénitique du Vourinos. La partie ascendante de la courbe de l'alumine dans le Vourinos peut être reliée à l'augmentation des plagioclases dans les cumulats tandis que la partie descendante serait due au fractionnement des plagioclases de plus en plus sodiques comme la répartition du sodium semble l'indiquer. En Corse, l'augmentation en fer et en titane couplée avec la décroissance de la silice au niveau des ferrogabbros est due à la précipitation massive des minéraux opaques. Dans le massif du Vourinos, l'augmentation sensible de Na et K dans les granophyres (non visible sur ce diagramme) peut être liée à la cristallisation des amphiboles (CAWTHORN R. G., 1976). Celles-ci n'apparaissent plus dès que les teneurs de calcium sont inférieures à 2% (CAWTHORN R. G., *op. cit.*).

c) Caractérisation des deux évolutions à l'aide du diagramme de M. BESSON et M. FONTEILLES (1974).

L'importance de la nature, de la proportion et de l'association des phases qui précipitent sur le comportement des éléments majeurs peut être mieux saisie dans le triangle de M. BESSON et M. FONTEILLES (1974) (fig. 6) où ont été reportées des lignées tholéitiques et calco-alcalines de référence des cumulats gabbroïques. Le report des points des gabbros de Corse et du Vourinos se fait respectivement sur les lignées tholéitique et calco-alcaline. L'évolution vers des termes riches en fer avec un rapport Al_2O_3/MgO presque constant est visible dans le cas de la Corse tandis que l'évolution vers un pôle alumineux dans le cas du Vourinos y apparaît nettement. Dans ce massif, il existe une continuité entre les gabbros et diorites d'une part et les termes acides granophyriques d'autre part. A l'opposé, entre les ferrogabbros de Corse et les granites albitiques, il existe un vide. Sur ce diagramme où sont également reportées les péridotites des deux massifs, le caractère réfractaire des harzburgites du Vourinos et celui moins appauvri en CaO et Al_2O_3 des lherzolites de Corse apparaissent également.

2 — Nature du fractionnement

Sur la figure 5, lorsque les éléments majeurs évoluent dans le même sens, les pentes des courbes représentatives de la Corse et du Vourinos pour un même élément ne sont pas identiques. Les courbes de SiO_2, de MgO et dans une moindre mesure Al_2O_3 sont faiblement pentées pour les gabbros de Corse à l'opposé de celles du Vourinos. D'une

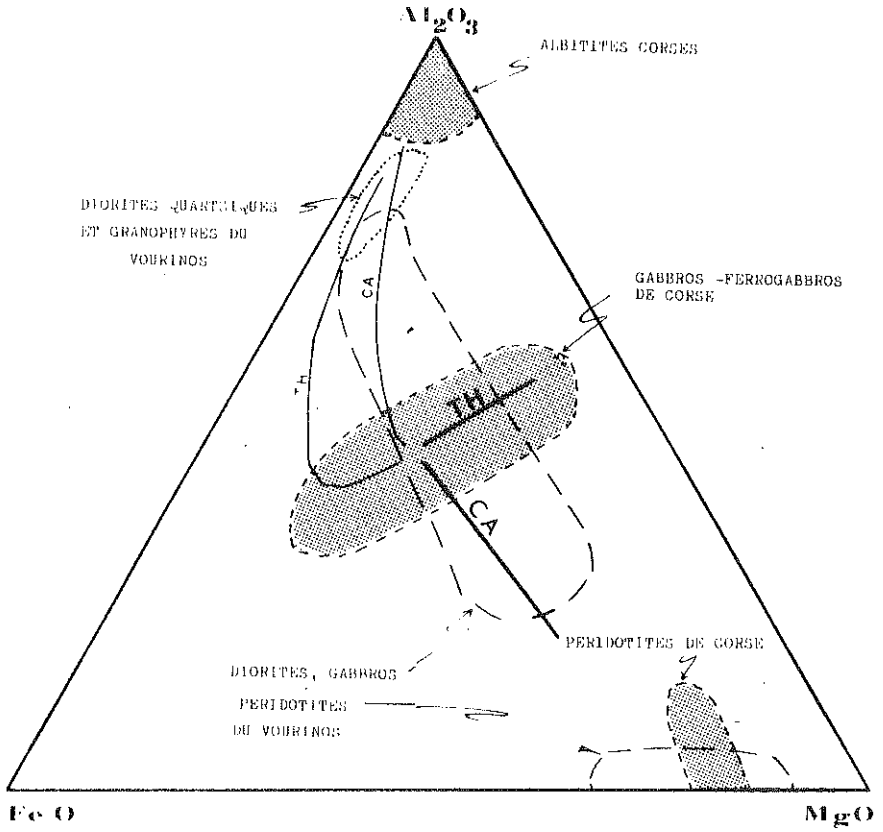

FIG 6. Diagramme Al_2O_3 FeO total MgO. CA = lignée calco-alcaline; TH = lignée tholéitique de référence.

manière plus générale, les courbes de variation du Vourinos, tout en étant plus régulières sont aussi plus pentées que celles de Corse, ce qui les conduit dans le cas de MgO et de CaO à atteindre l'axe des abscisses.

Considérons par exemple les courbes de variation de MgO aux points A et B. En A, les minéraux mafiques du Vourinos contiennent plus de MgO que ceux de Corse alors qu'en B la situation est inverse. Dans le Vourinos, tout se passe comme s'il y avait un retrait successif des minéraux formés pour appauvrir rapidement le liquide résiduel en MgO. Dès lors, en fin de différenciation B, les minéraux mafiques qui cristallisent sont pauvres en MgO pour être en équilibre avec un liquide lui-même appauvri en cet élément. Cette évolution est caractéristique d'une cristallisation fractionnée plus ou moins parfaite. Inversement, dans le cas de la Corse, tout se passe comme s'il y avait un réajustement constant de la composition des phases mafiques au contact du liquide, de telle sorte qu'en fin de différenciation B, la concentration de MgO dans le liquide soit relativement importante; la formation de ces cumulats présente donc des caractéristiques de cristallisation à l'équilibre.

Par ailleurs, la précipitation précoce du feldspath dans les cumulats de Corse ne contribue pas à appauvrir le liquide en MgO.

On peut de même considérer les courbes d'évolution du calcium. Celle du Vourinos, régulièrement pentée à partir de I M = 25 et atteignant l'axe des abscisses, résulte probablement d'un processus de cristallisation fractionnée. La courbe évolutive de Corse pour I M compris approximativement entre 25 et 55 possède une pente assez forte bien qu'inférieure à celle du Vourinos. Il existe une similitude d'évolution du calcium et du magnésium dans le cas du Vourinos que l'on ne retrouve pas en Corse. Pour I M compris entre 25 et 55, à des diminutions importantes de CaO ne correspondent pas des variations importantes de MgO. De ce fait, si l'on prolonge la partie pentée de la courbe du calcium, on devrait obtenir en fin de différenciation des minéraux mafiques relativement riches en MgO, coexistant avec des plagioclases riches en Na_2O. Tel n'est pas le cas car CaO ne diminuant que peu ou plus à partir de I M = 60, la pente devient très faible et la courbe recoupe alors celle du Vourinos (cf. paragraphe 3).

Lors de la différenciation magmatique, FeO/MgO varie de 0,2 à 2,5 dans les cumulats de Corse et les liquides correspondants tandis que I M varie de 17 à 80. Les éléments incompatibles sont fortement concentrés en fin de différenciation (OHNENSTETTER D. et *al.*, 1975; BECCALUVA L. et *al.*, 1977). A l'opposé, le rapport FeO/MgO varie de 0 à 5 dans les termes effusifs et les cumulats du Vourinos et I M est compris entre 10 et 90 (BEBIEN J. et *al.*, 1975; PAUPY A., 1977). Le titane est faiblement concentré en fin de différenciation. Par ailleurs, la variation de la composition des minéraux dans le cas de la Corse est plus lente que dans les cumulats du Vourinos. Selon M. J. O'HARA (1977), lors d'une cristallisation à l'équilibre, la concentration des éléments incompatibles dans les liquides se fait pour de faibles variations de FeO/MgO tandis qu'au cours d'une cristallisation fractionnée parfaite, le rapport FeO/MgO augmente rapidement et la concentration des éléments incompatibles dans les liquides résiduels reste faible et inférieure à celles obtenues par le mécanisme précédent. Les évolutions magmatiques de Corse et du Vourinos sont de ce point de vue respectivement caractéristiques de cristallisation à l'équilibre et de cristallisation fractionnée.

Il s'ensuit que les textures des cumulats et par suite les conditions de formation qu'elles traduisent doivent être en accord avec les processus magmatiques envisagés. En Corse, l'abondance des adcumulats qui impliquent l'existence de phénomène de diffusion entre solide et liquide pour modifier leur composition respective et éliminer les fluides interstitiels (OHNENSTETTER D. et *al.*, 1975) est en accord avec un mécanisme de cristallisation à l'équilibre. Les cumulats du Vourinos qui possèdent des minéraux zonés et des phases en position interstitielle sont plus caractéristiques d'un processus de cristallisation fractionnée. De plus, l'existence de courants importants et continus, en favorisant le dépôt des minéraux donc leur recouvrement rapide, a permis l'isolement des cumulats formés et le piégeage des liquides interstitiels. Au contraire, en Corse, l'absence de courant continu, en ralentissant la formation des cumulats, a augmenté la durée des échanges entre minéraux précipités et liquide.

3 — Immiscibilité en fin de certaines différenciations

Les liquides et les cumulats gabbroïques ont été reportés dans le triangle

$$SiO_2 — CaO + MgO + FeO + TiO_2 — Al_2O_3 + K_2O + Na_2O$$

(fig. 7 et 8), qui est utilisé pour étudier le comportement des liquides en fin de différenciation (WEIBLEN P. N. et ROEDDER E., 1973; NAKAMURA Y., 1974; McBIRNEY A. R. et NAKAMURA Y., 1974; McBIRNEY A. R., 1975).

Dans ce diagramme, les points représentatifs des roches de la Corse ne tombent jamais dans le champ des deux liquides immiscibles. La droite rejoignant les troctolites aux euphotides s'oriente vers le pôle alumineux du triangle et rejoint la courbe de séparation du domaine à deux liquides. Les basaltes suivent la même évolution, montrant un appauvrissement relatif en alumine au cours de la différenciation. Les ferrogabbros et les albitites se placent de part et d'autre de cette enveloppe.

A la différence des gabbros de Corse, quelques cumulats du Vourinos sont dans le champ de démixion. La courbe caractéristique de l'évolution de ces cumulats commence près de la droite $SiO_2 — CaO + FeO + MgO + TiO_2$, évolue à SiO_2 constant tandis qu'Al_2O_3 augmente en passant par le domaine d'immiscibilité. Une fois sortie de ce champ, elle change de direction pour se rapprocher du pôle riche en silice. Les liquides du Vourinos ne tombent pas dans ce domaine. La courbe de différenciation a cependant les mêmes caractères que celles des cumulats supérieurs associés.

L'évolution géochimique des cumulats et des liquides de Corse apparaît encore fort différente de celle du Vourinos et peut être modifiée en fin de différenciation par un phénomène d'immiscibilité. En effet, lorsque la courbe d'évolution du liquide recoupe le champ de démixion, on

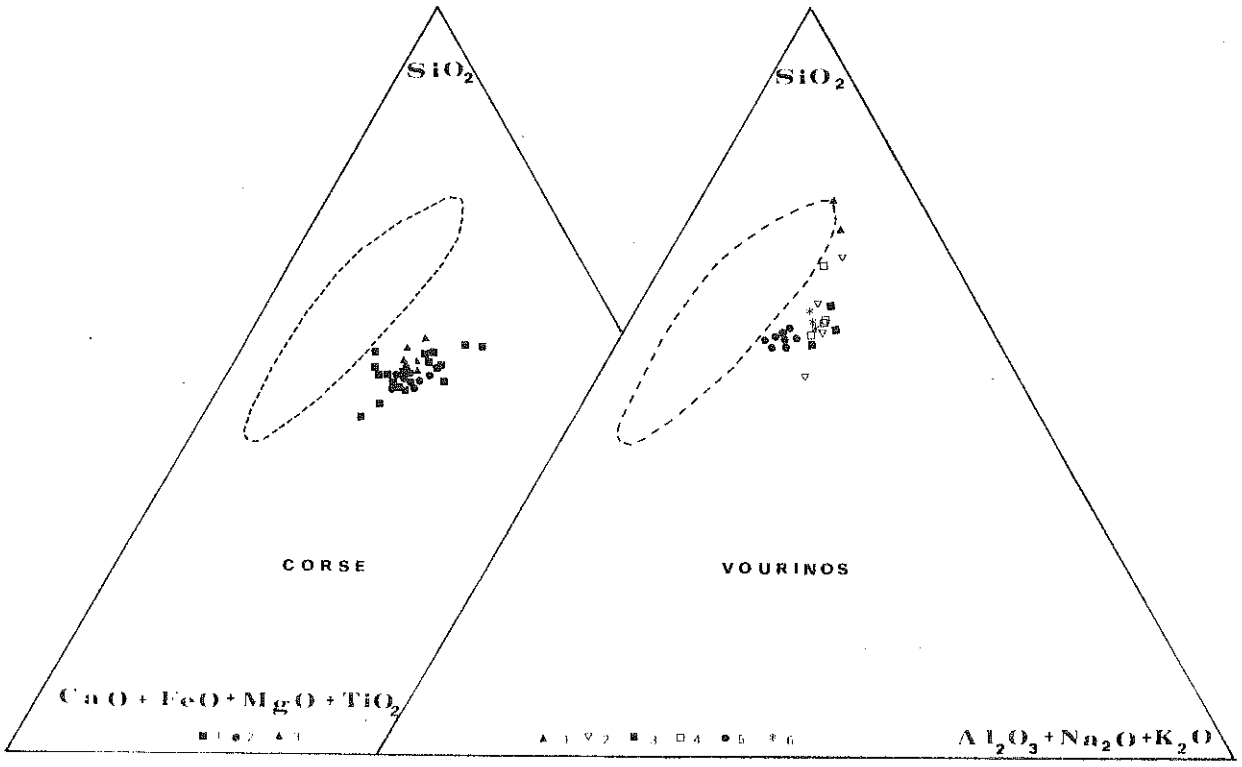

FIG. 7 — Diagramme SiO_2-Al_2O_3 + Na_2O + K_2O-CaO + FeO total + MgO + TiO_2 pour les liquides.
En pointillé, domaine d'immiscibilité de deux liquides d'après J. W. GREIG (1927) et E. ROEDDER (1951).
Corse : 1 = laves, 2 = dolérites, 3 = filons diabasiques. Vourinos : 1 = dacites, 2 = sills, 3 = laves massives, 4 = laves en coussins, 5 = basaltes, 6 = filons de basaltes, d'andésites basaltiques et de dacites.

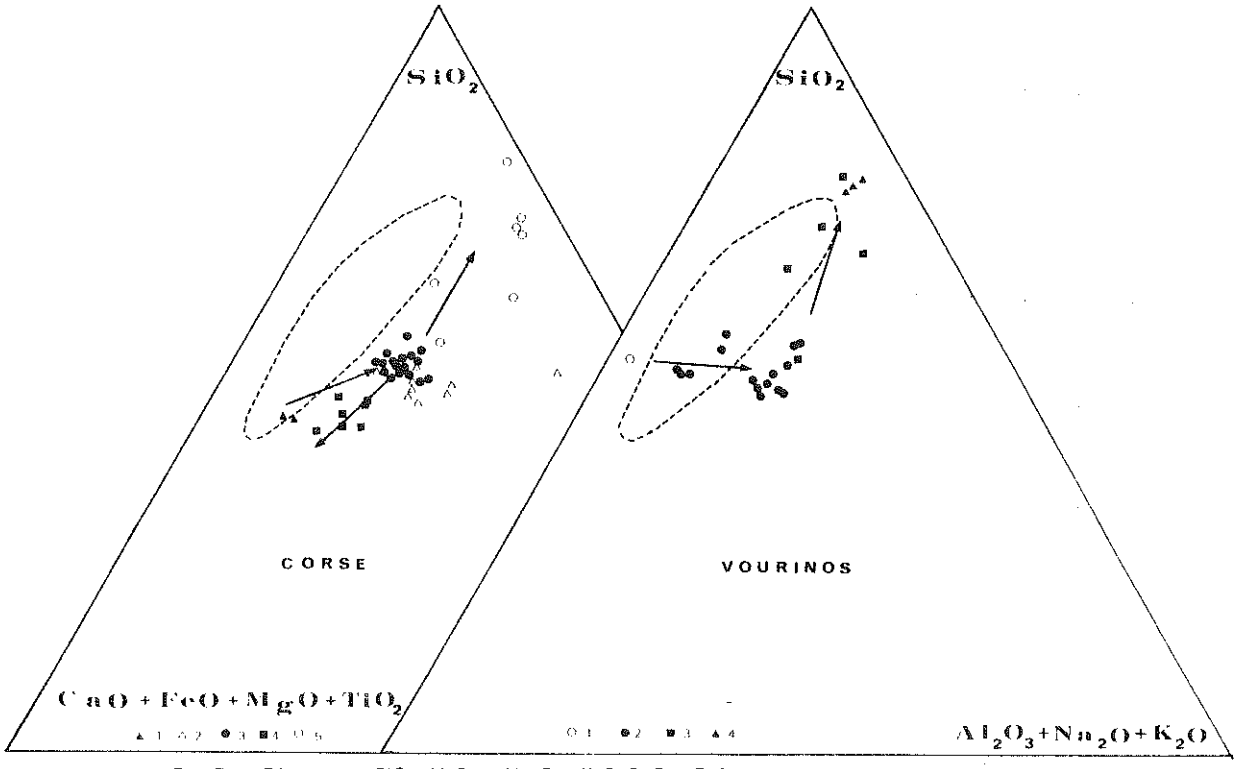

FIG. 8 — Diagramme SiO_2-Al_2O_3 + Na_2O + K_2O-CaO + FeO total + MgO + TiO_2 pour les solides.
Corse : 1 = dunites plagifères de l'unité 3, 2 = troctolites, 3 = euphotides, 4 = ferrogabbros, 5 = roches acides dont les albitites.
Vourinos : 1 = webstérite, 2 = gabbros et norites, 3 = diorites, 4 = granophyres.

peut supposer que deux liquides se forment, l'un pauvre en silice et sodium mais riche en fer, titane et phosphore et l'autre pauvre en ces éléments mais riche en silicium et sodium (le potassium étant toujours faible). Une telle hypothèse peut rendre compte du gisement des roches acides, de la distribution bimodale des ferrogabbros et des albitites, de l'existence du « trou » dans le diagramme de la figure 6, et du franchissement du plan critique de saturation par les derniers liquides. De plus ce mécanisme pourrait expliquer le fléchissement de la courbe du calcium observé sur la figure 5. En effet, la séparation d'un liquide riche en sodium à partir du magma l'appauvrit d'autant en cet élément. Les plagioclases qui y cristallisent sont alors relativement calciques et réduisent ainsi la pente de la courbe de CaO.

Dans le Vourinos, les cumulats qui tombent dans le domaine théorique d'immiscibilité indiquent que les conditions thermodynamiques n'étaient pas favorables, en particulier la température était peut-être trop élevée, pour qu'il y ait immiscibilité. On peut néanmoins constater qu'à partir du point d'inflexion les termes effusifs qui sont toujours en dehors du domaine des deux liquides en suivent néanmoins l'enveloppe.

Métallogénie des ophiolites

1 — Les gisements de chromite

T. P. THAYER (1972) a remarqué que le sous-type harzburgitique contient des chromites podiformes tandis que le sous-type lherzolitique n'en possède pas. Dans la classification des ophiolites en deux groupes, G. ROCCI et *al.* (1975) constatent la même distribution. Néanmoins, P. ROUTHIER (1963) avait déjà remarqué la présence de chromite dans les ophiolites de Yougoslavie, de Grèce, de Turquie et d'Iran, et leur absence dans celles de l'ensemble alpin s. s.

Au sein des ophiolites de type II, les chromitites peuvent occuper deux niveaux stratigraphiques (COLEMAN R. G., 1977) :

— soit dans les péridotites tectonisées : Vourinos (MOORES E. M., 1969), Turquie (VAN DER KAADEN G., 1970),

— soit près du contact péridotite-gabbros : Oman (GREENWOOD J. E. W. et LONEY P. E., 1968), Camaguey (Cuba) (FLINT D. E. et *al.*, 1948; THAYER T. P., 1964).

En fait dans ces ophiolites, la plupart des gisements de chromite importants sont associés aux dunites comme c'est le cas à Chypre, en Turquie et en Iran plutôt qu'aux harzburgites et pyroxénites.

En considérant que les harzburgites et dunites représentent le manteau supérieur appauvri, résidu de la fusion partielle d'une péridotite mantellaire, on a attribué aux chromitites qui leur étaient associées la même origine. Cependant la mise en évidence de phénomènes gravitaires dans les dunites et chromitites et les variations cryptiques dans la composition des chromites au sein d'un massif constituent des arguments en faveur d'une origine par accumulation (THAYER T. P., 1970; DICKEY J. S., 1975). Malgré la déformation intense des péridotites d'Iran, des preuves de tri de minéraux ont pu être observées (SABZEHEI M., 1975). De même, à Chypre, des textures magmatiques ont été reconnues par GREENBAUM D. (1972-1977) qui considère que les chromitites ont réagi comme un bloc rigide par rapport à l'encaissant, permettant ainsi entre les zones tectonisées la conservation des fragments et l'observation des textures primaires.

Les chromites des péridotites de type alpin évoluent d'un pôle riche en chrome vers un pôle riche en aluminium (THAYER T. P., 1970). Ceci peut être vérifié, en général, dans les chromitites des ophiolites. En Nouvelle Calédonie (GUILLON J. H., 1973), les teneurs de Cr_2O_3 évoluent de 54% dans les dunites à 41% dans les gabbros. Dans la masse péridotitique principale, la teneur en Cr_2O_3 dans les spinelles est toujours élevée. En Papouasie, l'alumine dans les spinelles augmente de 13,6% à 19,9% tandis que la teneur en Cr_2O_3 diminue de 55,7% à 32,5% des cumulats ultramafiques aux norites (ENGLAND R. N. et DAVIES H. L., 1973). Des variations semblables s'observent dans les chromitites d'Antalya (JUTEAU T., 1974) tandis qu'à Chypre, il n'est pas possible de mettre en évidence des évolutions cryptiques à l'échelle du massif bien que localement les concentrations de chrome varient rapidement (GREENBAUM D., 1977).

L'évolution des teneurs en Cr_2O_3 et en Al_2O_3 dans les chromites des ophiolites de type I se produit en sens contraire de celle observée dans les ophiolites de type II : les chromites des lherzolites sont plus riches en aluminium et plus pauvres en chrome que les chromites des cumulats ultrabasiques de l'unité de troctolite de Mattarana (BEZZI A. et PICCARDO G. B., 1971; PICCARDO G. B., 1976). Une même répartition s'observe en Corse entre chromites des péridotites du Monte Maggiore et celles des troctolites du bois de Pinetto (OHNENSTETTER D. et OHNENSTETTER M., 1975; à paraître).

On peut présenter le schéma des substitutions dans les chromites des deux types d'ophiolites :

	Péridotites	Gabbros	Liquides basaltiques associés
Type I	Al ⟶	Cr	riches en Cr
Type II	Cr ⟶	Al	pauvres en Cr

Les liquides qui ont été en équilibre avec les cumulats possèdent des concentrations de chrome différentes dans les deux types d'ophiolites. Le type II à gisements de chromites présente des liquides appauvris en chrome :

43 ppm dans le groupe de base du Troodos (DESMET A., 1977) et 106 ppm pour les laves en coussins inférieures (DESMET A., 1977; PEARCE J. A., 1975). Les concentrations en chrome dans les basaltes de Corse et des rides (type I) sont plus élevées : 216 ppm en Corse (BECCALUVA L. et *al.*, 1977). Les faibles teneurs en chrome des liquides des ophiolites de type II s'expliquent par la précipitation massive, précoce, de chromite riche en cet élément. Au contraire dans le cas de la Corse, l'absence de gisement important de chromite ne crée pas un appauvrissement notable du magma en cet élément. Le fractionnement important du spinelle dans un liquide basaltique peut être expliqué par une fugacité d'oxygène élevée (HILL R. et ROEDDER P., 1974), ce qui a fait suggérer à STRONG D. F. (1974) que les ophiolites issues de mers marginales pouvaient être plus riches en chromitite que les ophiolites issues d'océans vrais.

2 — Les minéralisations sulfurées

a) Les concentrations de nickel

D'une manière générale, les ophiolites ne possèdent pas de gisements importants de nickel. Néanmoins, on peut constater une diversité des associations nickélifères en fonction de la classification des ophiolites. En Corse, M. AZAIS et *al.* (1968) ont décrit des awaruites (fer nickelé) associées à des pentlandites et à des heazlewoodites, ces dernières étant caractéristiques des paragenèses sous-soufrées.

Dans les ophiolites de type II, les indices nickélifères : pentlandite et pyrrhotine associées à des heazlewoodites, awaruites et miliérites, sont relevés dans les péridotites riches en pyroxène de Nouvelle Calédonie (GUILLON J. H. et SAOS J. L., 1972; GUILLON J. H., 1973; 1975).

A Acoje (Philippines), les minéralisations de nickel, bien que disséminées, sont toutefois exploitées en même temps que les chromites (Philippines Bureau of Mines, 1972; PARINGIT R. U., 1975; THAYER T. P., 1976).

A Chypre (forêt de Limassol), des dépôts sporadiques de Cu, Ni et Co contenant de la pyrrhotine et de la pentlandite sont décrits dans les péridotites serpentinisées (CONSTANTINOU G., 1976).

b) Les amas sulfurés

Dans l'ensemble alpin (ophiolites de type I), J. BOULADON et P. PICOT (1968) décrivent quatre types de gîtes :

a) gîtes en amas généralement disloqués dans les serpentinites : Vigonza, Ponte Lecchia, Monte Catoni,

b) type pyriteux stratiforme dans les prasinites : Frangone, Erbajolo,

c) type filonien dans les gabbros : Casaluna, San Quilicio,

d) type stratiforme avec les niveaux de quartzite à riébeckite et aegyrine : Saint-Veran, Noceta, Vezzani.

Ces quatre types ont des paragenèses sous-soufrées : chalcopyrite, bornite, un peu de pyrite, blende + magnétite, pyrrhotite, oligiste et possèdent également des tellurures, notamment la mélonite. De telles paragenèses ont également été décrites dans les Apennins (BRIGO L. et *al.*, 1976).

Les gisements des ophiolites de type II de taille importante comme ceux de Chypre, Turquie et de Betts Cove, Terre Neuve, sont constitués de sulfures sur-soufrés. Ils sont liés à la stratigraphie des laves : à Chypre les minéralisations sont sous les laves supérieures (CONSTANTINOU G. et GOVETT G. J. J., 1973), et aux circulations de fluides riches en eau salée dans les failles (UPADHYAY H. D. et STRONG D. F., 1973; CHAPMAN H. H. et SPOONER E. T. C., 1977; SPOONER E. T. C. et BRAY C. J., 1977). Ces gisements étroitement liés à la formation des ophiolites résulteraient de l'activité hydrothermale, en particulier des circulations convectives de l'eau de mer dans les zones d'accrétion océanique. Néanmoins E. T. C. SPOONER et C. J. BRAY (1977), à partir de l'étude des inclusions fluides, soulignent particulièrement l'analogie entre les températures de congélation de ces fluides dans les gisements de Chypre et ceux de type Kuroko au Japon. Ces résultats sont en accord avec la création des ophiolites de type II et de leurs gisements associés dans un contexte moins franchement océanique et voisin d'un arc insulaire.

L'absence de gisements importants dans les ophiolites de type I est probablement liée à une insuffisance de soufre dans le magma (WEGMANN C. E., 1926; ROUTHIER P., 1946; BOULADON J. et PICOT P., 1968; BOULADON J. et *al.*, 1976). Au contraire, les ophiolites de type II, qui présentent le plus souvent des amas pyriteux importants ne se distinguent pas des gisements liés au volcanisme calco-alcalin épicontinental (BOULADON J. et *al.*, 1976).

Résumé et conclusion

L'évolution géochimique des cumulats et des laves associées est en grande partie contrôlée par la nature des minéraux et par le type de fractionnement. Les variations brutales dans la concentration des oxydes sont dues à la précipitation locale massive de certains minéraux : pics de CaO et de Al_2O_3 respectivement dans les zones de transition pyroxénitique et troctolitique, pics de FeO total et de TiO_2 au niveau des ferrogabbros. Dans le cas de la Corse, il semble que des cristallisations à l'équilibre aient eu lieu, telles que l'indiquent la texture et la minéralogie des cumulats, l'évolution du rapport FeO/MgO et la variation des éléments incompatibles en fonction de ce rapport.

Pour les mêmes raisons, l'évolution des liquides dans le cas du Vourinos est due à une cristallisation fractionnée plus ou moins parfaite. Les fractionnements de l'olivine, du plagioclase, du clinopyroxène et dans une moindre mesure de l'orthopyroxène et des oxydes ont lieu sous de faibles pressions : 3 kb dans le cas de la Corse et des rides et au voisinage de 1 kb pour le Vourinos. Sous ces pressions, les magmas à l'origine des cumulats et des basaltes sont des tholéïtes à olivine ou à quartz. Le magma du Vourinos, plus saturé que celui de la Corse, semble aussi plus riche en CaO mais plus pauvre en alcalins et titane. Par ailleurs les éléments incompatibles tels que Zr, Nb et Y et le chrome et le nickel sont en faible concentration dans les tholéïtes à quartz du Troodos et du Vourinos (KAY R. W. et SÉNÉCHAL R. G., 1976; BECCALUVA L. et *al.*, 1977, 1978). En conséquence, les termes effusifs et la plupart des cumulats qui représentent une section importante de la séquence ophiolitique sont issus, sous de faibles pressions, de fractionnements différents à partir de magmas tholéïtiques également différents.

La validité de la classification des ophiolites en deux types (ROCCI G. et *al.*, 1975) repose en partie sur la distinction des tholéïtes à olivine et des tholéïtes à quartz et donc sur l'efficacité du plan Di-Pl-Opx de sursaturation. Pour des pressions inférieures à 5 kb (O'HARA M. J., 1968) des liquides situés de part et d'autre de ce plan ne peuvent pas le franchir par un processus normal de fractionnement. Sous ces conditions, les tholéïtes à olivine et les tholéïtes à quartz évoluant vers des eutectiques différents, la classification des ophiolites en deux groupes est alors justifiée. Comme il est improbable que des fusions importantes aient eu lieu dans ces zones de basses pressions, il s'en dégage que les caractères magmatiques respectifs des deux types d'ophiolites ont été acquis à de plus hautes pressions.

Il est généralement admis que les harzburgites appauvries et tectonisées à la base des complexes ophiolitiques représentent les résidus de fusion d'une péridotite mantellique dont les produits de fusion auraient donné basaltes et cumulats. En Corse et dans les Alpes de telles roches restent à découvrir, de même que dans les rides où les péridotites ont une composition proche de lherzolites. En admettant que ces harzburgites représentent les résidus de tous les magmas ophiolitiques, il peut y avoir suffisamment de cheminements divers dans les zones de hautes pressions pour obtenir près de la surface des magmas différents. On peut supposer que les sources n'étant pas identiques, les deux magmas qui en seraient issus à des taux de fusion suffisamment importants pour qu'une deuxième phase disparaisse, seraient alors différents. Mais on peut également considérer que les sources étant voisines, des fractionnements de hautes pressions de caractères particuliers permettraient de diversifier les magmas. Dans le cas de la Corse ou des basaltes océaniques, la fusion d'une péridotite lherzolitique donnerait un liquide qui fractionnerait essentiellement de l'olivine pendant son ascension et un résidu : harzburgite ou péridotite appauvrie. La formation des tholéïtes à quartz du Vourinos ou de Chypre nécessite la mise en œuvre de modèles à plusieurs stades. La faible concentration des alcalins et des éléments incompatibles peut être expliquée par une fusion à partir d'une péridotite appauvrie ou par une fusion plus importante à partir d'une péridotite non appauvrie (SMEWING J. D. et *al.*, 1975). Dans les deux cas, le résidu sera très réfractaire, constitué par exemple de harzburgites. De même, le peu de chrome et de nickel dans les liquides nécessite, en plus de la précipitation de l'olivine, le fractionnement du spinelle dans des zones de hautes pressions.

Quelle que soit l'hypothèse retenue, on est obligé d'envisager des cheminements différents dans la création et l'évolution des magmas ophiolitiques. Cette évidence pose en fait le problème de l'homogénéité de la croûte océanique dans l'hypothèse où toutes les ophiolites représentent des fragments de cette croûte. De ce point de vue et étant donné la nature pétrographique et géochimique des ophiolites de Corse, celles-ci représentent le meilleur exemple de croûte océanique créée dans une dorsale de type médio-océanique et incorporée postérieurement au continent. Le modèle de fonctionnement d'une ride actuelle à partir du modèle corse peut alors s'appliquer (OHNENSTETTER D. et OHNENSTETTER M., 1976). Au contraire, les contraintes pétrologiques imposées par la nature de la fusion et des fractionnements de hautes pressions ne permettent pas d'appliquer la genèse des tholéïtes océaniques aux basaltes des ophiolites de type II. Par ailleurs, la nature du fractionnement, l'existence de courants continus dans les chambres magmatiques situées à faible profondeur résultent de conditions thermodynamiques particulières, différentes de celles des rides et liées à un autre environnement géotectonique. En conséquence, la création de ce type d'ophiolite dans un environnement plus continental au voisinage d'un arc ou dans une mer marginale est alors plus plausible.

D'un point de vue métallogénique, la distribution des gisements en deux grandes catégories : type (I) sans chromite et à amas sulfurés sous-soufrés de faible extension, et type (II) à chromite et à amas sulfurés sur-soufrés, qui est en accord avec la classification des ophiolites en deux groupes : type I-Corse et type II-Chypre-Vourinos, souligne le bien-fondé de l'hypothèse de la dualité.

La répartition des gisements en massifs stériles (type I) et fertiles (type II) peut être expliquée par des conditions physico-chimiques propres à chaque type de magma. Le magma de type I est caractérisé par une forte activité de FeO et une faible fugacité d'oxygène qui se traduisent par un enrichissement en fer au cours de la différenciation et par la rareté des chromites. De plus les paragenèses sous-soufrées indiquent une fugacité du soufre peu élevée dans le bain silicaté responsable de l'absence de gisement important. De la même façon, les concentrations de chromitites et de sulfures dans les ophiolites de type II peuvent être causées par de fortes fugacités d'oxygène et de soufre dans un magma sursaturé en silice. Les ophiolites sont réputées stériles en gisements de nickel sulfuré, toutefois dans certains complexes du type II, il existe des indices (Kellaki, Chypre) ou de petits gisements (Acoje, Philippines; Thetford, Québec).

Dans l'état actuel de nos connaissances, compte tenu de la stérilité relative des ophiolites de type I qui se seraient formées dans les zones d'accrétion médio-océanique, on peut raisonnablement envisager que les océans sont dépourvus de concentrations importantes de chromites et de sulfures. Les gisements liés au type II se seraient développés dans un contexte plus continental : arc insulaire ou bassin inter-arc.

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COMPARISON BETWEEN CORSICAN ALBITITES AND OCEANIC PLAGIOGRANITES

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ABSTRACT

The nomenclature of low-K leucocratic rocks belonging to the upper part of the cumulate sequence in ophiolites is reconsidered. In Corsica, the geological setting, textural and petrographical features of the acid differentiates mainly found in the Rospigliani series lead to the following classification. Coarse-grained leucoferrodiorites with accumulate textures, sometimes deformed, have crystallized with ferrogabbros and ferrodiorites. The most abundant medium-grained albitites (meta-tonalites or albite-syenite), rich in accessory minerals, locally marked by flow and accumulate features, possess subhedral granular and/or cataclastic textures. These rocks and associated dolerites may belong to an upper border group of the magmatic chamber. The fine-grained albite-granite (meta-plagiogranites) with allotriomorphic granular textures occur mainly in dykes cutting the upper gabbros and the volcano-sedimentary formations of the Rospigliani series. Granophyric intergrowths recognized in the less evolved term are often obscured by brittle and plastic deformation related to magmatic injection (cataclastic to mylonitic textures).

The oceanic leucocratic rocks may be classified in the same way. A geochemical compilation of Corsican and oceanic samples has reinforced the similarities between the two suites. A compositional gap does not occur in the range 50-80% SiO_2. However, three main populations at about 55, 60 and 70% silica corresponding respectively to leucoferrodiorites, albitites (tonalites) and albite-granites (plagiogranites) have been emphasized.

The genesis of these rocks is related to the evolution of high-level magma chambers. Immiscibility may occur after the beginning of ferrogabbro crystallization, inducing various magmatic breccias and rapid ascent of more or less unmixed acid and basic magmas. As shown by REE patterns fractionation may occur in the silicic liquid after immiscibility. This process and late stage hydrothermalism as well as oceanic tectonism have favoured the dismantling of the upper border group, well exposed in the Rospigliani volcano-sedimentary series possibly formed in a transform-fault zone.

INTRODUCTION

Leucocratic rocks have long been known to occur in the upper part of ophiolitic cumulates but in variable proportion and composition: rare and plagioclase-rich in Corsica (type I ophiolites) and more abundant and quartz-rich (granophyres) in type II ophiolites (Rocci *et al.*, 1975) reaching 20% if keratophyres are also con-

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sidered (BROWN *et al.*, 1979). As in type I ophiolites, acid residues are rare in the ocean, even in transform-fault zones, where lower cumulates and peridotites are well exposed. Therefore, a more extensive comparison between Corsican albitites and oceanic leucocratic rocks has been undertaken.

In Corsica, pre-Malm slices of the Liguro-piemontais oceanic crust are composed of lavas, mainly pillow-lavas, dolerites, gabbros and more or less serpentinized lherzolitic peridotites formed at mid-oceanic ridges, transform fault zones and seamounts. Low pressure fractionation of liquids corresponds to the crystallization of troctolites, olivine-gabbros, euphotides and ferrogabbros ending with acid differentiates (OHNENSTETTER and OHNENSTETTER, 1975; BECCALUVA *et al.*, 1977).

The large gabbro and lava pile are more or less devoid of leucocratic acid rocks. Most of them are found in the Rospigliani series formed of a volcano-sedimentary sequence resting on a serpentinite or gabbro floor. Contemporaneous acid and basic magmas, which have intruded the sediments, may be related to a peculiar type of oceanic accretion occuring in a possible transform-fault zone (OHNENSTETTER, 1979*a*). The setting of the leucocratic residue may give information on the development of oceanic magma chambers.

I. NOMENCLATURE OF LEUCOCRATIC RESIDUES

Nomenclature of acid residues must reflect mineral assemblages and not a geotectonic environment. For this reason the term oceanic plagiogranites (COLEMAN and PETERMAN, 1975; COLEMAN and DONATO, 1979) used as a general term for the leucocratic rocks in the ophiolitic suite must be restricted only to dredged and cored samples from the oceans. Similarly, if a general term is needed to design the quartz-bearing leucocratic assemblage in the field, ophiolitic plagiogranite may be employed.

Most of the problems of these rocks come from their low K_2O contents, which result in the scarcity of K-feldspar and biotite. Nevertheless, the proposed classification of STRECKEISEN (1967, 1974) may be used, whatever the nature of the mafic minerals. Thus, in the presence of one feldspar (plagioclase) most of the rocks fall on the QP join and depending upon the quartz content would be called diorite, quartz-diorite ($5 < Q < 20$), tonalite ($20 < Q < 40$) and plagiogranite ($Q > 40$) using the subdivision of GINSBURG *et al.* (1962) in STRECKEISEN (1967). Trondhjemites belong to the diorite-tonalite family with a color index < 15 (STRECKEISEN, 1967). Using the CIPW normative proportion of feldspar (O'CONNOR, 1965) most of the acid residues fall in the trondhjemite and tonalite fields (COLEMAN and PETERMAN, 1976). If the An content is less than 5%, then the leucocratic rocks fall on the AQ join. Depending upon the amount of quartz, they are called albite-syenite, quartz albite-syenite and albite-granite.

The granophyres often described in type II ophiolites are mostly tonalite, trondhjemite and plagiogranite. The granophyric texture appears in interstitial position in quartz-diorite and invades progressively the groundmass in plagiogranite. In type I ophiolites, this texture is scarce partly due to the rarity of quartz-bearing rocks (BROUWER and EGELER, 1952). Moreover, as the composition of plagioclase has always been modified by several metamorphic events, the position of rocks around either the A or P apex is debatable. For this reason the term albitite has generally been used in the alpine chain (JAFFE, 1955; SALIMI, 1965; STEEN, 1972; BERTRAND and VUAGNAT, 1979) and corresponds to Q-poor rocks (see TRÖGER, 1935 and JOHANNSEN, 1937) in the syenite (LACROIX, 1922; PUSZTASZERI, 1969) or diorite families. With the presence of quartz, albite-granite was employed.

II. EMPLACEMENT OF ACID ROCKS RELATED TO THEIR TEXTURES IN CORSICA

Leucocratic rocks are mainly found in the upper cumulates and volcanics (OHNENSTETTER and OHNENSTETTER, 1975). As they represent the latest magmatic products (ROUTHIER, 1945; NETELBEEK, 1951; BROUWER and EGELER, 1952; SPIJER, 1955) their emplacement is closely related to the history of the upper part of the magma chamber.

In cumulates they occur in dykes or are closely associated with ferrogabbros. In the dykes cutting euphotides, intermediate gabbros and ferrogabbros, the composition of the leucocratic rocks varies from diorite, quartz-diorite, leucoferrodiorite to albitite. Generally, the dioritic rocks are coarse grained, with large more or less zoned plagioclase, relicts after green hornblende, some quartz and scarce and small zircon. Local recrystallization around plagioclase porphyrocrysts and at the dyke border are linked to magma emplacement (Pl. III C).

In the upper cumulates, leucocratic rocks, mainly leucodiorites are interbedded or intermingled with ferrogabbros and some noritic gabbros. Various ortho and mesocumulates may be defined depending upon the phase proportion and form contacts. Leucodiorites clearly separated from ferrogabbros present a hiatal porphyritic granular textures with coarse grained plagioclase, sometimes deformed and surrounded by smaller ones. Locally, they contain scattered aggregates of clinopyroxene, opaque minerals and apatite forming clusters characteristic of the iron-rich gabbros and diorites. The distribution of mafic minerals is more regular in leucoferrodiorites, which possess elongate euhedral juxtaposed plagioclases and in intersertal position, cumulus apatite and pyroxene enclosed in opaque minerals (Pl. II A).

Less mafic quartz-diorites possess zircon, more abundant when quartz is notably present. In all of these cumulates, magmatic lamination is often well developped. They are locally deformed and cut by basic dykes as in the Alps s.l. (GIANELLI and

PRINCIPI, 1974; OHNENSTETTER and OHNENSTETTER, 1975; 1976; STEEN *et al.*, 1977; MEVEL *et al.*, 1978; BERTRAND and VUAGNAT, 1979). Elsewhere magmatic melange composed of basalt, leucoferrodiorite, diorite and gabbro indicate that injection of liquids in hot cumulates was contemporaneous with deformation and in some places may induce it (OHNENSTETTER and OHNENSTETTER, 1975).

Large boulders of albitites with a finer grain size may be found in slices belonging to the Rospigliani series. In the highest cumulates these rocks may enclose pieces of ferrogabbro picked up when rising to the surface. Above, they are locally associated with ferrodolerites and leucoferrodolerites. They may also occur in dykes cutting serpentinites (PETERLONGO, 1968). They are fine grained, rich in accessory minerals such as zircon, apatite, opaque minerals and allanite and exhibit a wide range of textures. Zircons give a middle Jurassic age (OHNENSTETTER *et al.*, 1976; and submitted). Some samples clearly show orthocumulate features with granular textures marked by euhedral plagioclase developed during igneous lamination (Pl. IA; B). Crude internal layers of zircon and opaque minerals have been observed (Pl. IVC). These rocks remind one of the congelation cumulates of the Upper Border Group of Skaergaard (WAGER and BROWN, 1967). Other albitites with subhedral to anhedral granular textures possess randomly oriented plagioclase in a finer matrix (hiatale texture) of plagioclase, zircon and minor quartz (Pl. IC; D; II B). Relict forms after pyroxene or amphibole closely associated with opaque minerals and some apatite (comparable to clinopyroxene, titanomagnetite and apatite clusters in ferrogabbros) are organized in a decussate texture. In rare places, primary plagioclase has grown until in contact giving a pegmatitic equigranular texture. Moreover, in some outcrops monogenic acid breccias with a matrix mainly of plagioclase clasts may result from a late hydrothermal process (Pl. III B).

Leucocratic rocks varying from albitites to albite granites are also found cutting polygenic breccias and ophicalcites of the Rospigliani volcano-sedimentary series (OHNENSTETTER, 1979). They are fine grained with allotriomorphic granular (Pl. II C; III A) and sometimes intersertal textures (Pl. IV A; B). Myrmekites around plagioclase are present in the less evolved albite granites (Pl. IV D). With increase in silica, early crystallization of quartz with Na-plagioclase prevents the development of granophyric texture. Moreover, graphic intergrowths are often destroyed during brittle and ductile deformation related to magmatic injection. Cataclastic rocks possess xenomorphic plagioclase with undulose extinction scattered in a fine grained albite and quartz matrix (Pl. III C; D). At a more plastic stage, subgrains in foliated albite granites crystallized from the porphyroclasts (Pl. III A). Cataclastic textures are also common in Canyon Mountain acid rocks (THAYER and HIMMELBERG, 1968). Acid magmas have often been emplaced simultaneously with a more basic liquid (Pl. IV A; B). The following features are relevant of this phenomenon with some of them probably linked to an immiscible process (ROEDDER, 1979): round to oval trapped inclusions of acid and basic magma in each other,

mechanical mixing of the magmas, nucleation and perpendicular growth from the margins of the globules, grain size increase far from the contact and common foliation in dykes. Discrepancies in the textures from acid and basic rocks are linked to the different rheological states of the magmas. Both liquids are also intermingled in extrusives (OHNENSTETTER, 1979).

III. PETROGRAPHY AND MINERALOGY

1. CORSICA

As previously mentioned, leucocratic rocks may be placed in three main groups; diorites especially leucoferrodiorites, albitites (albite-syenite) and albite-granite. Leucoferrodiorites possess zoned (sometimes reversed) plagioclase of oligoclase composition, clinopyroxenes with herringbone twins, euhedral apatite enclosed in ilmenite and titanomagnetite. Quartz and zircon may be present as well as brown hornblende. In albitites similar phases may be recognized, except for fresh magmatic clinopyroxenes. The existence of Na-pyroxene rich albitite and Na-amphibole rich albitites (NETELBEEK, 1951) has lead to the supposition that magmatic pyroxene and amphibole have occurred (OHNENSTETTER *et al.*, 1976). The proportion of apatite and opaque minerals decreases for the lower acid cumulates, whereas zircon increase. Allanite is common, rimmed by pistacite. In one sample biotite flakes have been observed. The previous existence of magmatic aegyrine (ROUTHIER, 1945; 1946) may be suggested by relict form after elongated pyroxene. The last products, albite granites, are poor in mafic and accessory phases, especially zircon. In all three groups, especially albitites and albite granites blueschist facies assemblages, occurring between two greenschist metamorphisms, have overprinted the magmatic assemblages (OHNENSTETTER *et al.*, 1976; in preparation). The compositions of fresh magmatic clinopyroxene from leucoferrodiorites and ferrogabbros have been plotted (Fig. 1). They are more Mg and Ca rich than clinopyroxene from the ferrogabbro. Plagioclase is oligoclase in leucoferrodiorite and nearly pure albite in albitite and albite granite (see Fig. 2).

2. OCEANIC AREAS

A few occurrences of leucocratic rocks have been reported from the ocean in different settings: seamounts near Mid Atlantic Ridge at 45°N (AUMENTO, 1969); in fracture zones (MIYASHIRO *et al.*, 1970; ENGEL and FISHER, 1975; OHNENSTETTER, in prep.) in the Aves Rise (WALKER *et al.*, 1972), in the Kjushu-Palao ridge from marginal basin (ISHIZAKA and YANAGI, 1975). They are often associated with ferrogabbros, gabbros, olivine gabbros and serpentinized lherzolitic peridotites in the fracture zones. Oceanic samples may be ranged in the three main groups defined for Corsican rocks. The coarse grained leucoferrodiorite from the Vema fracture zone

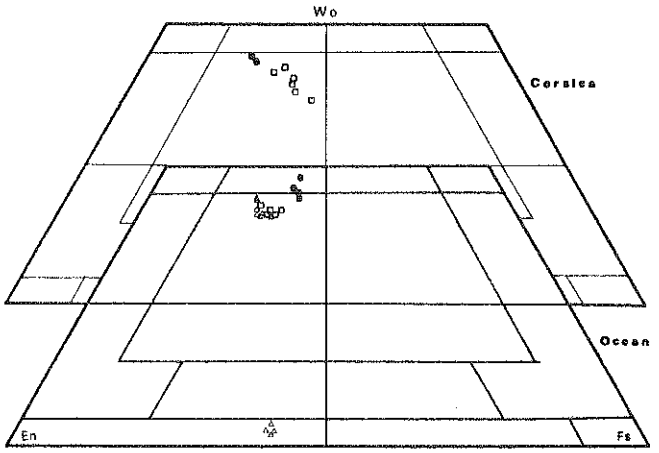

FIG. 1. — **Clinopyroxene compositions in the Hess quadrilateral.** Corsican samples: dot, leucoferrodiorite (75-4.6.8.); open square, ferrogabbro (20.9.5.). Oceanic samples from the Vema fracture zone: dot, tonalite (CH78-DR10-002A); open square, ferrodiorite (CH78-DR10-74); open triangle, ferrodiorite (CH78-DR1P-003).

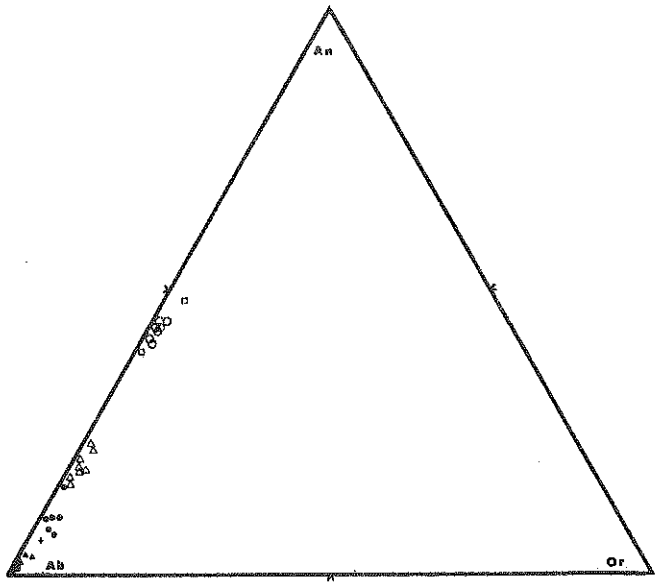

FIG. 2. — **Plagioclase composition of oceanic and corsican samples in the albite-orthose-anorthite triangle.** Corsican samples: dot (75-4.6.8.) and cross (75-7.6.9.), leucoferrodiorite; triangle, albitite (23.8.6-1A and 24.7.21.).Oceanic samples: open circle, ferrodiorite (CH78-DR10-003); open square, leucoferrodiorite (CH78-DR1P-74); open triangle, tonalite (CH78-DR10-002A).

is a cumulate with brown hornblende after clinopyroxene, relict forms after orthopyroxene, euhedral apatite, zircon and poikilitic titanomagnetite and ilmenite. Sample ANTP 2 (ENGEL and FISHER, 1975) belongs to the albitite-tonalite-trondhjemite group because of the medium grained texture and abundant accessory minerals like apatite and zircon. A fine grained rock of the same dredge may belong to the plagiogranite stem with minor accessory minerals like zircon (up to 25 μm) and micropegmatitic texture K-feldspars if they occur (quartz-monzonitic dykelet of ENGEL and FISHER, 1975) shown a heterogeneous repartition.

In a general way, andesine, albite, quartz, brown hornblende, apatite, magnetite, zircon, titanite are in varying proportions the most important phases which form the oceanic leucocratic differentiates. Clinopyroxene compositions of ferrodiorite and leucoferrodiorite from the Vema fracture zone are plotted in Fig. 1. As for Corsican samples, clinopyroxenes from the leucoferrodiorite are richer in wo content than the associated cumulates and may be due to the increase in alkalinity. Plagioclase compositions (Fig. 2) are similar to those of the Corsican leucoferrodiorite and much more sodic than those of the dredged ferrodiorites.

IV. CHEMISTRY

The compositions of Corsican and oceanic acid rocks are reported in Tables 1, 2 and 3. The low initial $^{87}Sr/^{86}Sr$ ratio in dredged samples is in agreement with derivation from a sub-oceanic mantle (ISHIZAKA and YANAGI, 1975). From textural evidence, the leucocratic rocks may be considered in some cases as solids with variable amounts of trapped liquid or as liquid with some primocrysts and clusters. No major gap is found in the 50-80% silica range. Leucoferrodiorites contain about 55% silica, the most important group, albitites around 60% and the plagiogranites reach 70-80%. These three main groups occur also in oceanic samples (Table 2). In both series the (FeO^T/MgO) ratio varies considerably. Aluminia decreases in the albite granites. SINTON and BYERLY (1980) have proposed a high-temperature vapor phase transport for removal of potassium during late-stage crystallization of silicic differentiates of abyssal oceanic magma. The concentrations of incompatible elements like Zr, Hf and RE may reach high values compared to type II ophiolites (COCHERIE, 1978). Zirconium and hafnium show strong maxima at 60% silica (albitites) whereas phosphorus is rather concentrated in ferrodiorites and leucoferrodiorites. Variable concentrations in LIL elements may be due to late stage hydrothermal processes or metamorphic alteration (COLEMAN and DONATO, 1979).

On an AMF diagram (Figs. 3 and 4) the acid rocks occupy the field between ferrogabbros and the A apex. In the GREIG's diagram (Figs. 5 and 6) oceanic and Corsican cumulates show a similar repartition linked to a common mechanism of formation. Ferrogabbros and acid rocks may be obtained through an immiscible

TABLE 1

Chemical compositions of corsican albitites. 1, leucoferrodiorite (in OHNENSTETTER and OHNENSTETTER, 1975 and BECCALUVA *et al.*, 1977); Inzecca massif; albite, oligoclase, Na-amphibole, zircon, apatite, ores. 2, pegmatoïde albitique (in BAUD, 1975); Balagne. 3, faciès grenu mésocrate (in COUTURIE, 1964); Pont de Piedicorte. 4, Pegmatoïde albitique (in BAUD, 1975); Balagne. 5, leucoferrodiorite, Punta di Puriolo; orthocumulate; oligoclase, albite, quartz, clinopyroxene, brown hornblende and ores. 6, metagabbro (in FRANCONI, 1967); Casaluna. 7, albitite-trondhjemite (in COCHERIE, 1978, sampled by OHNENSTETTER); Inzecca massif, orthocumulate with igneous lamination; albite, minor quartz, Na-pyroxene, Na-amphibole, zircon, ores. 8, albitite-trondhjemite (in OHNENSTETTER and OHNENSTETTER, 1975); Inzecca massif; metaplagioclasite without relict of primary texture; quartz, pumpellyite, chlorite. This sample is not reported in diagram. 9, albitite-trondhjemite (in OHNENSTETTER and OHNENSTETTER, 1975 and BECCALUVA *et al.*, 1977); Inzecca massif; same petrographic features as sample 7. 10, albitite, Punta di Puriolo; cataclastic and heterogeneous granular texture; albite, pistacite, quartz, chlorite, ores. 11, albitite; Punta Muzzone; granular texture; clinopyroxene, albite, green hornblendes. 12, albitite (in LACHAR-

	1	2	3	4	5	6	7	8	9	10	11
Sample No.	297131	105	25.62	133	P7985A2	188	2386 IAA	1.8.10/2	23.8.8 IAA	P79-83 1	M79 95B
SiO_2	54.00	55.30	57.05	57.92	58.09	58.70	59.90	60.08	60.48	60.87	60.98
TiO_2	1.44	.61	.50	.56	1.11	1.00	.88	.49	.79	.43	.49
Al_2O_3	14.15	15.44	16.75	18.84	15.98	16.20	17.23	16.74	16.99	17.67	17.82
Fe_2O_3	15.78	8.58	3.40	5.24	8.02	4.20	6.32	1.29	6.25	5.18	2.57
FeO			5.10			5.60					
MnO	n.d.	.15	.15	.11	.22	.20	.06	.04	.06	.06	.05
MgO	2.76	4.23	3.30	2.48	2.01	1.90	1.38	2.88	1.49	.77	4.19
CaO	2.22	6.72	4.85	3.68	2.97	3.50	1.78	13.95	1.84	4.92	1.50
Na_2O	8.02	6.56	7.00	8.48	8.22	6.70	10.80	.15	11.28	8.21	9.01
K_2O	.09	.27	0.00	.25	.47	.10	.10	.01	.05	.05	.17
H_2O^+	1.36	1.06	2.05	1.15	2.97	1.10	.71	4.15	.65	1.71	2.67
H_2O^-			0.00			.45			.15		
P_2O_5	.40		.10			.40	<.10				
CO_2	—		—			—					
Total	99.82	98.92	100.25	98.71	99.88	99.70	99.10	99.77	100.03	99.87	99.45
Sr	37	223	—	129	—	—	—	—	27	—	—
Ba	—	10	—	17	—	—	37	—	—	—	—
Cr	—	118	—	149	—	—	35.3	—	—	—	—
Zr	620	—	—	—	—	—	1638	—	2200	—	—
V	49	101	—	25	—	—	—	—	—	—	—
Ni	110	336	—	10	—	—	1	—	22	—	—
Cu	18	58	—	89	—	—	—	—	10	—	—
Co	25	19	—	10	—	—	24.7	—	10	—	—
Sc	—	—	—	—	—	—	11.9	—	—	—	—
Zn	208	—	—	—	—	—	—	—	65	—	—
Pb	—	—	—	—	—	—	—	—	—	—	—
Ga	—	—	—	—	—	—	—	—	—	—	—
B	—	—	—	—	—	—	—	—	—	—	—
Rb	<6	—	—	—	—	—	.09	—	<6	—	—
Nb	11	—	—	—	—	—	—	—	16	—	—
Y	170	—	—	—	—	—	—	—	—	—	—
Yb	—	—	—	—	—	—	—	—	—	—	—
Li	15	—	—	—	—	—	—	—	11	—	—

PAGNE, 1970); Pont de Teddia; dike in euphotide. 13, albitite, Inzecca massif; equant granular texture; zoned plagioclase with albite rims, glaucophane, stilpnomelane, chlorite, zircon. 14, albite-crossitite (in BROUWER and EGELER, 1952); Bocca Serna. 15, albite-granite-trondhjemite (in COCHERIE, 1978 sampled by OHNENSTETTER); Inzecca massif; orthocumulate with igneous lamination; albite, quartz, Na-pyroxene, Na-amphibole, chlorite, ores, titanite and zircon. 16, Pegmatoïde albitique (in BAUD, 1975); Balagne. 17, albite-granite-trondhjemite (in OHNENSTETTER and OHNENSTETTER, 1975); mylonitic texture; albite, quartz, chlorite, ores and rare zircon. 18, albite-granite-trondhjemite (in OHNENSTETTER and OHNENSTETTER, 1975); Inzecca massif granular texture; albite, interstitial quartz, actinote, pumpellyite, titanite, zircon. 19, albite-granite; Casaluna; dike in intermediate gabbros; granular texture with myrmekite; albite, quartz, chlorite, titanite, zoïsite, rare zircon. 20, albite-granite-trondhjemite (in COCHERIE, 1978 sampled by OHNENSTETTER); Inzecca massif; hiatal granular texture with myrmekite; albite, quartz, chlorite, ores and titanite. 21, residual differentiate (in BROUWER and EGELER, 1952); Noceta. 22, albite-granite-trondhjemite (in OHNENSTETTER and OHNENSTETTER, 1975); Inzecca massif; cataclastic and hiatal textures; anhedral albite in a fine grained albite and quartz matrix with a planar texture; stilpnomelane. 23, albite-granite, Puriolo; fine-grained euhedral and planar textures; albite, quartz, crossite, pistacite.

12	13	14	15	16	17	18	19	20	21	22	23
JCL 136	23.8.6/2	VII	2873	77	30.7.3.1	1.8.2.2	1.8.6	25.8.1	VI	28.7.9	79.84B
62.45	62.47	63.57	64.10	66.08	68.91	69.40	70.20	71.00	74.19	79.45	79.65
.70	.80	.70	.91	.26	.19	.25	.25	.27	.28	.15	.12
16.35	16.21	16.16	13.67	18.96	16.52	16.12	15.95	13.87	13.99	12.08	11.48
6.15	5.78	3.71	8.17	1.18	3.01	.70	1.79	3.89	.81	1.35	.52
1.30		2.66							.92		
.10	.04	.08	.12	.03	—	.02	.08	.03	.03	n.d.	.01
2.00	1.55	1.58	1.46	.72	2.17	.78	.45	2.98	.94	<.01	.08
1.20	1.11	1.10	1.40	1.47	.03	3.10	1.67	.10	.68	.10	.32
7.35	10.23	9.55	8.65	10.17	7.84	7.49	6.93	5.80	7.32	6.59	6.59
.30	.18	.15	<.10	.17	<.01	<.01	.77	<.10	.08	<.01	.05
1.70	1.41	.84	1.90	.48	1.28	1.41	1.39	.48	.64	.54	.59
.45		.10						<.10	.07		
.35		.17	.13						.03		
—		—	—						.02		
100.40	99.78	100.37	100.50	99.52	99.95	99.27	99.48	98.30	100.00	100.26	99.41
—	—	—	84	1.41	—	—	—	—	—	9	—
—	—	—	20	18	—	—	—	13	—	—	—
—	—	—	28.4	177	—	—	—	25.3	—	—	—
—	—	—	1028	—	—	—	—	436	—	305	—
—	—	—	—	10	—	—	—	—	—	—	—
—	—	—	35.3	10	—	—	—	77.7	—	—	—
—	—	—	—	222	—	—	—	—	—	22	—
—	—	—	41.4	10	—	—	—	28.5	—	8	—
—	—	—	9.1	—	—	—	—	2.2	—	—	—
—	—	—	—	—	—	—	—	—	—	17	—
—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	.6	—	—	—	—	1.8	—	<6	—
—	—	—	—	—	—	—	—	—	—	25	—
—	—	—	—	—	—	—	—	—	—	180	—
—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—

TABLE 2

Chemical compositions of acid oceanic rocks. 1, leucoferrodiorite, Vema fracture zone; clinopyroxene, brown hornblende, actinote, oligoclase, titanomagnetite, magnetite, apatite and zircon. 2, granophyric diabase, Argo fracture zone, Western Indian ocean, (in ENGEL and FISHER, 1975); albitized andesine, K-feldspar, hornblende remplacing augite, apatite and opaque minerals. 3, diorite, Mid-Atlantic Ridge (in AUMENTO, 1969); Plagioclase (An_{40}-An_{20}), K-feldspar, albite, green hornblende after pyroxene and olivine, biotite, magnetite, spinel, apatite, sphene, allanite. 4, diorite, Mid-Atlantic Ridge at 45°N (in AUMENTO, 1969), same mineralogy as sample 3.5, Quartz monzonitic dikelet in sample 2, Argo fracture zone, Western Indian ocean (in ENGEL and FISHER, 1975), quartz, K-feldspar, hornblende, biotite, apatite, zircon and ores. 6, aplite, Argo fracture zone, Western Indian ocean (in ENGEL and FISHER, 1975); quartz, oligoclase, albite, aegyrine, apatite, sphene, zircon, chlorite. 7, aplite, Mid-Atlantic Ridge near 23°N31 at the ridge and fracture zone junction (in MIYASHIRO *et al.*, 1970); albite (5-8% An).

Sample No.	1 CH78 DR10.0029	2 ANTP 125.4b	3 AG 159.35	4 AG 159.39	5 ANTP 125.4c	6 ANTP 125.16	7 E
SiO_2	55.73	57.27	61.97	72.47	75.07	76.37	78.39
TiO_2	1.42	1.76	.94	.33	.15	.42	.09
Al_2O_3	16.41	14.57	16.00	14.17	13.18	12.78	12.68
Fe_2O_3	8.56	1.50	3.22	1.85	.76	.39	.38
FeO		9.57	3.57	1.19	1.15	.46	.41
MnO	.20	.16	.09	.08	.03	.02	.01
MgO	2.91	2.31	2.43	1.39	.23	.87	.54
CaO	5.80	5.94	3.24	1.48	1.10	.84	.55
Na_2O	6.69	5.17	5.55	5.55	4.55	7.70	6.66
K_2O	.36	.72	.75	.24	3.27	.07	.06
H_2O^+	1.90	.44	1.28	.90	.20	.25	.38
H_2O^-		.04			.08	.03	.03
P_2O_5		.32	.22	.06	.12	.02	.01
CO_2		—	<.10	<.10	—	—	—
Total	99.98	99.77	99.26	99.71	99.89	100.22	100.19
Sr	—	150	140	89	18	86	—
Ba	—	160	260	200	120	180	—
Cr	—	6	<20	<20	8	7	—
Zr	—	97	760	200	250	550	—
V	—	63	38	22	5	9	—
Ni	—	9	46	22	<10	29	—
Cu	—	16	24	30	5	5	—
Co	—	17	<20	<20	<10	<5	—
Sc	—	29	<10	<10	<10	10	—
Zn	—	—	52	27	—	—	—
Pb	—	<20	2.2	1.8	<20	47	—
Ga	—	21	32	40	20	21	—
B	—	—	2.2	1.7	—	—	—
Rb	—	—	<30	<30	—	—	—
Nb	—	<20	—	—	29	25	—
Y	—	150	—	—	150	180	—
Yb	—	12	—	—	13	17	—

TABLE 3

REE, Rb, Sr, Th, U, Ta and Hf contents of corsican albitites and oceanic plagiogranites. Corsican samples: 23.8.6.1A; 28.7.3; 25.8.1; (in COCHERIE, 1978). For the mineralogy, see Table 1. Oceanic samples: ANTP 1254b, 1254c (in HEDGE *et al.*, 1979); for the mineralogy, see Table 2. GDP 8.12.1, 8.12.4, 11.7: plagiogranites from the Kyushu-Palau ridge and the Arami Plateau (in ISHIZAKA and YANAGI, 1975).

Sample No.	23.8.6 1A	28.7.3	25.8.1	ANTP 125.4b	ANTP 125.4c	GDP 8.12.21	GDP 8.12.4	GDP 11.17.1
La	11.8	16.2	28.5	—	—			
Ce	43.8	60	92.3	63.9	53.2			
Nd	45.4	50.5	64	49.7	30.3			
Sm	13	15.5	17.1	14.5	7			
Eu	4.76	4.37	1.19	4.17	.616			
Gd	12.5	16	12.9	17.	8.22			
Tb	2.3	3	—	—	—			
Dy	14.4	20.7	10.9	19.2	10.8			
Er	8.29	12	6.74	15.7	7.68			
Tm	1.17	2.05	—	—	—			
Yb	8.46	11.6	7.67	16	8.22			
Lu	1.5	1.65	1.16	—	—			
Σ REE	178	227.	261	200.17	126.036			
Rb	.9	.6	1.8	30.5	89.5	3.16	13.2	4.9
Sr	25	84	—	133	27.7	1.91	217.8	957
$^{87}Sr/^{86}Sr$	—	—	—	.70338	.70418	.70336	.70338	.7032
Th	.28	.84	1.96					
U	.23	.21	.53					
Ta	.80	1.22	1.40					
Hf	35.3	28.4	19.1					

process (OHNENSTETTER *et al.*, 1979). The evolution of the leucocratic rocks testifies to a late fractionation event controlled mainly by plagioclase (Fig. 7).

The high REE (Σ REE ≈ 200 ppm) concentrations (Table 3) as well as the high Zr content, unexplainable solely by a single fractional crystallization model (COCHERIE, 1978) may be due also to an immiscible process. Moreover, similar REE patterns may be observed for Corsican and oceanic leucocratic differentiates according to their petrographic classification (Fig. 8). Flat REE patterns with slight negative Eu anomalies are exhibited by oceanic ANTP 125 4b granophyric diabase (HEDGE *et al.*, 1979) and Corsican 2873 albitite which shows more evident rapid cooling features. Compared to the previous samples, the Corsican 23861A orthocumulate possesses a slight positive Eu anomaly and a HREE depletion. The albite granite 2581 and the quartz-monzonitic dykelet ANTP 125 4c show REE patterns with strong Eu negative anomalies coupled to low HREE. Plagioclase, apatite and zircon

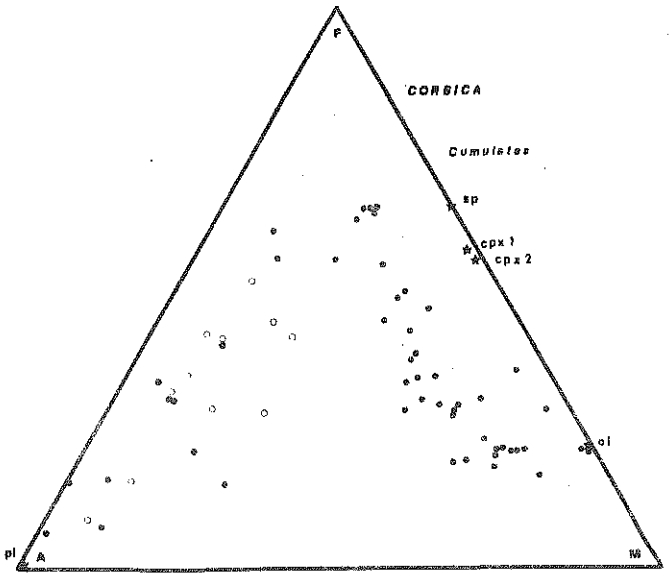

FIG. 3. — **AMF triangle for corsican samples.**
Dot, data after OHNENSTETTER, D. and M. Open circle, data from the litterature (see Table I). Olivine, clinopyroxene, spinel and albite from the Corsican cumulates are also reported.

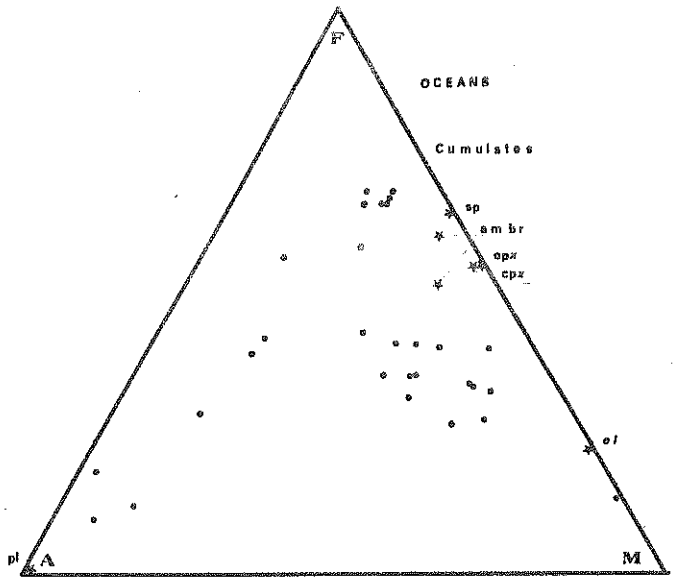

FIG. 4. — **AMF triangle for oceanic samples.** Data from AUMENTO (1969), BONATTI *et* al. (1971); CANN (1971); PRINZ *et al.* (1976); BONATTI *et al.* (1975); ENGEL and FISHER (1969; 1975); MIYASHIRO *et al.* (1970); THOMPSON (1973); OHNENSTETTER, M., unpublished data. Compositions of the plotted minerals are from HODGES and PAPYKE (1976); PRINZ *et al.* (1976); OHNENSTETTER, M., unpublished data.

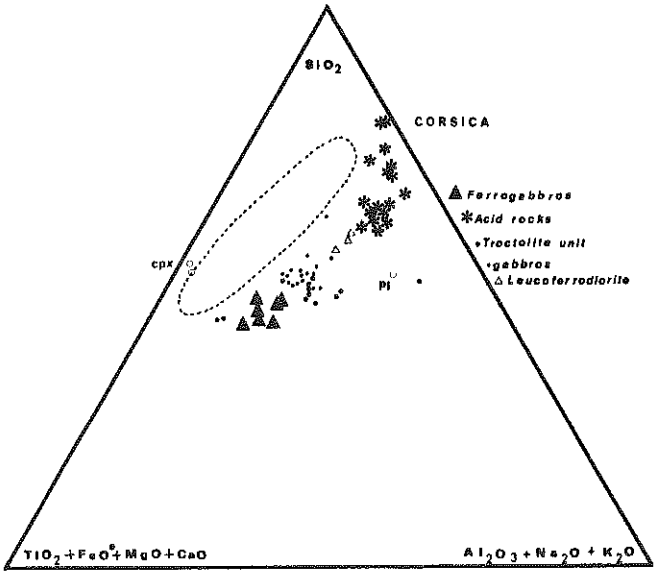

FIG. 5. — **Pseudoternary Greig diagram for Corsican samples.**
The immiscibility envelope is from ROEDDER (1978);
data source is the same as for Fig. 3.

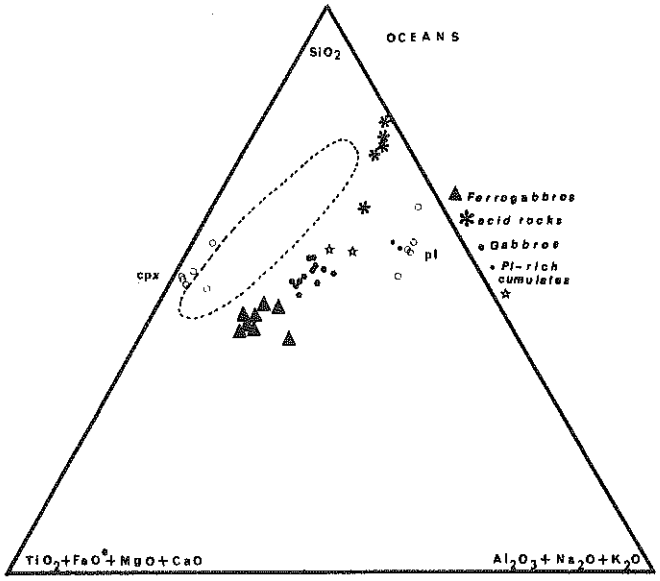

FIG. 6. — **Pseudoternary Greig diagram for oceanic samples.**
The data source is similar as for Fig. 4.

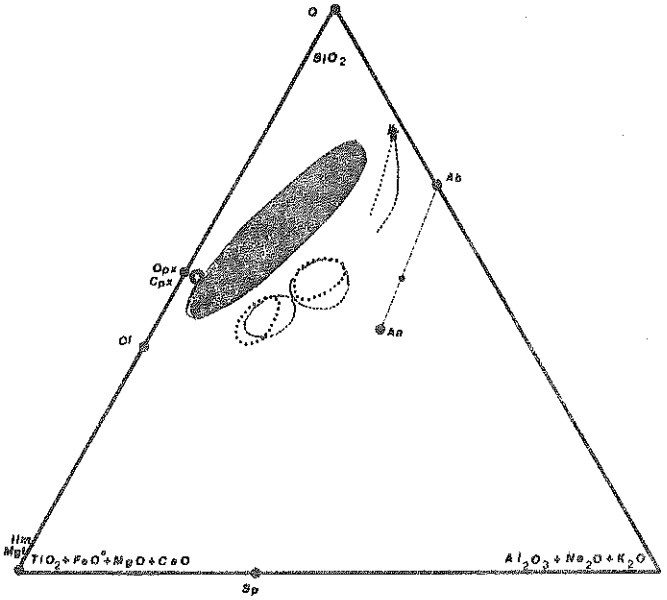

FIG. 7. — **Envelope and evolutionary trend of plutonic rocks from Corsica and the oceans in the Greig diagram:** dot, oceanic samples; dashed, corsican samples.

(NAGASAWA, 1970; WATSON, 1979; 1980) decrease and quartz increase in the final products may be responsible for this pattern. Feldspar and apatite removal has been envisaged to explain the negative europium anomaly in oceanic samples (HEDGE *et al.*, 1979). Consequently, the diversity pointed out for REE distribution is due to various proportions of solid and liquid in rocks at different stages of fractionation after the immiscible process. Hydrothermal fluids at late stage of crystallization may also explain a LREE enrichment (EXLEY, 1980).

V. CONCLUSION

A compilation of dredged acid differentiates has emphasized their scarcity in oceanic environment. Similarly in the Alps they are not frequently described and their occurrence in Corsica is sporadic and mainly limited to the Rospigliani series. Therefore, the genesis of these rocks must testify to a special tectonic environment in the ocean (COLEMAN and PETERMAN, 1975) and the plagiogranite layer must thus

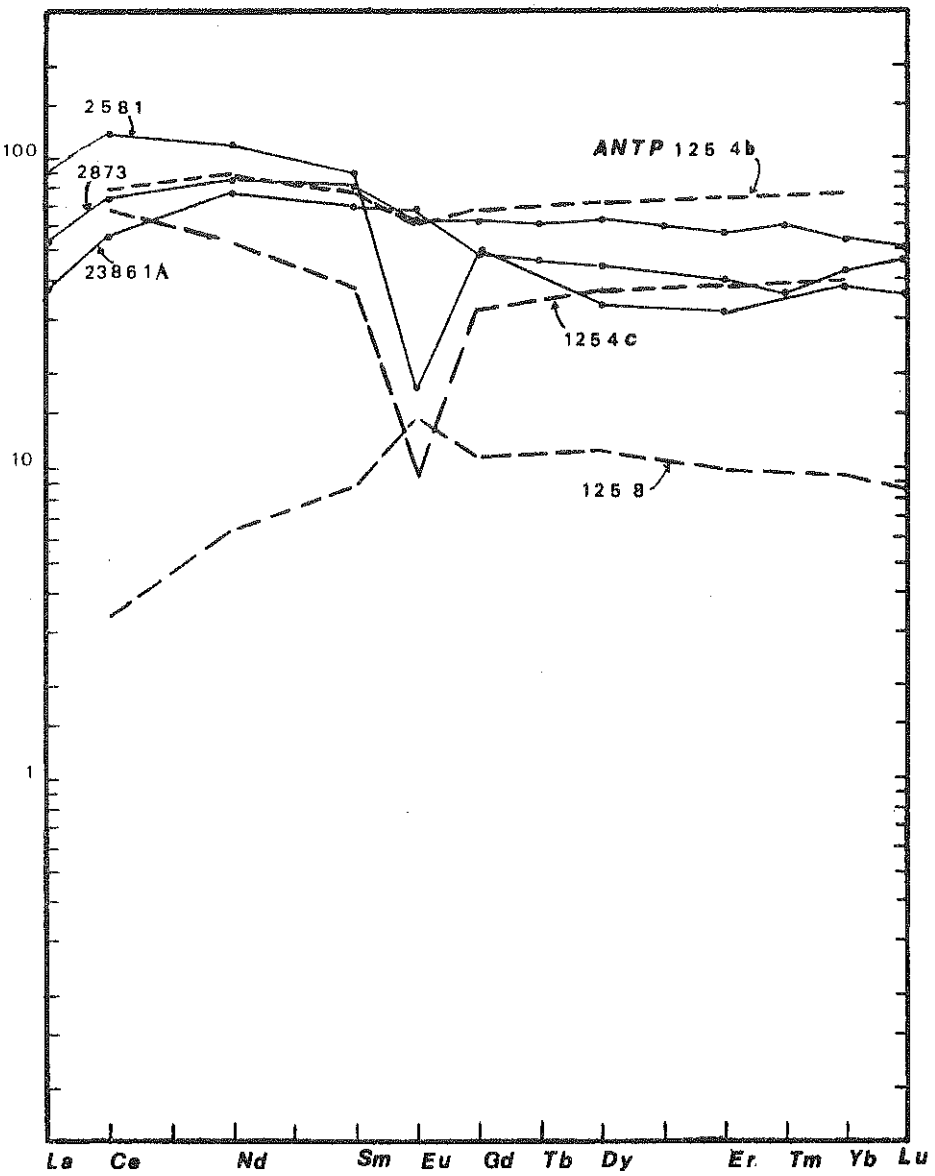

FIG. 8. — **Rare-earth normalized patterns.** Data after COCHERIE (1978) for Corsican samples: 23.8.6.1A and 28.7.3., albitite; 25.8.1, albite-granite. Data after HEDGE *et al.* (1979) for oceanic samples: 12.5.8., Ti-Fe gabbro; ANTP 1254c, granophyric diabase; ANTP 1254b, quartz-monzonite dikelet.

be laterally discontinuous (CHRISTENSEN, 1977). Textural, petrographic and geochemical (major and trace element) comparisons between Corsican and oceanic acid differentiates have stressed the close resemblance between these two suites. Similar magmatic processes are expected to be responsible for their genesis.

An immiscible phenomenon may occur after the crystallization of some ferrogabbros for a residual liquid fraction <35 (BECCALUVA *et al.*, 1977) leading to an iron-rich liquid and a silicic one (OHNENSTETTER *et al.*, 1979) in agreement with experimental work on oceanic tholeiites (DIXON and RUTHERFORD, 1979). The high density iron-rich liquid, giving clinopyroxene, apatite and ore clusters in close association with leucoferrodiorites may form troughs in the underlying cumulates. The low density and viscous silicic magma showing a "peralkaline tendency", rapidly escaping towards the surface participates in the building up of the roof of magma chamber. Albitites (tonalites-trondhjemites) have well recorded this process with congelation cumulates (solid and trapped liquids) and anhedral granular textures obtained during more rapid cooling (COX *et al.*, 1979). They constitute a border group with less evolved dolerites and ferrodolerites which may locally disappear during the latest successive acid injections. Albite granites (plagiogranites) result from the fractionation of this demixed liquid and are injected mostly in dykes showing locally plastic deformation related to their emplacement. Late stage hydrothermal fluids induce cataclasis and brecciation of the acid differentiates. Sulfur deposits may be related to this event. The more or less demixed acid and basic liquids reaching the surface form the agglomerates of the Rospigliani series and may correspond to the evolved liquids of the oceans (e.g. Galapagos and transform fault zones). The rapid dismantling of the upper cumulates and the border group is favoured by rapid uprise of unmixed magmas and by late stage hydrothermal processes. Soon after their formation, these rocks may be incorporated in slight reworked sediments associated with ophicalcites. Erosional features are also induced by serpentinite protrusions linked to oceanic tectonism (LEMOINE, this volume). Therefore, the development of the magma chamber leading to a complete cumulate sequence may have occurred in a dynamic environment compatible with a transform fault zone.

ACKNOWLEDGMENTS

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PLATE I

IA. Albitite. Congelation cumulate: broad plagioclase primocrysts in a subhedral albite-rich matrix with Na-pyroxene (Na-pyr), zircon (zr); opaque minerals and rare apatite and quartz. Inzecca massif.

IB. Albitite. Congelation cumulate showing igneous lamination: euhedral plagioclase laths and sodic amphiboles (crossite rimmed by glaucophane) are separated by interstitial xenomorphous albite and quartz. Accessory phases: opaque minerals, zircon and apatite. Inzecca massif.

IC. Albitite. Mesocumulate: subhedral plagioclases are surrounded by a rim of albite. Aggregates of zircon or mafic minerals, mainly glaucophane (gl) are scattered in the rock; stilpnomelane occupies interstitial areas. Inzecca massif.

ID. Albitite. Subhedral granular texture with large poikilitic albite. Some mechanical twins and suture boundaries are related to slight deformation. Other minerals: magnetite, chlorite, clinozoïsite, sphene and zircon. Casaluna massif.

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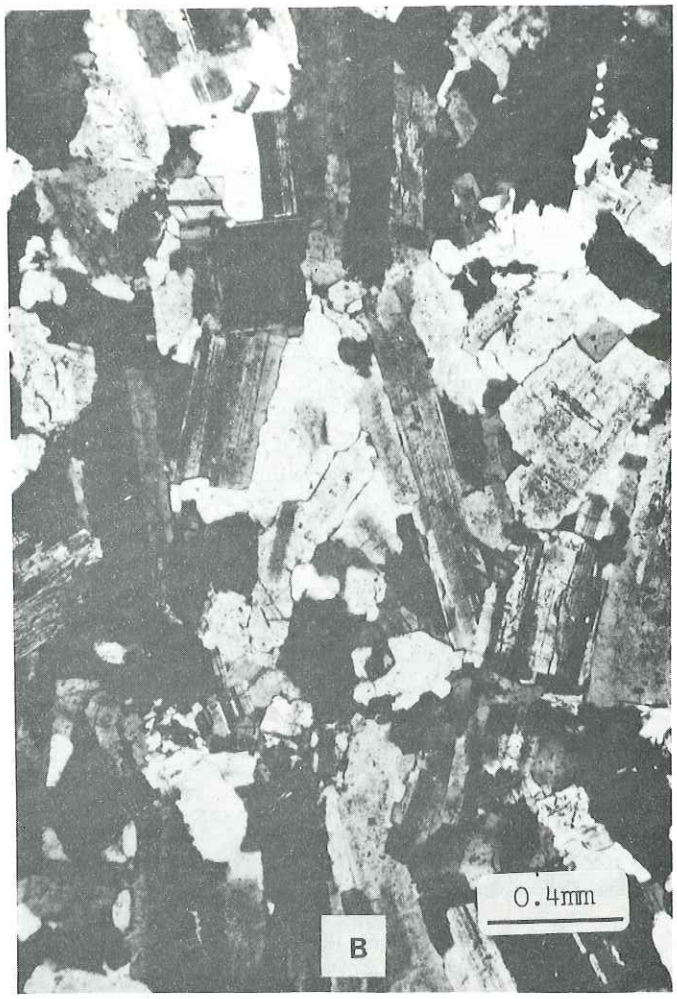
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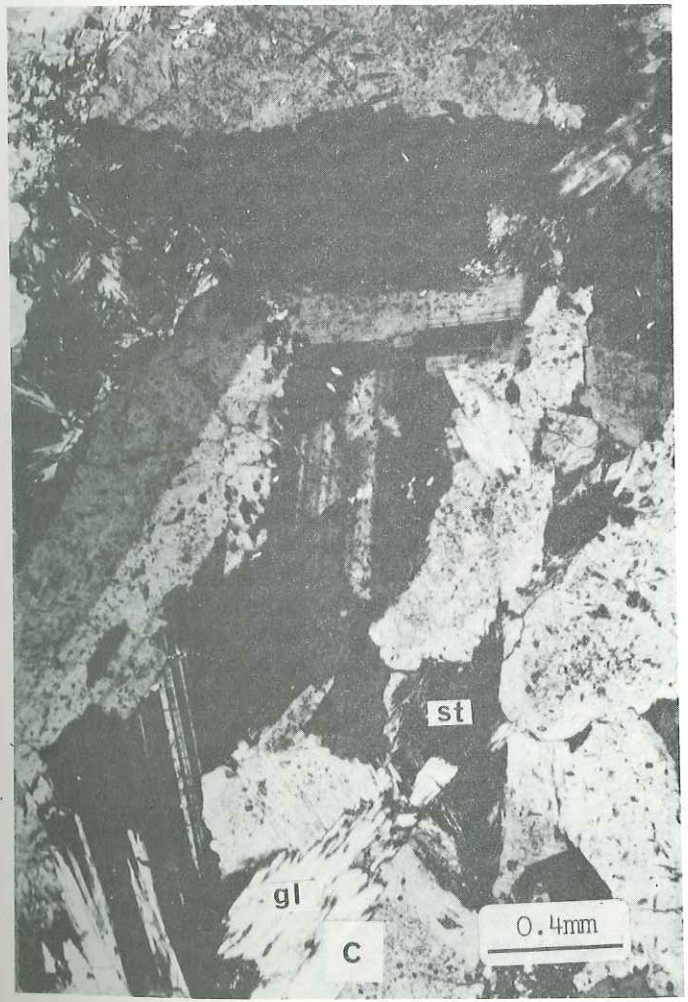
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PLATE II

IIA. Leucoferrodiorite. Orthocumulate. Coarse andesine, clinopyroxene, apatite are cumulus minerals whereas titano-magnetite, ilmenite, sodic plagioclase and some quartz and zircon are intercumulus. Casaluna massif.

IIB. Albitite. Subhedral to anhedral granular texture in a coarser grained albitite with small interstitial area now filled with chlorite and stilpnomelane. Inzecca massif.

IIC. Albite-granite. Xenomorphic inequigranular texture. Sutured boundaries along strained minerals, porphyroclasts of sodic plagioclase and quartz, and local development of mortar texture are evidence of deformation. Opaque minerals, chlorite, oxychlorite and rare zircon. Inzecca massif.

IID. Leucoferrodiorite. Equigranular tabular texture in which large porphyroclasts of plagioclase and clinopyroxene are sometimes present. Opaque minerals, apatite, sodic amphiboles and clinopyroxenes, lawsonite, sphene and chlorite are associated with the strained sub-grains of plagioclase. Inzecca massif.

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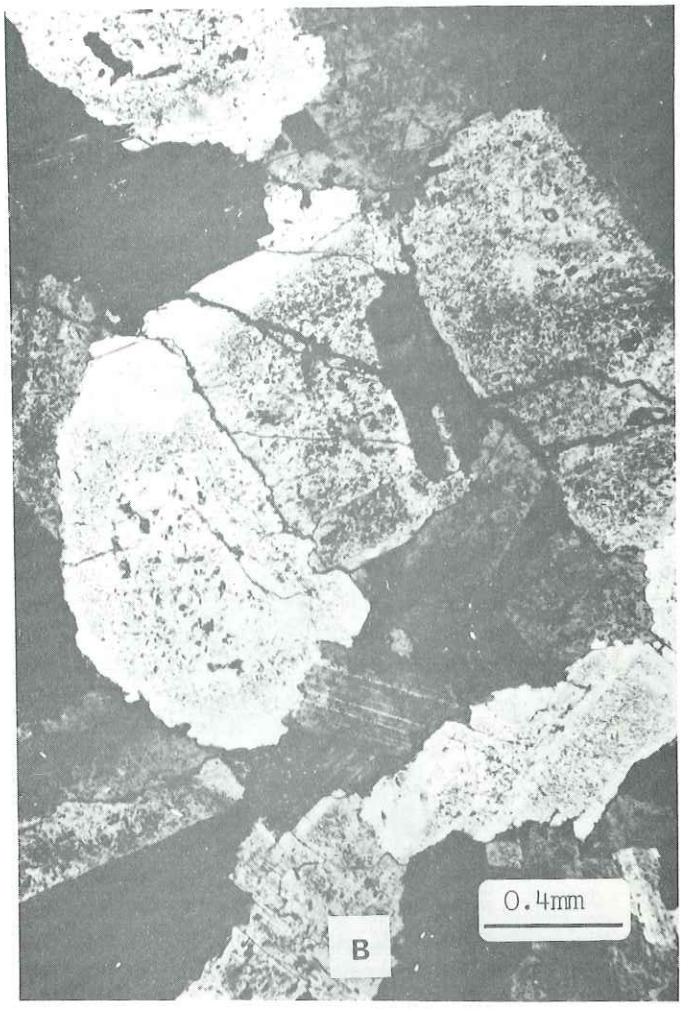
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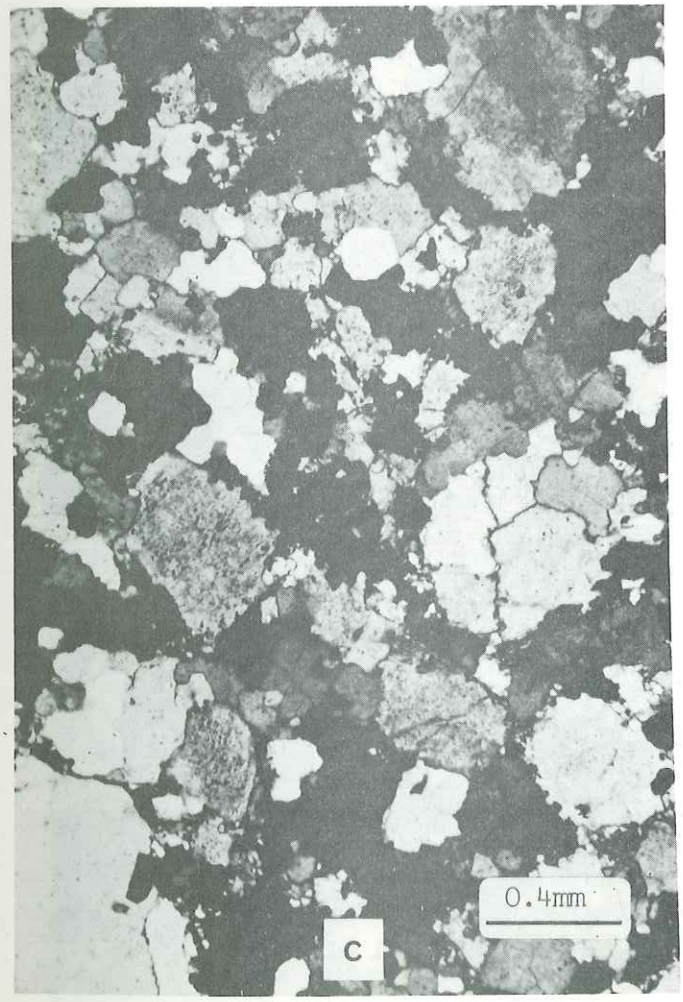
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PLATE III

IIIA. Albite-granite. Mortar texture developped from a xenomorphic inequigranular sample. Abundant albite and quartz associated with glaucophane, opaque minerals, sphene and rare zircon. Inzecca massif.

IIIB. Albitite. Protoclastic texture in a congelation cumulate. Hornblende, sodic amphibole, opaque minerals, apatite and zircon. Chlorite has mainly crystallized in the matrix. Inzecca massif.

IIIC. Albitite. Mylonitic texture formed at the border of a dike. Abundant xenocrysts of gabbro and pyroxene. Epidote, amphiboles, chlorites, opaque minerals and rare zircon. Mte Piano Maggiore, Casaluna massif.

IIID. Albitite-granite. Mylonitic texture locally developped from an anhedral inequigranular one with sub-grains showing curved boundaries. Epidote, sodic amphiboles, chlorites and opaque minerals. Casaluna massif.

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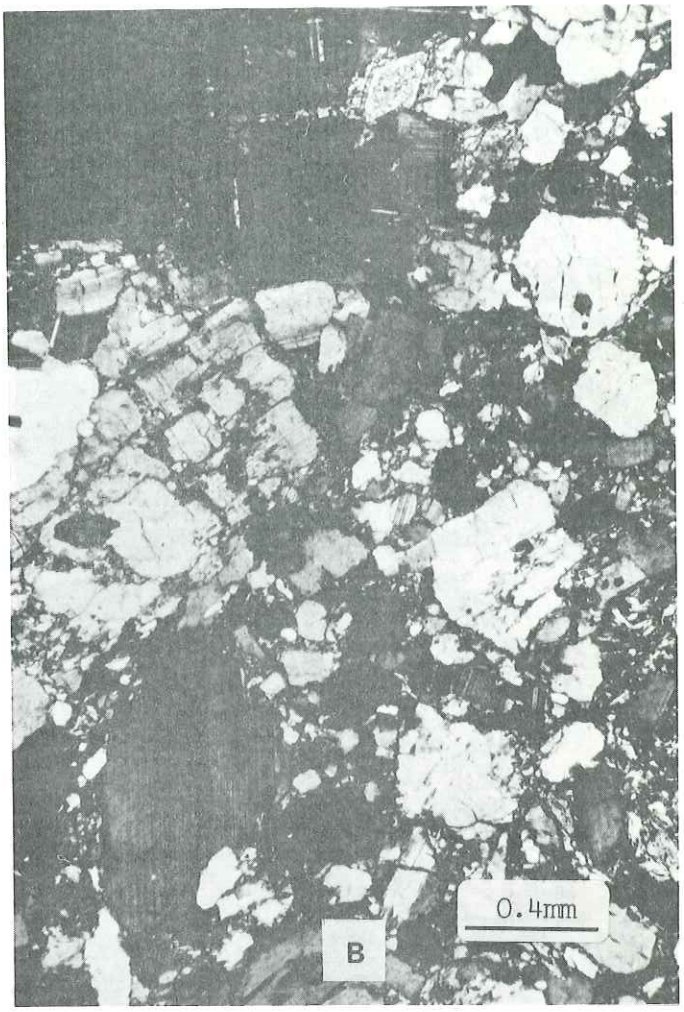
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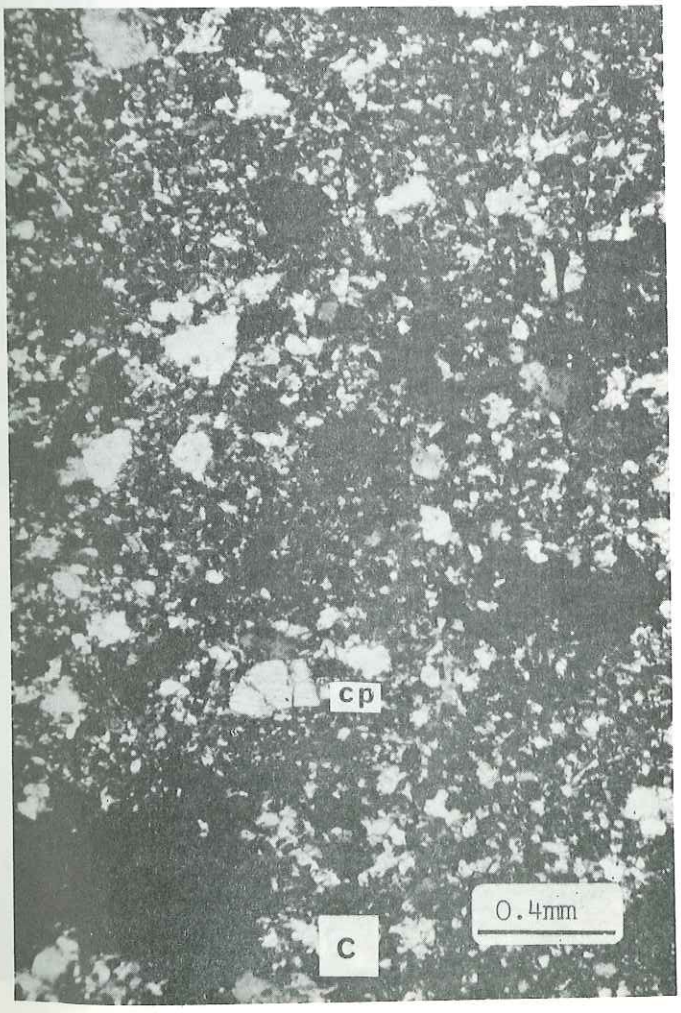
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PLATE IV

IVA. Contemporaneous basic and acid liquids. Radiating plagioclase developped from a trapped more basic bubble. San Cervone, Casaluna massif.

IVB. Contemporaneous basic and acid liquids perhaps related to immiscibility (same sample as for plate IVA). Contorted boundaries between basic and acid magmas. The holocrystalline matrix (albite-granite) shows an intersertal texture and some phenocrysts of plagioclase. Acid bubbles may be seen in the basic liquids.

IVC. Albitite. Local concentration of zircon and opaque minerals. Inzecca massif.

IVD. Albite-granite. Myrmekite around plagioclase laths. Casaluna massif.

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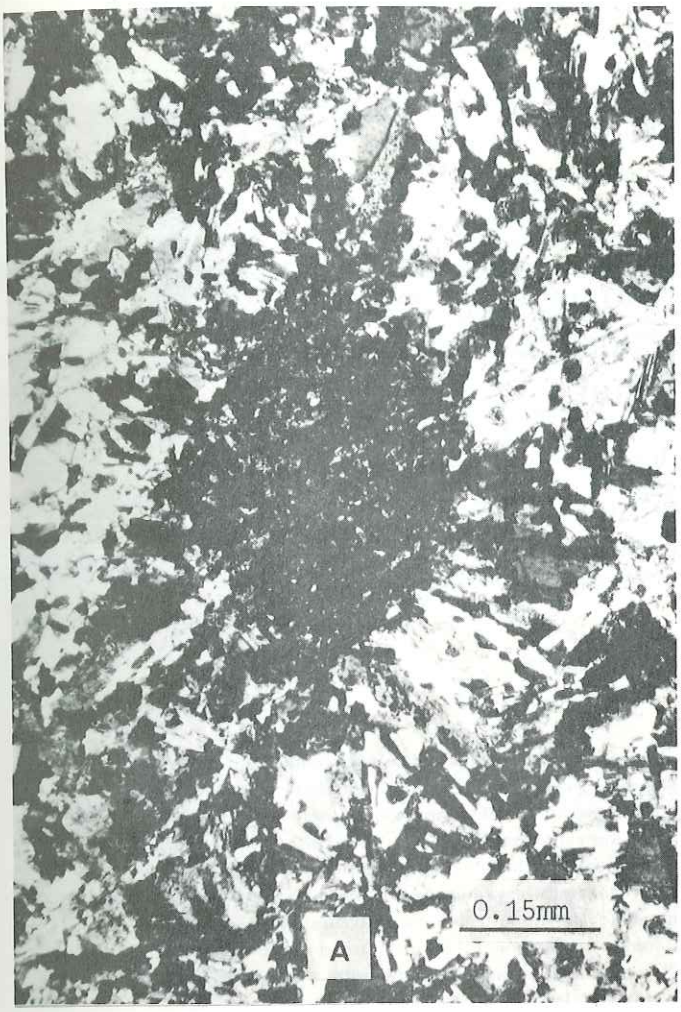
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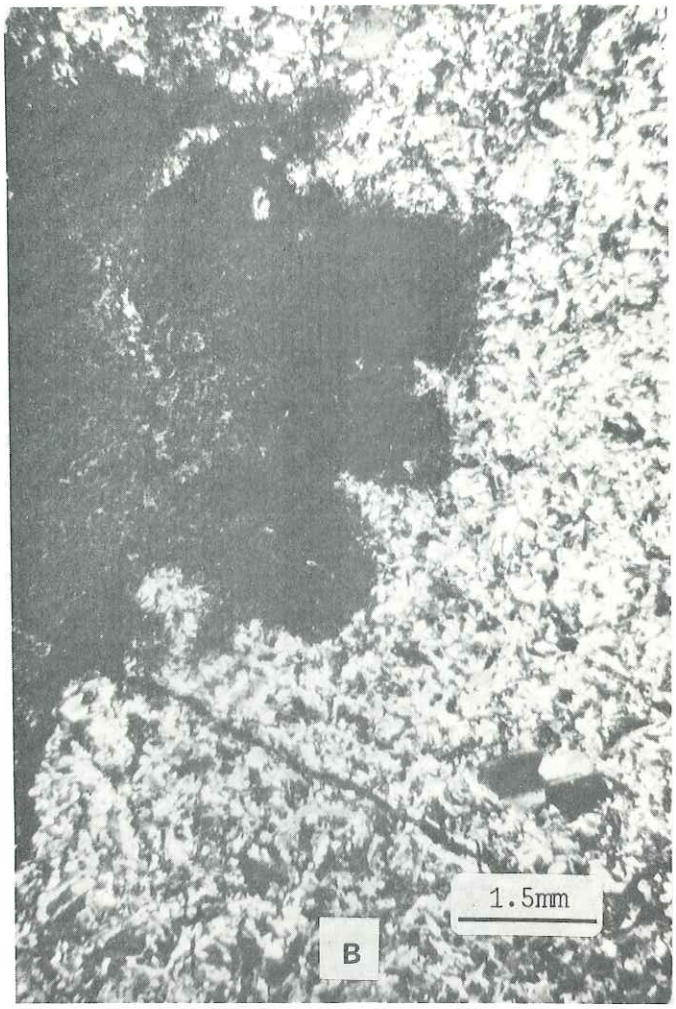
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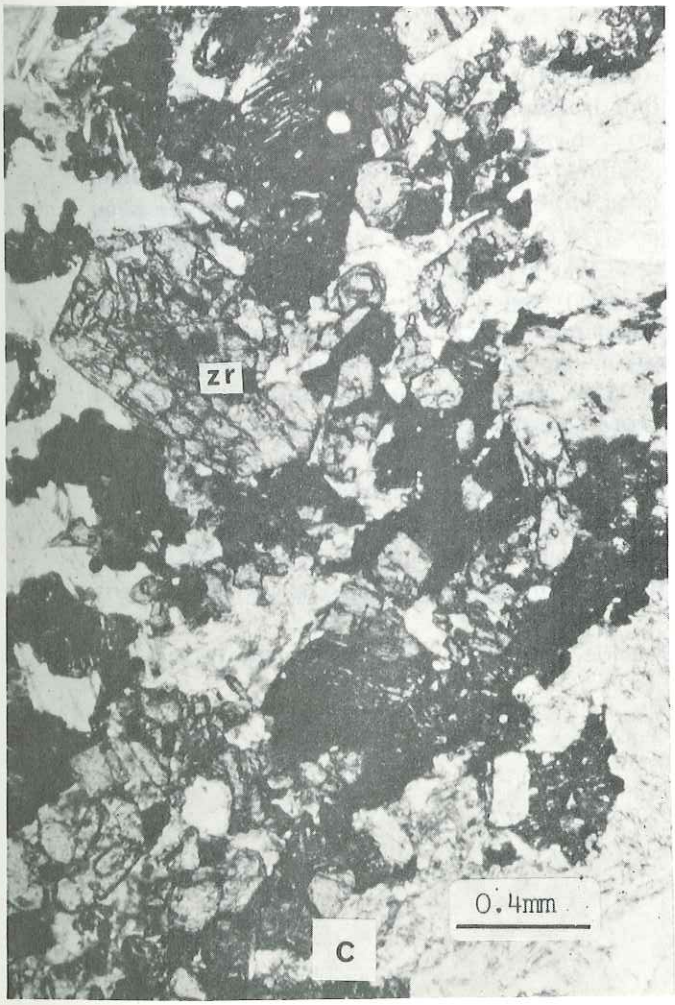
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[1]

Crystallization and age of zircon from Corsican ophiolitic albitites: consequences for oceanic expansion in Jurassic times

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In Corsica, albitites and albite-granites are considered to be the latest products of an ophiolitic magma. Two albitites have been selected for a study of the morphology, typology and age of their zircons, applying for the first time the U-Pb method to Western Tethyan ophiolites. Both samples come from the Rospigliani series, possibly formed in a transform fault zone. The first one shows a cumulate texture, whereas the second has features indicative of more rapid cooling. Zircons have precipitated at the end of the fractional crystallization process after a strong enrichment in alkali and incompatible elements. In a typological diagram they plot at high temperature and high agpaicity indexes. These abundant large, euhedral to subhedral zircons, sometimes with fine zoning, may also contain inclusions of granitic composition which testifies to K fractionation in ophiolitic magmas. The U-Pb age of 161 ± 3 Myr is in the range of the ages obtained for Alpine and Apennines ophiolites. All these ophiolitic remnants belonging to the Liguro-Piedmont Basin may represent a portion of the Central Atlantic Ocean created during Middle and Upper Jurassic times.

1. Introduction

In Corsica ophiolites belonging to the Liguro-Piedmont Basin show close petrographical and geochemical analogies with the present-day oceanic crust [1]. By comparison with the Apennines, metasediments covering lavas are thought to be of Malm age [2]. K-Ar data exist (148 ± 15 Myr) on a brown hornblende from a metagabbro [3] from Balagne, a Liguride unit [4]. Fission tracks in zircon from the Inzecca Massif [5] gave an age of 43.7 ± 0.2 Myr, corresponding to the end of the latest Alpine tectonometamorphic phase. As K-Ar ages on amphiboles and micas, and fission tracks on zircons and apatites, may be partially or totally reset by subsequent thermal events, the age of the ophiolites in Western Tethys is not firmly established. The occurrence of zircons in acidic residues from Corsica is a good chance to obtain the age of crystallization of these minerals by using the U-Pb isotopic systematics. Moreover, morphological and typological studies were also undertaken to determine the conditions of crystallization of the subhedral zircon sometimes reaching 1% during the latest magmatic stage.

2. Geological setting and sample description

In eastern Corsica, ophiolite slices are composed of serpentinized lherzolites, troctolites, gabbros, ferrogabbros, ferrodiorites, albitites, albite-granites, dolerites and mainly pillow lavas [1]. Two main sedimentary covers [6] show lateral variation depending on the distance from the passive continental edge [7]. The ophiolitic sections exhibiting important oceanic tectonism may have been formed on a mid-oceanic ridge and in a transform

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fault. Most of the large gabbro massifs and lava flows are more or less devoid of acid rocks which form only about 1% of the ophiolite. Acid rocks are best developed in the Rospigliani Series, possibly created in a transform-fault zone [8] where leucoferrodiorite, albitites and albite-granites have participated building up the roof of the magma chamber in a dynamic environment. Their textures are characteristic of accumulation processes and/or rapid cooling, locally obliterated by late-stage hydrothermal and cataclastic events [9].

The rare magmatic phases still preserved are Ca-rich clinopyroxene and abundant accessory minerals such as allanite, apatite and zircon. These later phases may be responsible for the extreme-incompatible and rare earth element enrichment and distribution [10,11], unexplained by a single model of fractional crystallization [11]. Alternatively, immiscibility has been proposed [9,12,13] releasing a Si- and Na-rich magma with a "peralkaline tendency" [9]. Secondary parageneses belong to three metamorphisms [1]. The blueschist facies is particularly imprinted in these rocks because of their high Na and Si contents. Crossite rimmed by glaucophane and ferro-glaucophane, aegyrine-augite rimmed by jadeitic pyroxene surrounded by a second generation of aegyrine, are associated with stilpnomelane, chlorite, pumpellyite and lawsonite. In some places greenschist facies may be recognized before and/or after the high pressure phase.

Two albitites were chosen for dating. The first one (I) was sampled from a small mass resting tectonically on serpentinites and ophicalcites near Punta Corbara. It is a fine-grained orthocumulate showing a hiatale texture. Euhedral plagioclase porphyroids with fine albite and Carlsbad twins produce the igneous lamination. Other smaller randomly oriented subhedral feldspars are separated by interstitial anhedral untwinned albite and quartz. The proportion of scattered Na-pyroxene varies. When they present a cloudy greenish core of aegyrine augite bordered by jadeitic pyroxene or Na-amphibole, they may be relicts after magmatic clinopyroxenes. Elsewhere, jadeitic pyroxenes correspond to the breakdown of plagioclase. The small opaque minerals sometimes observed in plagioclase primocrysts are more generally associated with clinopyroxene relicts. Apatite is rare and some biotite flakes have been observed. Abundant (about 1%) euhedral zircons 0.1–0.2 mm in size are scattered through the plagioclase matrix where they may be in contact with opaque minerals. They possess some cavities and inclusions (see later).

The second sample (II) was taken from the acid matrix of a magmatic breccia (near Vezzani). In some places the coexistence of two rapidly quenched magmas is evident. The rock is medium grained and possesses a subhedral equigranular texture. When coarser grained it becomes pegmatitic. Coarse and cloudy, subhedral to anhedral, randomly oriented plagioclase laths are bordered by a narrow rim of pure albite. Rare interstices are occupied by stilpnomelane, oxy-chlorite, chlorite and sometimes albite. Small apatite grains are associated with titano-magnetite transformed to sphene. Zircon is less abundant and the crystals are smaller than in the previous sample but it may be highly concentrated in some places. It is corroded, broken and rich in inclusions, perhaps linked to its rapid growth.

3. Morphology, typology and U and Th contents of zircon

3.1. Morphology

Various morphological characteristics observable under the microscope have been recorded. The study reveals:

- the preponderance of euhedral and subhedral grains
- elongation ratios usually between 2 and 2.5, rarely greater than 3
- a weak beige coloration in the zircons
- at least 21% of all zircons with finely developed zoning (Fig. 1B)
- some growth-interference structures
- the existence of crystalline aggregates of the prismatic or multiple type (2% of the total population)
- an abundance of inclusions of different types (Fig. 1A and B): light- and dark-coloured rods connected by menisci; tubes oblique to the $\vec{c}$-axis

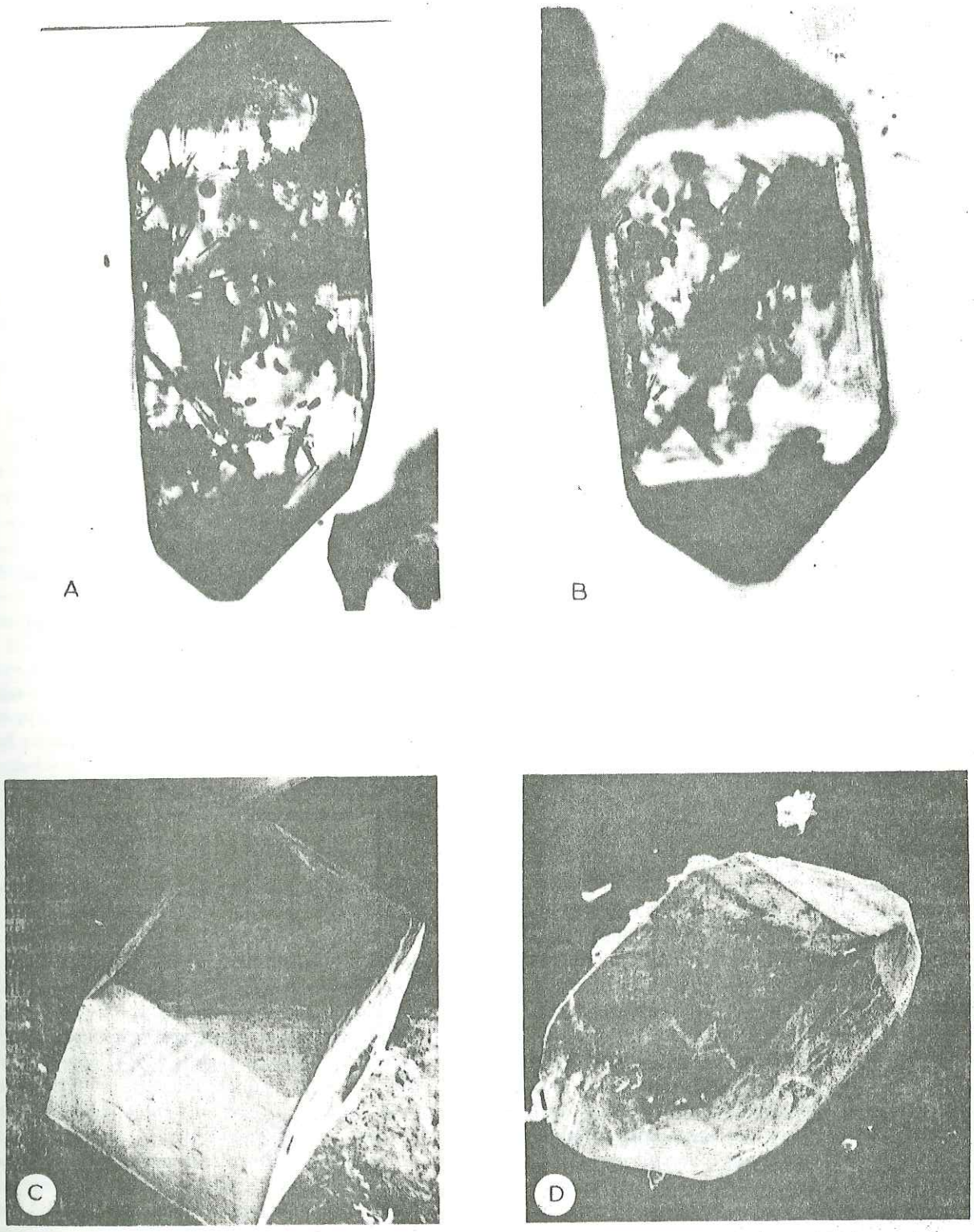

Fig. 1. Crystals of zircon. Each crystal is about 0.1 mm. A. Tube oblique to $\vec{c}$ axis and irregularly shaped inclusions. B. Irregularly shaped inclusions and fine zoning. C, D. Scanning electron microscope (SEM) photographs of euhedral zircon. Note the development of {101} and {301} forms.

TABLE 1

	A	B	C	D	E
SiO_2	32.54	32.40	75.85	71.9	69.68
Al_2O_3		–	15.15	16.3	12.41
FeO^{total}		0.18	0.06	0.59	5.09
MnO	–	0.09	0.03	n.d.	n.d.
MgO	–	0.01	–	0.21	0.40
CaO	–	–	0.63	2.92	2.85
Na_2O	–	–	6.30	6.94	4.29
K_2O	–	–	1.07	0.28	1.47
TiO_2	–	–	0.01	0.06	0.61
P_2O_5	0.28	0.20	n.d.		0.07
ZrO_2	67.30	67.30	n.d.		
Total	100.12	100.18	99.10	99.20	96.87

Analyses of zircons (A and B) and inclusion in zircon of sample I (C). The electron probe Camebax of the University of Nancy I was employed with the following conditions: an acceleration voltage of 15 kV, a beam current of 8 nA, a beam diameter of less than 3 μm, a six-second counting time for a group of three elements. The corrections were obtained through the MBXCOR program of Henoc and Tong [57]. Y and Th have not been detected with the microprobe. Analysis D: granophyric patch, Atlantic Ocean [30]; E: rhyodacite, Galapagos [28].

filled with amorphous material; two-phase inclusions; light or dark irregularly shaped inclusions and "bubbly" inclusions. An inclusion analyzed with the microprobe presents a granitic composition (Table 1).

3.2. Typology

Crystal faces were indexed following the method of Caruba and Turco [14]. When faces are plotted in the typological classification diagram of Pupin and Turco [15] and Pupin [16], crystals are assigned to various types and sub-types according to the relative growth of prismatic (I T-index) or pyramidal forms (I A-index). The development of pyramidal faces appears related to the agpaicity (I A-index) of the rock, whereas prismatic faces are favoured by increased temperature of crystallization (I T-index). The zircons from albitites have the following characteristics:

— weighted mean co-ordinates $I\ \bar{A} = 624$; $I\ \bar{T} = 785$ (Fig. 2)

— a high population density centred around sub-type J_5 and D (monoprismatic {100} forms) corresponding to a high temperature of crystallization (around 900°C) using geothermometric correlations [17]

85% of indexed zircons with monoprismatic {100} (Fig. 1C and D) faces

— very little scattering of types and sub-types, especially along the I T-index axis

— abundant development of {101} and {301}

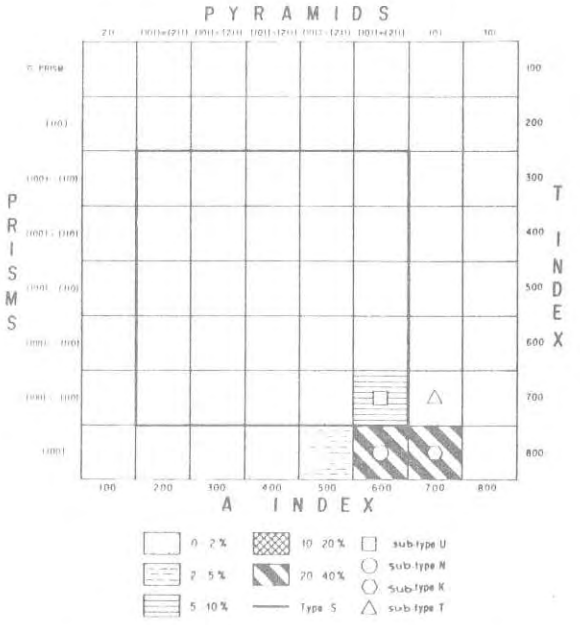

Fig. 2. Typological diagram of Pupin and Turco [15]. T-index: index of temperature; A-index: index of agpaicity.

forms (Fig. 1C and D) amongst N, K and T-type zircons. 36% of the indexed population sub-types U, N, K, and T) show {301} forms indicating a high agpaicity.

A preliminary study of zircons [16,18] from a Corsican trondhjemite has revealed analogies with zircon from rhyolites and granites from anorogenic complexes of the Permian alkaline province of Esterel and Corsica, suggesting high temperature, and high agpaicity and low water content [19,20] in magmatism related to oceanic rifting.

3.3. U and Th concentrations

The zircons are pure under microprobe (Table 1); U and Th contents are extremely low, at least ten times lower than the average concentration of zircons in acid rocks. Using mass spectrometry (isotope dilution) U averages 40 ppm, and Th (calculated through the $^{208}Pb/^{206}Pb$ ratio which gives a Th/U ratio of 0.85) averages 34 ppm. These contents are similar to those in kimberlite [21–23].

In contrast, a zircon in an ophiolitic albitite from the Chenaillet Massif, Alps, shows higher values of uranium and thorium, of 284 ppm and 166 ppm respectively [24,25]. This reflects the general paucity of these elements as well as that of potassium in basic magmas from ophiolitic suites [26], and even in the more differentiated liquids, where other incompatible elements may be highly concentrated [1,10].

3.4. The significance of zircon in the ophiolitic differentiates

Rare acid differentiates in western ophiolites contain relatively abundant zircons, especially in albitites. The euhedral zircons are larger in cumulates (congelation cumulate of Wager and Brown [27] in sample I) than in more rapidly cooled rocks (sample II), as testified by their skeletal habit. They have cristallized in residual liquids issued from fractional crystallization of an olivine-tholeiitic magma. The high concentrations of incompatible elements in these evolved liquids have been favoured by adcumulus growth in most of the gabbroic sequence [1,10]. Later immiscible processes between iron-rich and high silicic and sodic liquids [9,12,13] have caused a higher Zr concentration in the latter one which shows a "peralkaline tendency" even though the rock possesses an agpaicity coefficient less than 1. This character permits late precipitation of zircon to occur when silica is in excess (quartz is often present in zircon-bearing rocks) after plagioclase and clinopyroxene fractionation. Zircon may contain inclusions of granitic composition (sample I), similar to some of the evolved rocks in the ocean (see Table 1), e.g. Galapagos [28], East Pacific Rise [29], Atlantic Ocean at 25°N [30] and Indian Ocean [31]. These inclusions may testify to the concentration of K_2O in the residual liquid and its escape by late magmatic vapour transport [30].

4. Dating of the zircon

4.1. Crushing and separation

Two lots of albitite weighing about 50 kg each were studied. For sample II a circular saw was used to cut out the leucocratic part. The samples were crushed in jaw crushers and drum rollers and sieved until all fragments belonged to the 250–125 mesh fraction ($\sim$0.1 mm). A good separation of the heavy minerals from the bulk sand was achieved using a panning method in the laboratories of the "Usine Pilote of Nancy". Magnetic methods allowed the separation of zircon from aggregates and from zircons with inclusions. The light fragments were removed in potassium iodomercurate ($d = 3.17$) liquid. The zircons were then purified using a Frantz electromagnetic separator.

4.2. Chemical procedures

The chemical methods used for the dissolution of zircon and the extraction of uranium and lead are based on those of Krogh [32]. Zircon was washed in hot 8*N* nitric acid for 30 minutes, then dissolved in a hydrofluoric nitric acid mixture in a bomb. Sample weights were of the order of several tens of milligrams. Separation of lead and uranium was carried out in ion-exchange columns packed

with Bio Rad anionic resins (AG1 × 8, 200–400 mesh). One sample of a zircon separate (I-3) was also analyzed at the Department of Geological Sciences of the University of California at Santa Barbara (Ph. Vidal), following the procedure of Chen [33]: apart from the use of a ^{205}Pb spike, the method was very similar to that given above.

Triple distilled water, from Heraeus stills, followed by sub-boiling distillation in Quartex, is used for washing the apparatus and for the preparation of reagents. The sub-boiling distillation in FEP teflon bottles leads to very low reagent blanks. Partial blanks obtained on the dissolution bombs are better than 300 pg, while column blanks and loading blanks amount to about 100 pg and 10 pg, respectively. The total reagents blanks amount to about 100 pg. Several full analytical blank determinations were performed during the course of the analyses; they all fall between 0.5 and 1 ng. Isotopic ratios for two of these complete blanks were as following: $^{206}Pb/^{204}Pb$: 17.67 and 17.67; $^{207}Pb/^{204}Pb$: 15.48 and 15.49; $^{208}Pb/^{204}Pb$: 37.49 and 37.40.

4.3. mass spectrometry

Measurements were performed on a Cameca TSN 206 SA mass spectrometer (30 cm radius, accelerating voltage 10 kV). Lead was run by loading samples with silica gel–phosphoric acid onto rhenium filaments. Uranium was run at Rennes on oxidized tantalum filaments, whereas at Santa Barbara samples were prepared in a phosphoric acid–graphite mixture and loaded onto rhenium filaments. The signal is detected by an electron multiplier and measured on a HP 5328 A digital voltmeter. All measurements and data reduction are programmed by an on-line HP 9825 calculator followed by a HP 9871 A printer.

The apparatus is checked regularly by the analysis of NBS lead and uranium standards. Values obtained are in close agreement (less than 1‰ per mass unit) with those of Catanzaro et al. [34] so no correction factor has been applied to the present data.

The within-run errors (2 σ_m) are about 1% for the $^{206}Pb/^{204}Pb$ ratio and 1‰ for the $^{207}Pb/^{206}Pb$ ratio.

Uranium and lead concentrations were determined by isotopic dilution after adding a mixed ^{235}U-^{208}Pb spike to sample aliquots taken after dissolution (except at Santa Barbara, where a mixed ^{235}U-^{205}Pb spike was added before dissolution). The U and Pb concentrations of the spikes were determined by calibrating against standard solutions freshly made up from NBS 983 and specpure Johnson-Matthey leads salt and a uranium salt provided by the French Atomic Energy Commission (CEA).

The isotopic compositions of the Oak Ridge spike was not found to be significantly different from recommended values, so these certified values were used in all calculations. The isotopic composition of lead in the Johnson Matthey salt was measured as: $^{204}Pb = 0.0141$, $^{206}Pb = 0.2439$, $^{207}Pb = 0.2185$, $^{208}Pb = 0.5235$. The concentration of ^{208}Pb in the spike was determined by calibrating against the salts NBS 983 (on three dilutions: 1.6859 ± 0.0030 $\mu g/ml$) and Johnson-Matthey (on five dilutions: 1.6802 ± 0.0040 $\mu g/ml$).

The isotopic composition of uranium in the uranium spike was: ^{234}U, 0.0117; ^{235}U, 0.9785; ^{238}U, 0.0098. The concentration of ^{235}U in the spike was determined by calibrating against a CEA standard salt (on five dilutions: 4.770 ± 0.017 $\mu g/ml$).

The uncertainties in standard concentrations, added to the known precision of measurements, has led to an estimate of 1% error on the U/Pb ratio.

The uranium decay constants adopted in this study are: $\lambda(^{238}U) = 0.155125 \times 10^{-9}$ yr^{-1}; $\lambda(^{235}U) = 0.98485 \times 10^{-9}$ yr^{-1}.

4.4. Interpretation of $^{207}Pb/^{206}Pb$ and U/Pb ages

The radiometric dating of the relatively young zircons presents rather specific problems. In particular, the uncertainties of the ages calculated from the $^{207}Pb/^{206}Pb$ ratio is very large and this is due to three factors.

(1) The almost linear nature of the Concordia curve in this age range. An error of 2‰ in the $^{207}Pb/^{206}Pb$ ratio leads to an error of more than 3% in the $^{207}Pb/^{206}Pb$ age at 150 Myr, whereas the same analytical error influences an age of 2000 Myr by only 0.2%.

TABLE 2

Variations in the $^{207}Pb/^{206}Pb$ age vs. measured $^{206}Pb/^{204}Pb$ ratio

$^{206}Pb/^{204}Pb$	$^{207}Pb^*/^{206}Pb^*$	$^{207}Pb/^{206}Pb$ age (Myr)	$^{206}Pb/^{238}Pb$ age (Myr)	$^{207}Pb/^{235}U$ age (Myr)
538	0.050426	215	159	163
530	0.050004	195	159	161

* Designates radiogenic lead.

(2) The error of the $^{206}Pb/^{204}Pb$ ratio has a very noticeable effect, which can be seen by taking the example of zircon fraction I.2 in Table 2, where $^{206}Pb/^{204}Pb = 534 \pm 4$. $^{207}Pb/^{206}Pb$ ages are very sensitive to small changes due to analytical errors in the $^{206}Pb/^{204}Pb$ ratio, but the two U/Pb ages are hardly affected at all.

(3) The influence of the correction for common lead. Common lead is made up from (a) contaminating lead (laboratory blank), (b) lead from silicate inclusions and (c) primary lead fixed in the zircon structure.

The isotopic composition of leads (b) and (c) can be estimated by measuring the lead of an associated feldspar. Feldspar from the same rock as the zircon give: $\alpha = 18.02 \pm 0.03$, $\beta = 15.47 \pm 0.05$.

Since the μ values are very small in the feldspar (U = 0.0185 μg/g, Pb = 1.24 μg/g, $\mu = 0.93$), the observed ratios can be considered as practically the same in the primary lead. One can also indirectly estimate the composition of primary leads by referring to the normal terrestrial growth curve [35–37]. Depending on which model is chosen, α falls between 18.0 and 18.8 and β between 15.4 and 15.66 for the age range under consideration (100–200 Myr). The choice of different compositions for the lead correction has less and less effect as the $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ ratios increase. In the present case, however, $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ ratios are small because of the relatively young age and the very low uranium contents of the zircons. By taking the same example as in the Table 2, where $^{206}Pb/^{204}Pb = 534$, it can be shown that variations in the isotopic ratios assumed for the common lead correction have hardly any effect on the calculated U/Pb ages, whereas $^{207}Pb/^{206}Pb$ ages are changed by as much as 5% (see Table 3).

The isotopic compositions of primary lead and of the lead blank are so similar in the present case that any possible variations would have no effect on the U/Pb ages. Common lead is thus corrected for by using the isotopic ratios found in the laboratory blanks: $\alpha = 17.67$, $\beta = 15.48$. The zircon lead analyzed at Santa Barbara is corrected with a different blank isotopic composition: $\alpha = 19.0$, $\beta = 15.6$.

TABLE 3

Variations in the $^{207}Pb/^{206}Pb$ age vs. common lead correction

Isotopic ratios for the non-radiogenic lead correction	$^{207}Pb^*/^{206}Pb^*$	$^{207}Pb/^{206}Pb$ age (Myr)	$^{206}Pb/^{238}U$ age (Myr)	$^{207}Pb/^{235}U$ age (Myr)
$\alpha = 17.67$ $\beta = 15.48$	0.05014	204	159	162
$\alpha = 18.50$ $\beta = 15.60$	0.04997	194	159	161

* Designates radiogenic lead.

TABLE 4

Uranium and lead concentrations, U-Pb and $^{207}Pb/^{206}Pb$ ages for the different fractions of the two samples

Samples (I,II) fractions (1, 2, 3)	Observed ratios			$^{206}Pb^*/^{238}U$ age (Myr)	$^{207}Pb^*/^{235}U$ age (Myr)	$^{207}Pb^*/^{206}Pb^*$ age (Myr)	Contents (ppm)	
	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{206}Pb$	$^{208}Pb/^{206}Pb$				Pb *	U
I.1 > 149 μm "non-magn"	871	0.06611	0.2938	162	163	166	0.939	32.7
I.2 > 149 μm "magn"	534	0.07748	0.3006	159	162	205	0.986	35.6
I.3 > 149 μm "magn"	501	0.07841	0.3038	162	162	160	1.315	46.2
II.1 > 149 μm "non-magn"	846	0.06611	0.2938	161	162	179	1.125	40.3
II.2 > 149 μm "magn"	570	0.07510	0.2976	159	160	175	1.266	45.5
II.3 > 149 μm "non-magn"	648	0.07186	0.2511	142	143	165	1.043	43.5

* Designates radiogenic lead.

4.5. Results

Data for the non-magnetic zircon fraction are presented in Table 4. I and II refer to the two rocks samples and 1, 2, 3 identify different magnetic or grain size (> 149 μm; < 149 μm) fractions.

$^{206}Pb/^{204}Pb$ ratios of the Corsican ophiolitic zircons fall between 500 and 900. As outlined above, three separate factors can explain the spread

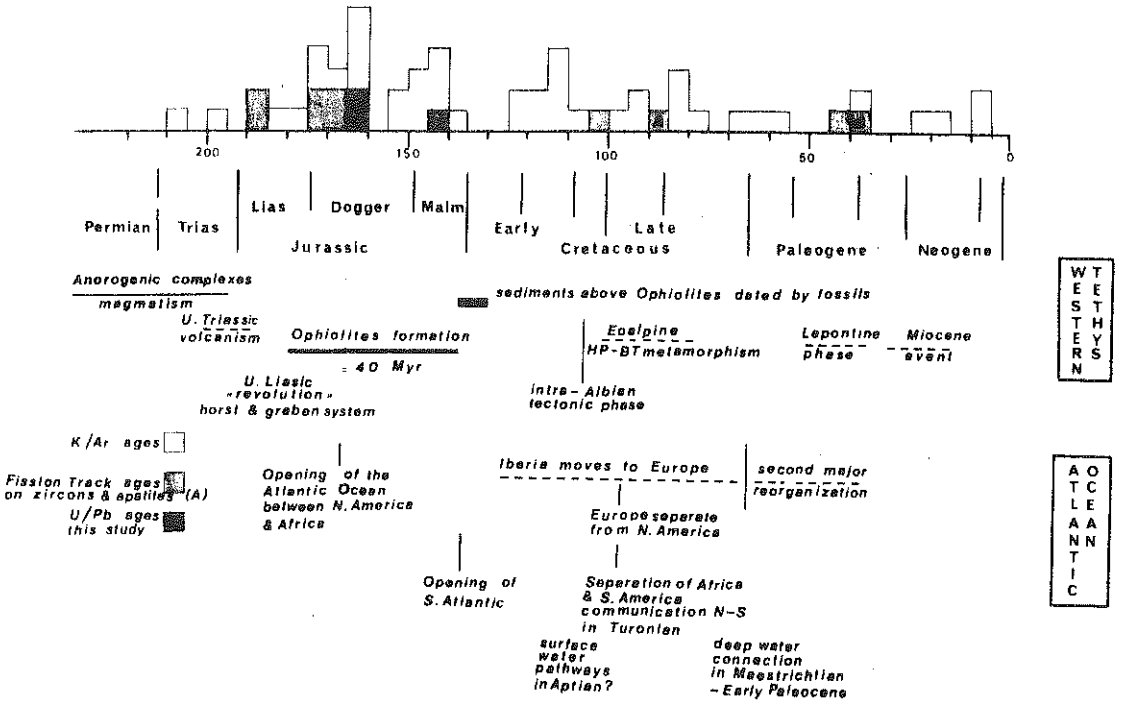

Fig. 3. K/Ar, U/Pb and fission track ages of western Tethyan ophiolites. Data from references 3,5,24,25,40,41,42,43,44,45,46. Data for the Atlantic Ocean are taken from Sclater et al. [58] and Van Andel et al. [59]. Times scales for the Jurassic and the Cretaceous are from Van Hinte [39,60] Fossils dated above the ophiolites in the Ligurian domain (Balagne, Corsica and the Apennines) are of Portlandian/Berriasian age [47–49,61]: *Saccocomidea?*, and *Stomiosphera moluccana* Wanner have been found. In calcarenites interbedded in radiolarian cherts followed by *Calpionella* limestones with *Calpionella alpina* Lorenz; *Calpionella elliptica* Cadish and *Calpionella oblonga*.

of $^{207}Pb/^{206}Pb$ zircon ages between 160 and 205 Myr. An age of 190 ± 25 Myr [38] was considered significant on the basis of preliminary analyses but it was relying on $^{207}Pb/^{206}Pb$ ratios. It is now proposed that the concordant U/Pb ages (or subconcordant ages as with fraction I.2) should form the basis for new age interpretations (Fig. 3).

The very low concentrations of uranium in the zircons make continuous lead loss, or even episodic lead loss during Alpine metamorphisms, a very unlikely possibility. Thus the crystallization age of the zircons is most likely 161 ± 3 Myr *. It corresponds to mid-Jurassic times ([39], and G.S. Odin, written communication).

5. The Liguro-Piedmont basin may be a remnant of the Central Atlantic Ocean in Jurassic time

In the Mediterranean area no U-Pb dating of the relatively recent ophiolite formation existed. In the western part, the main dates obtained by the K-Ar method [3,24,25,40–44] gave, due to late metamorphism, a wide spread of values in the range 180–38 Myr, with two main maxima [42] at 180–165 and 150–135 Myr, the former being the more important (Fig. 3). Fission tracks on zircon [5,24,25,45,46] gave a more restricted range of values between 185 and 142 Myr with a maximum around 160–180 Myr, except for albitites from Chenaillet [24,25] and Inzecca [5]. The age proposed here (161 ± 3 Myr by U-Pb method) falls in this range. As the samples come from the Rospigliani Series, the dated magmatic event may be related to the activity of a possible transform fault zone. Similar values from the two albitites (a fine-grained cumulate and a more rapidly cooled granular rock mixed with basic lavas) indicate that formation of the magmatic chamber and outpouring of lava were coeval.

In slightly metamorphosed areas such as Balagne, Corsica, microfauna enclosed in sediments above the ophiolitic lavas given a Malm age (see extensive review in references 4, 47, 48, 49). As similar faunas are known in the Apennines it seems that the most important ophiolitic magmatism has ceased before the Malm. Hence from radiometric and sedimentation ages, the main oceanization event was of a short duration about 40 Myr, during Middle/Upper Jurassic time. The Liguro-Piedmont Basin, possibly a remnant of the Atlantic Ocean, was composed mainly of short ridge segments interrupted by numerous transforms [8,50,51]. Its opening was contemporaneous with that of the Central Atlantic [45,46], beginning with horst and graben systems in the continental margin since middle Lias [52]. Later, a tholeiitic volcanism of Cretaceous age occurred in the Valaisan-Versoyen trough. Sinistral transcurrent movements between African and European plates were going on until the Lower Cretaceous and favoured the formation of olistoliths and olistostromes [53]. Compressional movements linked to the closure of the ocean acted since Middle Cretaceous [54,55] and were followed by an important Eoalpine metamorphic (LT-HP) phase during the upper Cretaceous [56].

6. Conclusions

(1) The morphology of zircons from two albitites from Corsican ophiolites indicates rapid growth of these crystals in residual acid liquids participating in the building-up of the roof of the magma chamber.

(2) K-rich inclusions of granitic composition in the zircons testify to a potassium enrichment in the residual liquids, before its separation possibly in a high-temperature vapour phase as suggested by Sinton and Byerly [30].

(3) In the typological diagram of Pupin and Turco [15], the zircons plot in a restricted area and have high temperature and high agpaicity indexes.

(4) The alkali enrichment "peralkaline tendency" in the silicic liquids has permitted a late precipitation of zircon at the end of the fractional crystallization process. The high Zr concentrations in the residual magma have been favoured by adcumulus growth, the plagioclase effect and immiscible process.

* N.B. Zircon fraction II.3 gives a concordant U-Pb age at 143 Myr. Lack of material has prevented duplicate analyses which might show whether this age difference is analytical, as suspected, or real.

(5) U-Pb ages obtained for the first time from zircons in Tethyan ophiolitic albitites gave a 161 ± 3 Myr age corresponding to mid-Jurassic time.

(6) The Liguro-Piedmont Ocean created during the Jurassic was probably a portion of the Central Atlantic ocean.

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The Trace Element Geochemistry of Corsican Ophiolites*

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Abstract. Ultramafic rocks, gabbros, plagiogranites, dolerites and basaltic dikes and pillows lavas of the ophiolite slices of eastern Corsica have been metamorphosed in both oceanic and orogenic environments. The trace element content of the metabasaltic rocks which exhibit a tholeiitic differentiation trend towards a ferrobasaltic composition, is similar to that of oceanic basalts. The cumulate sequence is interpreted to have been formed by fractional crystallization from an olivine-tholeiite magma. The troctolites representing cumulates derived from melts representing liquid fractions in the range $F=1$–0.85, gabbros from melts with $F=0.85$–0.45, and ferrogabbros and plagiogranites from melts with $F<0.45$. An oceanic environment with moderate spreading rates and magmatic processes similar to present-day normal oceanic ridge segments is considered the more probable original setting of the Corsican ophiolites.

Geological Setting and Previous Work

In the Mediterranean area the ophiolites of Corsica form the most western fragment of the Tethyan ophiolitic belts. In north eastern Corsica, numerous ophiolite units are found as imbricated slices within the "Schistes lustrés" nappe (Durand Delga, 1974, 1975) to the east of the western shield (Maluski et al., 1973) (Fig. 1). As in the Apennines radiolarian cherts and limestones of Malm age cover the pillow lava sequence (Amaudric du Chaffaut et al., 1972).

Three main types of metamorphism have been recognized in the ophiolite sequence. The first metamorphism, compared to sub-sea floor metamorphism, produced hydrous minerals of the greenschist facies and involved migration of elements such as Ca, Mg and Na as well as high temperature metamorphism

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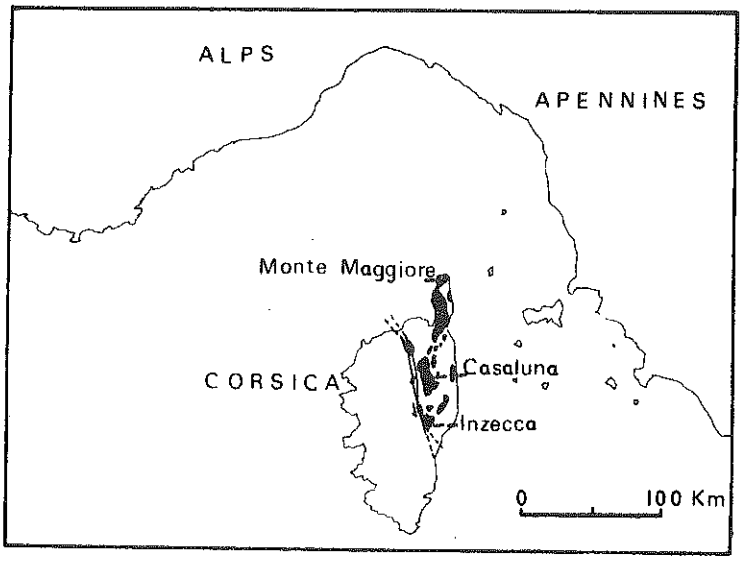


Fig. 1. Schematic map showing Corsican ophiolite outcrops (in black) in the Alpine-Apennine system

and deformation (amphibolite to granulite grade) of some gabbros and ultramafic rocks. The second metamorphic event of high pressure- low temperature type may be related to subduction processes possibly during an "eoalpine" (Upper Cretaceous) tectonism (Mattauer and Proust, 1976). Not every ophiolite fragment has been affected by all these metamorphic events which together have been only recognized in some outcrops (Ohnenstetter et al., 1976).

The ophiolite sequence reconstructed by Ohnenstetter and Ohnenstetter (1975) has been considered on the basis of petrographic and major element geochemical studies to be a remnant of Jurassic oceanic crust. From base to top these ophiolites are built up of 1) an ultramafic unit, 2) a mafic unit composed of troctolite, euphotide, ferrogabbro and acid residue, 3) hypabyssal rocks mainly massive dolerites and 4) pillow lavas with some dikes. The ultramafic rocks consist of poorly layered lherzolites, clinopyroxene-poor lherzolites and minor harzburgites and dunites with accessory aluminous phases (plagioclase and/or spinel). Alternating layers of dunites, often with squeletal olivine, anorthosites, and troctolites with accessory clinopyroxene and chromite form the base of the gabbroic sequence. The troctolites are overlain by poorly layered euphotide rocks, the most abundant type of gabbro, in which altered plagioclase and calcic augite are locally associated with transformed olivine. In some areas, euphotide rocks alternate with iron-rich gabbros (intermediate gabbros) in which branching clinopyroxene occurs. In the ferrogabbros cumulus magnetite, apatite and sulfides are well developed along with cumulus plagioclase, clinopyroxene and transformed accessory orthopyroxene and olivine. Albite-granites with albite as the main phase and lesser amounts of pyroxene and amphibole now converted to sodic pyroxene and sodic amphibole, quartz, zircon, magnetite, apatite and allanite, are locally abundant within the ferrogabbros and gabbros and the

pillow lava units. Adcumulates, heteradcumulates and crescumulates (Wager et al., 1960; Wager and Brown, 1968) are important components of the cumulate sequences. Mesocumulates and orthocumulates are also locally well developped in the ferrogabbros and acid rocks. However no evidence for the existence of continuous magmatic currents is found. Massive dolerites have been previously considered as the chilled margins of a magmatic chamber (Bebien et al., 1975). They are overlain by pillow lavas which although mainly variolitic and aphyric are locally porphyritic. The effusive and hypabyssal rocks are olivine tholeiites. However the more differentiated dolerites fall in the quartz tholeiite field. The massive dolerites and dike rocks show a clear tholeiitic evolution with strong Fe and Ti enrichment and Si and Al decrease. The major element variation can be directly related to fractional crystallization reflected not only in the crystallization order in the cumulates and dolerites and the nature of phenocrysts in basaltic rocks, but also in the physical nature of the cumulates formed by diffusion in a supersaturated medium. The early stage of fractionation is characterized by olivine and plagioclase precipitation whereas the more advanced stage is marked by the formation of clinopyroxene and magnetite.

In this paper new data on the trace element content of the basaltic rocks and cumulates of the Corsican ophiolites is provided in support of the hypothesis that they formed by crystal fractionation within a magma chamber below a mid-ocean rift.

Analytical Methods

The trace element contents of 32 samples out of 87 rocks already analysed for major elements were determined by L. Beccaluva and G. Venturelli using atomic absorption (Perkin-Elmer 303) and X-ray fluorescence (Siemens Kristalloflex IV) with the following precision:

Atomic absorption: Sr 3%, Li 2%, Ni 3%, Co 4%, Cr 4%, Cu 4%, Zn 4%, V 4%.

X-ray fluorescence: Rb 5–15%, Nb 18%, Y 3–8%, Zr 3%.

Al_2O_3, Fe_2O_3 as total iron, MgO, CaO, K_2O and TiO_2 were obtained by X-ray fluorescence (Siemens SRS) by R. Montanari. SiO_2 and Na_2O were determined by atomic absorption (Perkin-Elmer 306) by Y. Perrin. P_2O_5 determinations by colourimetric methods were carried out by L. Beccaluva and G. Venturelli. Trace element values obtained on international standard rocks during the analyses of Corsican ophiolites are reported on appendix II. The data are reported on Tables 1 and 2. Localities and mineralogical composition of 4 massive dolerites, 5 pillows, 4 dikes, 6 peridotites, 10 gabbros and 3 plagiogranites are reported in Appendix 1.

Chemistry and Original Setting of Basaltic Rocks

Although fresh basalts from different tectonic settings show distinctive chemical features for some major and trace elements (Pearce and Cann, 1973) such an approach is difficult to apply when the rocks are chemically altered. It is well known that weathering, spilitization and metamorphism induce strong chemical mobilization, particularly of the alkali and alkaline-earth elements. Ophiolites, presumed to have formed in an oceanic environment, are often strongly altered and have undergone secondary element mobilization (Montigny et al., 1973; Allegre et al., 1973; Mevel, 1975; Ohnenstetter and Ohnenstetter, 1975; Pearce, 1975). In Corsica, basaltic rocks and cumulates, especially those

Table 1. Chemical data of volcanic and plutonic ophiolite rocks from Corsica, western Mediterranean area (oxidas as wt %, elements as ppm)

Basaltic rocks													
Samples	Dykes				Dolerites				Pillow lavas				
	2093	596a	594	16911	2761	2763	7991	25611	2563	2295	2293	3084A	3087
SiO_2	49.03	52.82	49.86	54.54	48.35	48.00	50.00	48.31	52.51	50.95	49.16	49.41	51.47
TiO_2	1.13	1.93	2.41	2.55	1.13	1.32	2.33	3.03	1.61	1.70	1.37	1.69	1.61
Al_2O_3	16.06	13.83	14.11	12.87	16.32	16.26	14.73	13.56	18.14	15.55	17.28	13.41	15.08
Fe_2O_3tot	8.14	8.65	11.92	11.20	6.95	9.23	12.64	14.74	7.22	9.80	9.55	9.69	8.43
MnO	0.15	0.15	0.20	0.15	0.13	0.17	0.17	0.28	–	–	0.15	–	–
MgO	8.76	6.84	5.72	5.25	9.67	8.42	7.30	7.42	6.35	5.32	3.20	9.93	2.68
CaO	8.20	8.41	8.03	5.05	8.18	7.02	5.12	5.46	1.92	6.44	9.03	9.26	16.57
Na_2O	3.08	4.82	4.60	6.37	2.69	4.74	5.56	5.24	3.77	4.80	5.25	2.33	–
K_2O	1.25	0.03	0.02	0.15	1.22	0.05	0.01	0.20	4.27	1.26	0.42	0.05	<0.01
P_2O_5	0.10	0.20	0.20	0.36	0.11	0.13	0.24	0.30	0.13	0.14	0.17	0.19	0.19
L.I.[a]	3.83	2.30	2.76	1.77	4.46	3.93	2.83	2.47	3.89	3.23	4.21	4.14	4.34
Total	99.76	99.98	99.83	100.26	99.21	99.27	100.92	101.01	99.81	99.19	99.79	100.10	100.18
Fe O*/MgO	0.84	1.14	1.88	1.92	0.65	0.98	1.56	1.79	1.02	1.69	2.69	0.89	2.83
Ni	160	93	79	50	170	137	39	41	120	153	95	134	85
Co	39	41	51	37	31	56	38	53	36	48	44	44	50
Cr	380	152	88	n.d.	231	30	n.d.	34	379	350	350	301	186
Cu	22	7	41	52	53	76	13	36	60	84	48	59	35
Zn	49	27	83	27	51	68	60	103	58	69	92	81	90
V	183	268	313	292	182	193	277	395	184	235	202	213	252
Sr	152	123	48	44	215	64	41	34	12	121	63	23	20
Li	7	9	11	6	18	15	13	19	16	32	24	19	4
Rb	24	8	6	<6	18	<6	<6	<6	–	36	16	8	<6
Nb[b]	–	–	8	6	7	–	–	8	7	–	–	–	–
Zr	88	161	180	260	96	104	203	188	132	129	120	166	121
Y	36	62	68	85	40	36	67	50	40	54	49	48	37
88/Zr	1.00	0.55	0.49	0.34	0.92	0.85	0.43	0.47	0.67	0.68	0.73	0.53	0.73
Zr/Nb	–	–	22	43	14	–	–	–	24	19	–	–	–

of the upper units, have been affected by weathering and several metamorphisms and therefore tend to be spilitic. Some meta-basaltic pillows also show a K_2O enrichment, probably due to sea-water interaction. Such alteration is not found in the massive dolerites.

In this paper, Ti, P, Cr, Y, Nb, Zr are selected to characterize the magmas (Pearce and Cann, 1971, 1973; Pearce, 1975; Floyd and Winchester, 1975; Winchester and Floyd, 1976). A clear tholeiitic affinity appears in the P_2O_5-Zr and TiO_2-Zr/P_2O_5 diagrams of Figure 2 as well as by the high Y/Nb ratios. On the Ti-Zr-Y discrimination diagram (Fig. 3) all the samples fall in field B. The Ti-Zr diagram shows that most of the samples group in or near the ocean-floor basalt field (OFB) but that some iron-rich dolerites and dikes extend beyond the limits of this field at high Ti and Zr values (Fig. 4). The Ti-Cr

Table 1 (continued)

Samples	Troctolites				Gabbros			Ferrogabbros			Plagiogranites		
	11914	11912	11911	119151	12924	2096	1711	2383	2095	297131	2879	30731	23861A
SiO_2	45.62	39.58	47.65	49.50	48.81	48.61	49.60	40.31	46.34	54.00	79.45	68.91	60.48
TiO_2	0.07	0.05	0.09	0.20	0.30	0.92	0.27	4.78	5.01	1.44	0.15	0.19	0.79
Al_2O_3	19.40	9.65	23.35	18.88	17.00	15.47	14.50	12.90	11.27	14.15	12.08	16.52	16.99
Fe_2O_3tot	4.78	7.64	2.76	3.70	4.02	8.31	4.00	16.25	18.06	15.78	1.35	3.01	6.25
MnO	0.08	0.10	0.08	0.09	0.10	0.10	0.08	0.25	0.15	n.d.	n.d.	n.d.	0.06
MgO	13.40	26.14	5.53	10.35	9.29	8.02	12.25	10.27	5.51	2.76	0.10	2.17	1.49
CaO	9.86	5.12	12.00	12.68	13.55	10.87	13.42	3.96	8.86	2.22	0.01	0.03	1.84
Na_2O	2.18	0.40	3.04	2.60	2.57	3.83	3.24	5.34	3.53	8.02	6.59	7.84	11.28
K_2O	<0.01	<0.01	0.01	<0.01	<0.01	0.01	<0.01	<0.01	0.10	0.09	<0.01	<0.01	0.05
P_2O_5	<0.02	0.02	<0.02	0.03	<0.02	<0.02	0.02	0.04	0.13	0.40	0.05	0.02	0.15
L.I.[a]	4.40	11.17	3.19	1.29	4.39	3.42	3.20	4.64	1.57	1.36	0.54	1.28	0.65
Total	99.79	99.91	99.78	99.32	100.13	99.66	100.58	98.70	100.55	99.82	100.26	99.97	100.03
Fe O*/MgO	0.32	0.26	0.33	0.32	0.39	0.93	0.29	1.42	2.95	5.14	12.10	1.25	3.78
Ni	474	1130	254	283	153	167	217	76	64	110	–	63	22
Co	50	80	30	43	52	83	34	59	59	25	8	12	10
Cr	110	1100	490	1080	484	378	1260	45	–	–	–	–	–
Cu	48	119	35	63	211	73	17	54	57	18	22	71	10
Zn	29	36	20	30	26	51	20	130	97	208	17	25	65
V	26	28	33	77	100	174	135	507	558	49	–	37	–
Sr	147	70	176	109	71	236	108	19	99	37	–	10	27
Li	2	1	2	2	2	3	19	47	4	15	9	13	11
Rb	<6	<6	<6	–	<6	<6	<6	<6	<6	<6	<6	–	<6
Nb	–	–	<2	2	3	4	6	7	10	11	25	26	16
Zr	14	–	20	20	25	32	46	85	113	620	305	345	2200[c]
Y	<3	–	<3	3	10	12	24	49	41	170[c]	180[c]	370[c]	124
88/Zr	6.28		4.4	4.4	3.52	2.75	1.91	1.03	0.78	0.14	0.29	0.26	0.04

[a] loss on ignition
[b] only some relatively high values are reported
[c] approximate values
n.d. = not detected, – = not determined

diagram, which is useful in discriminating between OFB and island arc tholeiites (LKT), confirms the ocean-floor origin of the Corsican ophiolites. It is interesting to note, however, that some dolerite samples have an LKT affinity. This may be due to fractionation of small amounts of chromite bringing about a drop of Cr content for little change in the titanium content of the magma during the early differentiation stages (see also later). Use of the Ti-Cr diagram to discriminate OFB and LKT may lead therefore to an erroneous interpretation of the tectonic environment in which the basalts formed. Figure 4 also shows the field of ophiolite basalts from the Northern Apennines (Beccaluva et al., 1975, 1976; Ferrara et al., 1976). On all these diagrams, except for Ti-Zr-Y,

Table 2. Ultramafic rocks (oxides as wt %, elements as ppm)

	2591	2594	1398c	1398b	227152	29619
SiO_2	40.38	46.36	37.81	42.32	39.59	38.84
TiO_2	0.06	0.15	0.04	0.09	0.02	0.11
Al_2O_3	2.46	3.70	1.74	1.27	0.62	3.13
Fe_2O_3tot	7.55	6.16	9.92	8.55	9.94	8.54
MnO	0.12	0.10	0.10	0.10	0.09	0.10
MgO	34.53	26.00	43.10	42.85	35.80	32.20
CaO	2.70	9.40	0.47	0.98	0.01	4.86
Na_2O	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01
K_2O	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01
P_2O_5	<0.02	0.02	0.02	<0.02	0.02	0.02
L.I.[a]	11.49	7.85	5.83	2.01	13.20	11.78
total	99.29	99.72	99.01	98.17	99.26	99.56
FeO*/MgO	0.20	0.21	0.21	0.18	0.25	0.24
Ni	2040	1080	2400	2260	2450	2150
Co	100	62	129	123	118	90
Cr	2870	4100	5980	2090	3270	2960
Cu	15	17	19	29	11	29
Zn	–	39	67	60	50	58
V	72	95	44	61	n.d.	93
Sr	n.d.	7	n.d.	n.d.	n.d.	n.d.
Li	n.d.	4	2	n.d.	n.d.	3

[a] loss on ignition. n.d. = not detected, – = not determined

fractionation trends in dolerites and dikes are clearly demonstrated. Such processes should be considered (cf. Shibata and Fox, 1975) when attempting to correlate TiO_2 of basic volcanics to the spreading rate of the oceanic diverging plates. However, disregarding this effect, on the basis of Nisbet and Pearce's (1973) correlation Ti and the spreading rate, a moderate half-spreading rate about 2 cm/year is suggested for the accretion of the Corsican ophiolites (Ohnenstetter and Ohnenstetter, 1975). Similar values have been proposed for the Austrian Alps (Bickle and Pearce, 1975) and for the Northern Apennines (Beccaluva et al., 1975; Ferrara et al., 1976).

Zr/Nb ratios between 14 and 105 for the basalts are comparable to those of normal Mid Oceanic Ridge basalts (Table 3). The difference in Nb values, independant of shallow depth fractionation (Erlank and Kable, 1976) between the Apennine and Corsican basalts (Beccaluva et al., 1976) may reflect different degrees of depletion of the mantle source. However in terms of the chemical characterization of oceanic ridge basalts of Hart et al. (1973), Schilling (1973, 1975), and Bryan et al. (1976) both the Corsican and Apennine basalts seem to belong to "a normal ridge segment". The very flat REE pattern (Allegre et al., 1976; Noiret and Allegre, 1976) of some Inzecca samples, Corsica, agrees with this assumption.

Pronounced chemical differences also occur between western and eastern Mediterranean ophiolites e.g. Corsica and Apennines, versus Troodos, respec-

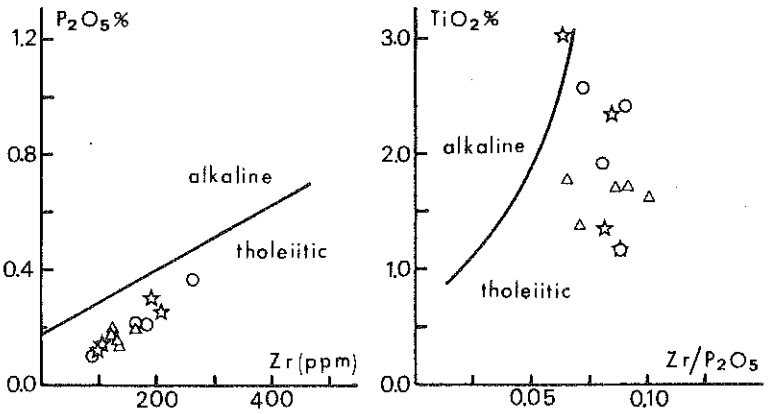


Fig. 2. P_2O_5%-Zr and TiO_2%-Zr/P_2O_5 diagrams (Floyd and Winchester, 1975; Winchester and Floyd, 1976) illustrating the difference between tholeitic and alkaline basic rocks. Corsican ophiolite basaltic rocks: *1* Dolerites; *2* Dikes; *3* Pillow lavas

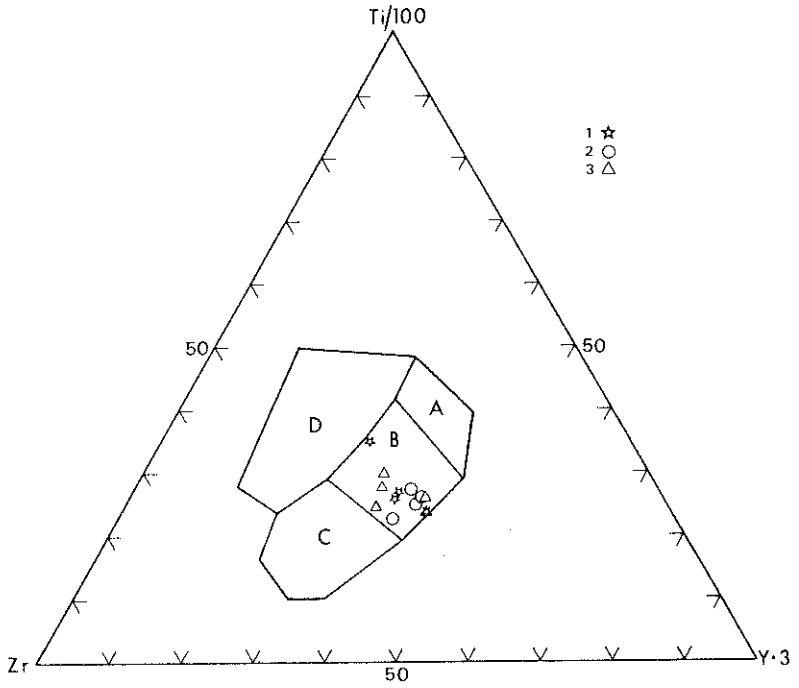
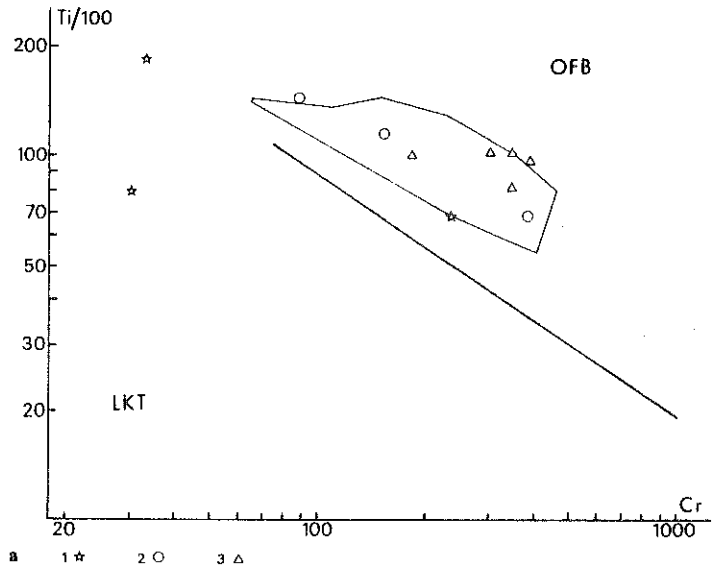


Fig. 3. Ti-Zr-Y discrimination diagram for basic volcanic rocks (Pearce and Cann, 1973). "Within plate" basalts plot in field *D*, ocean floor basalts in field *B*, low potassium tholeiites in fields *A*, and *B*, calc-alkali basalts in fields *C* and *B* Corsican ophiolite basaltic rocks: *1* Dolerites; *2* Dikes; *3* Pillow lavas



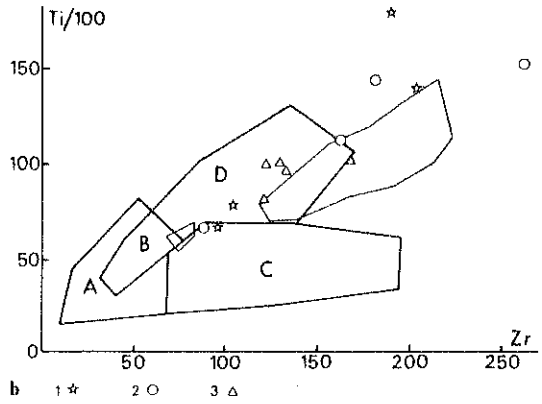


Fig. 4a and b. Ti-Zr discrimination diagram (Pearce and Cann, 1973). Ocean-floor basalts plot in fields *D* and *B*, low-potassium tholeiites in fields *A* and calc-alkali basalts in fields *C* and *B*. Ti-Cr discrimination diagram (Pearce, 1975). *OFB* ocean-floor basalts; *LKT* island-arc tholeiites. Corsican ophiolite basaltic rocks: *1* Dolerites; *2* Dikes; *3* Pillow lavas. Distribution fields (light line) of basalts from Apennine ophiolites are also reported (data after Beccaluva et al., 1975, 1976; Ferrara et al., 1976).

Table 3. Average chemical composition of some Mediterranean ophiolitic lavas, Mid-Atlantic ridge basalts, and island arc volcanics

	1	2	3	4	5	6
SiO_2	50.06	48.74	54.18	53.22	49.21	51.42
TiO_2	1.44	1.60	0.58	0.69	1.39	0.81
Al_2O_3	15.14	15.97	14.84	14.47	15.81	17.97
FeO Tot	8.25	8.69	8.48	7.94	9.11	9.88
MnO	0.16	0.17	0.14	0.12	0.16	–
MgO	7.12	7.21	5.59	5.88	8.53	4.89
CaO	8.16	8.17	5.34	7.69	11.14	8.90
Na_2O	4.02	4.48	4.11	2.62	2.71	2.94
K_2O	0.50	0.39	0.61	0.76	0.26	0.26
P_2O_5	0.19	0.24	–	–	–	0.09
Ni	104	110	38	47	123	23
Co	44	43	50	54	–	28
Cr	216	256	43	106	296	43
Cu	45	68	76	72	87	94
Zn	66	92	–	–	–	–
V	245	231	288	286	289	297
Sr	74	218	84	118	123	131
Li	15	16	–	–	8	–
Rb	9	8	–	8	–	2.3
Nb	7	5	–	1	–	–
Zr	150	155	–	66	100	37
Y	52	47	–	26	43	23

1 = Corsican basaltic rocks (Ohnenstetter and Ohnenstetter, 1975; and this work.)
2 = Northern Apennine basalts (Beccaluva et al., 1975, 1976; and Ferrara et al., 1976.)
3 = Pillow lavas, flows and dykes from Troodos Basal Group (Desmet, 1977.)
4 = Lower pillow lavas and sheeted dikes and sills of the Troodos (Desmet, 1977; Y, Nb, Zr and Rb data after Pearce, 1975 and Beccaluva and Venturelli, unpublished data.)
5 = Average Mid-Atlantic ridge basalt (Thompson, 1973.)
6 = Basalts and andesites from Eua, Tonga Islands (Ewart and Bryan, 1972.)

tively (Table 3). Metabasalts (Q tholeiites) from the lower pillow lava unit and sheeted complex of Troodos – the so-called axis sequence (Smewing et al., 1975) – show lower Ti, Zr, Nb, Y, Cr and Ni contents (Pearce, 1975; Kay and Senechal, 1976; Desmet, 1977) than Corsican and Apennine basaltic rocks. Stronger differences, except for Cr, exist between the Troodos upper pillow lavas and the basaltic rocks of Corsica and the Apennines. Thus, important geochemical differences clearly exist between western and eastern Mediterranean ophiolites, the eastern ophiolites appearing to be more calk-alkaline (Miyashiro, 1973; Thayer, 1973; Bebien et al., 1975; Rocci et al., 1975; Paupy, 1976). The differences in geochemical characters are too marked to assign the same origin to the two groups of ophiolites. On the basis of geochemical and petrographic data, it appears that the Troodos ophiolite has a closer affinity with the low-K tholeiites of immature island arcs e.g. Eua (Ewart and Bryan, 1972) than with typical ocean floor basalts (Table 3). The nature of these "calk-alkaline" ophiolites raises the problem of their genesis (Ohnenstetter and Ohnenstetter, 1975;

Rocci et al., 1975; Cann, 1976). Similarly in the Appalachian province such a diversity exists between the ophiolites of the Internal and External domains (Church, 1977; Church and Riccio, 1977).

Evolutionary Paths of Magmatic Liquids and Cumulate Products

1. Choice of Parameters and Variation Diagrams

The chemical variations of basaltic rocks in Corsica ophiolites are considered to be the result of fractional crystallization, on the basis of petrographic considerations and chemical differentiation indexes. Parameters, such FeO total/MgO, although slightly sensible to alteration are nevertheless often inadequate in evaluation of the amount of residual liquid. Thus in the following estimation immobile elements such P, Zr, Ti, and Y which strongly increase during the differentiation are used (Fig. 5). The P_2O_5-Zr pair shows the highest increase and the most significant correlation coefficient (0.93) with a regression line passing near the origin. This trend indicate that Zr and P have a residual character (solid/liquid mean partition coefficient $\bar{D}^{Zr}$ and $\bar{D}^{P} \ll 1$) during fractional crystallization (Treuil, 1973; Treuil and Joron, 1975).

An evaluation of the liquid fraction F at a defined differentiation stage is given by the ratio C^A/C_1^A: initial concentration in the original magma versus concentration in the liquid of the residual element A (Rayleigh model). In determining a value for different basalt samples (dolerites and dikes), the concentration of Zr is used because of its better analytical precision besides a residual character. The value obtained is only approximate, not only because of the limits of the theoritical assumptions, the possible occurence of trapped liquid in the comagmatic cumulates, the difficulty of estimating the initial content of Zr in the primary liquid, and analytical error, but also because of other factors such as metamorphic mobilization of the element, the presence of Zr-bearing cumulus phases, liquid inhomogeneity, etc...

The initial Zr content of the primary basaltic magma was estimated on the basis of the distribution of Ni and Cr which are strongly incorporated in the early crystallizing minerals. In Ni+Cr versus Zr diagram (Fig. 6), dolerites and dikes which define two different trends, show a strong variation of Cr+Ni at low Zr values. At this early differentiation stages, the lowest Zr values (88 ppm) approach the least differentiated composition and is chosen to calculate F. This value which in any case must be considered an excess estimation, is similar to those of relatively undifferentiated oceanic basalts from different areas (Cann, 1970). The least differentiated dolerites (sample 2761) and dikes (sample 2093) show the following compositional ranges: MgO 8.8–9.7, Fe_2O_3 tot. 8.1–7.0, Al_2O_3 16.1–16.3, TiO_2 1.1, P_2O_5 0.10–0.11%, Y 36–40, Zr 88–96, V 180, Ni 160–170, Cr 380–231, Co 39–31 ppm, FeO/MgO 0.84–0.65.

2. Mineralogical Control on Differentiation

In Figure 7, variations of some major and minor elements which give good indications of differentiation processes are plotted varsus 88/Zr ratio. Alkali

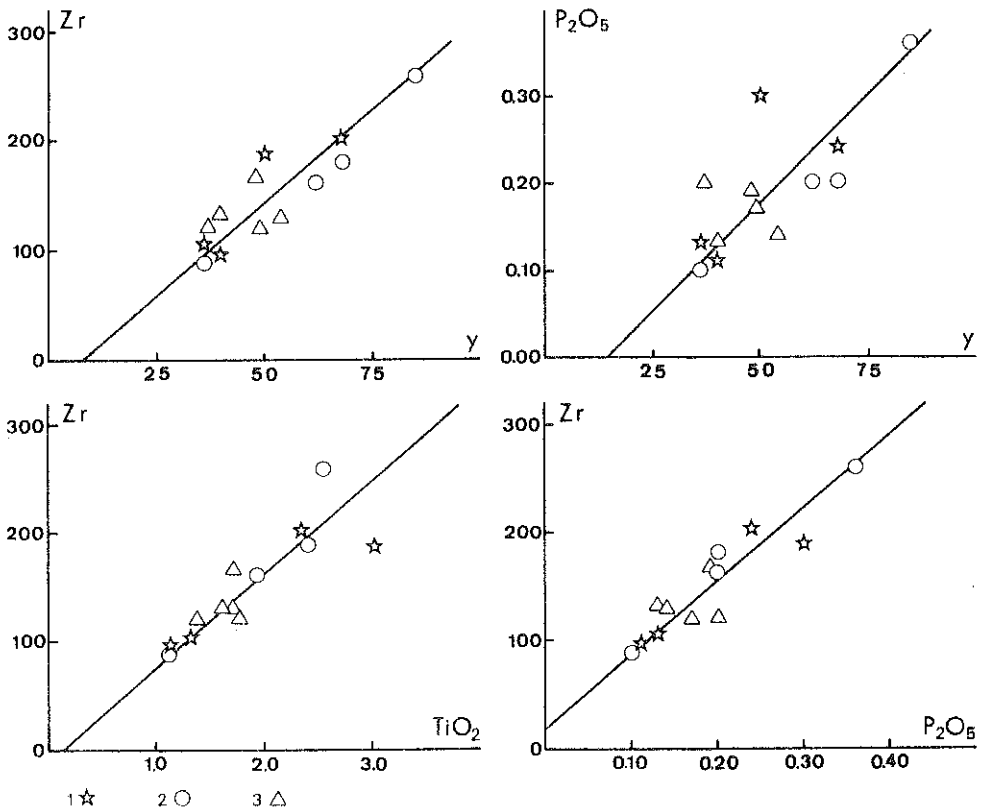


Fig. 5. Binary diagrams of 'immobile' element pairs, illustrating the strong increase exhibited by Corsican basaltic rocks during differentiation: *1* Dolerites; *2* Dikes; *3* Pillow lavas

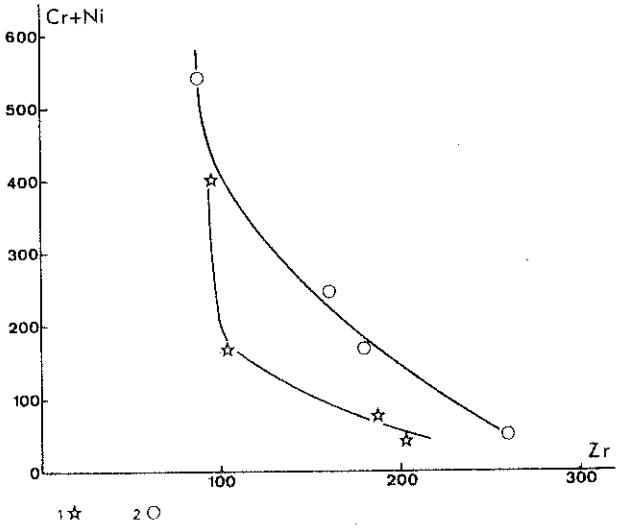


Fig. 6. Zr – Cr + Ni diagram. Evolutionary trends of dolerite. (*1*) and dike (*2*) basaltic rocks from Corsican ophiolites

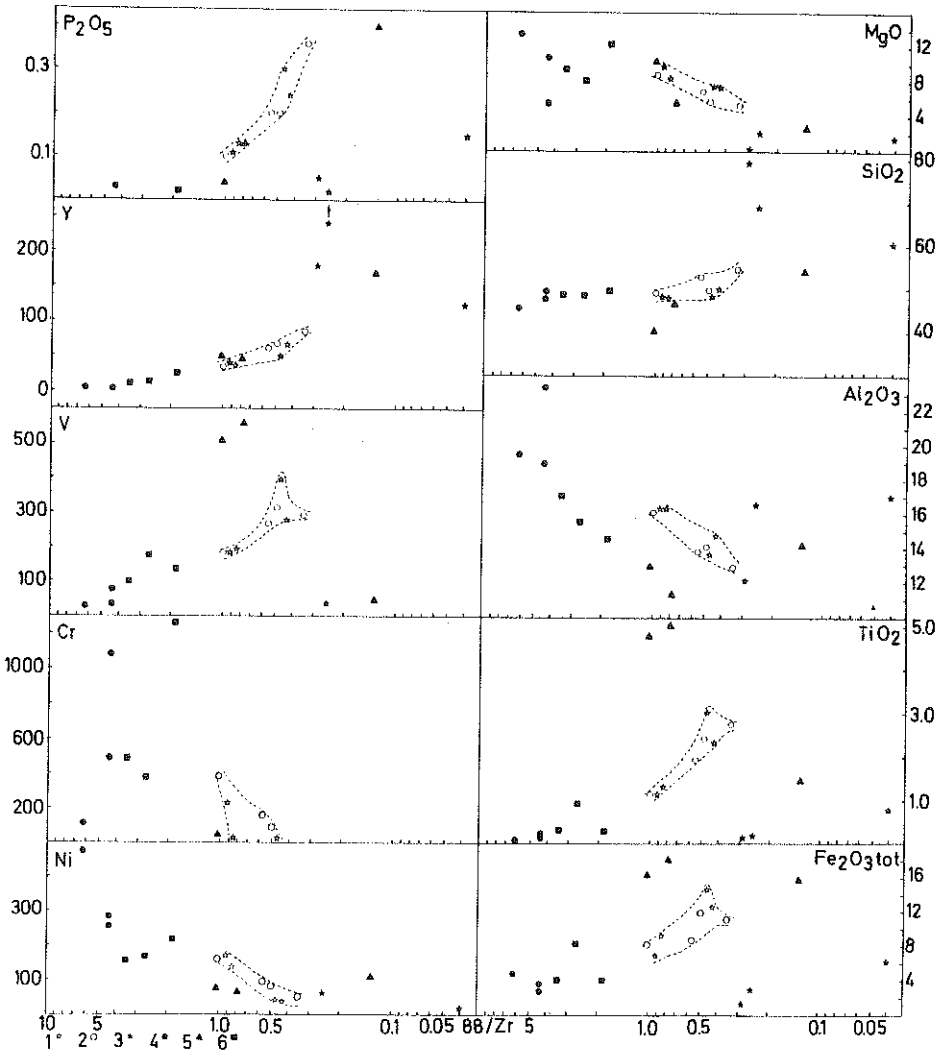


Fig. 7. Variations of some major and trace elements vs. 88/Zr ratio for Corsican ophiolites: *1* Dolerites; *2* Dikes; *3* Plagiogranites; *4* Troctolites; *5* Ferrogabbros; *6* Gabbros. Evolutionary trends of basaltic liquids (dolerites and dikes) are limited by dashed lines

and alkaline-earths which show wide dispersion are not reported. Sr, despite alteration, however seems to decrease in a regular way during the differentiation. Besides the dolerites and dikes, the basic cumulates, composed of troctolites, euphotides, ferrogabbros and plagiogranites are plotted because these rocks are thought to be partly responsible for the variation in composition of the basaltic rocks as may also be the case of Morb basalts (Melson and Thompson, 1970; Miyashiro et al., 1970; Bonatti et al., 1971; Shido et al., 1971; Thompson, 1973). The lavas are excluded because of their high degree of alteration.

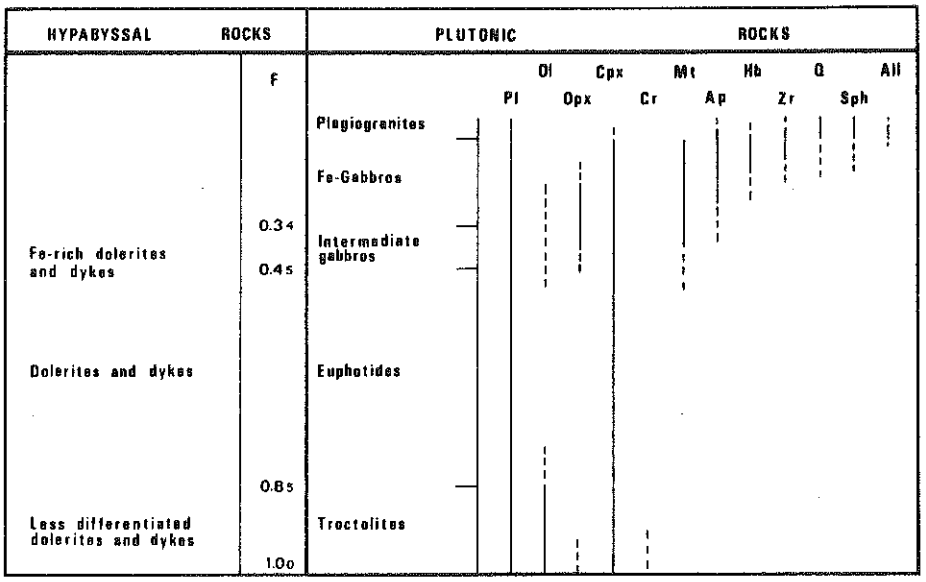


Fig. 8. Generalized section of Corsican plutonic mafic units showing the appearance and the disappearance of the main magmatic minerals: *Pl* plagioclase; *Ol* olivine; *Opx* orthopyroxene; *Cpx* clinopyroxene; *Cr* chromite; *Mt* magnetite; *Ap* apatite; *Hb* hornblende; *Zr* zircon; *Q* quartz; *Sph* titanite; *All* allanite. The corresponding fractionation stages of the basaltic liquid fractions (*F*) are shown on the left side of the diagram along with the relative rock types

The cumulates whose succession and mineralogy are reported in Figure 8 do not correspond to magmas represented by the basaltic effusive since they result from an average of cumulus phases with variable amount of pore material (trapped liquid). The 88/Zr may only serve to evaluate the proportion of this trapped liquid. Its value approaches infinity for rocks made entirely of cumulus minerals. If the magmatic relations between differentiation trends of the basaltic rocks and the cumulates are considered, it is apparent that the distribution of major and minor elements in basaltic rocks define two main discontinuities separating three stages of differentiation.

The early stages of basaltic liquid differentiation: $F=1$–0.85, if referred to dolerites, are characterized by a rapid fall of Ni and Cr which enter the early crystallizing mafic minerals, in particular the Cr content falls abruptly in dolerites. At this stage Al_2O_3 and MgO also decrease whereas SiO_2 is almost constant and all the other elements increase. The variation can be interpreted to result from the removal of plagioclase, olivine and Cr-spinel (troctolites) all of which occur as phenocrysts in the least differentiated rocks. Assuming that the troctolites formed from a magma with Zr content near 88 ppm, the high 88/Zr ratio together with the low content of other residual elements such Y, P and Ti of these cumulates suggest low amounts of trapped liquids (not more than 15–25%) in agreement with the adcumulate, mesocumulate and crescumulate growth structures shown by these rocks.

During the middle stage of differentiation: $F=0.85$–0.45 a typical tholeiite trend towards a ferrobasaltic composition is shown by the regular increase in Ti, Fe, V, Zn, Y, P, FeO total/MgO and Zr in the liquids and the decrease in Al, Mg, Ni and Cr. Removal of plagioclase, clinopyroxene and minor olivine, leaving gabbroic cumulates: olivine gabbros and euphotides, may yield such iron rich liquids. Similarly in normal ridge basalts, removal of plagioclase and olivine (Kay et al., 1970; Cann, 1971; Thompson et al., 1972; Bence et al., 1975) as well as clinopyroxene (in spite of the fact that phenocrysts of this mineral are scarce or completely lacking in oceanic tholeiites except in ferrobasalts (Mazzulo and Bence, 1976) has been used to explain such a trend (Miyashiro et al., 1970; Shido et al., 1971; Ayuso et al., 1976; Bryan et al., 1976; Byerly et al., 1976; Clague and Bunch, 1976; Schilling et al., 1976; Shibata, 1976). In Corsica, petrographic evidence show that clinopyroxene is always present as an accessory heteradcumulus phase in troctolite, olivine gabbro, and even more importantly in euphotide. It occurs only as liquidus phase in the late stages of differentiation when it is present as a phenocryst phase in ferrobasaltic rocks. Thus element diffusion between intercumulus liquid and the overlying magma (Hodges and Papike, 1976) may have operated. In this manner the presence of clinopyroxene in gabbros and its absence in the corresponding basaltic rocks during the early and middle stages may be explained. However, the decrease of the 88/Zr ratio in the euphotide gabbros does not necessarily mean a contemporaneous increase of pore materiel since the cumulate textures testify that elimination, to a greater or lesser extent, of the original intercumulus liquid has taken place, but may be only, or also, due to the increase of the zirconium content in the trapped liquid during the differentiation.

At advanced stages of differentiation: $F<0.45$ a decrease of Ti, Fe, V and Zn in the basaltic liquids is apparent in both dolerite and dike rocks. The decrease in concentration of these elements may be related to the removal of titaniferous magnetite and possibly minor sulfides as cumulus phases, leading to ferrogabbroic cumulates which are obviously enriched in the same elements. Orthopyroxene precipitation in such cumulates may indicate silica saturation (O'Hara, 1968). Finally, it must be noted that dolerite and dike samples showing a decrease in P and Zr have not been found. It may therefore be concluded that apatite and zircon, which appear in some ferrogabbros, are removed in appreciable amounts only from the most differentiated magmas: at least $F<0.35$ and that, apatite- and zircon-bearing ferrogabbros and the more evolved cumulates are formed from liquids not represented amongst the dolerite, dike and pillow lava units of the sequence. Sample 297131 low in V and Ti and high in Si, Na, Y and P may correspond to an intermediate fractionation stage between ferrogabbroic composition and more acid residues.

The fractionation products of the latest magmatic stages are represented by albitegranites composed of Na-rich plagioclase, clinopyroxene and/or amphibole, Fe-Ti oxides, zircon, apatite, titanite, allanite and Quartz (Table 1). Petrographic observations suggest that the albitegranites represent cumulus solid fractions with variable amounts of trapped liquid (ortho- to mesocumulates). The very low content of the residual elements K and Rb may be related to the lacking of K-feldspar and biotite as well as to secondary mobilization.

The Sr content in these albitegranites are lower than those reported by Coleman and Peterman (1975). In Corsica potassic-rich rocks representing an ultimate residual liquid have yet to be found. In oceanic environment such rocks are very scarce having been reported only from the Indian ocean in the form of a quartz-monzonitic dikelet with K-feldspar and biotite (Engel and Fisher, 1975).

Conclusion

The repartition of trace elements in basaltic rocks of Corsican ophiolites confirms their ocean-floor character. Their tholeiitic trend of differentiation is marked by a strong increase in Fe, Ti, V, Zr, Y, P, Zn and decrease in Mg, Al, Ni, Cr. The mineral phases which control the fractional crystallization are the same as those found in the underlying gabbroic sequence. The troctolitic rocks crystallized from magma with an F value in the range 1–0.85, euphotide from magma with an F value of 0.85–0.45, and intermediate gabbros, ferrogabbros and albitegranites from magma with an F value of less than 0.45. The petrographic and geochemical features of the ophiolites indicate that accretion occured during Jurassic time (Ohnenstetter et al., 1977) in the western Mediterranean area according to magmatic processes similar to those thought to be active in present-day normal mid-oceanic ridge segments. The abundance of gabbroic adcumulates in Corsica (Ohnenstetter et al., 1975) and in the Atlantic Ocean (Hodges and Papike, 1976) coupled with their low content of residual elements is compatible with a slow cooling of successive magma batches under a high geothermal gradient. Beneath mid-oceanic ridges, olivine-tholeiite produced or reequilibrated at pressures of 5–10 kb (Green and Ringwood, 1967; O'Hara, 1968) evolve towards liquids of ferrobasaltic composition.

In Corsica, the presence of ultramafics with cumulus textures raises the problem of the existence and significance of 'peridotite' fractionation. Such lherzolites, equilibrated at pressures of 7–9 kb (spinel-plagioclase transition) may have separated from a more primitive olivine-rich picritic liquid (O'Hara, 1968) derived from a deeper zone e.g. the low velocity zone. According to this assumption, the abyssal tholeiites reach the surface after polybaric fractionation, leaving peridotite with a lherzolitic character located between gabbros constituing layer 3 and the upper part of the underlying mantle of the oceanic lithosphere.

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Appendix 1: List of Samples

Pillow Lavas

They are issue from the Inzecca massif. Primary minerals are transformed into a greenschist assemblage. But high pressure minerals may be locally present.

30.8.4A. Col de Grata. Variolitic pillow with large varioles (1 cm), vacuoles (0.1 mm). ab, q, chl, epi, pum, act.

25.6.3. Défilé de l'Inzecca. Brecciated variolitic pillow with transformed olivine microphenocrysts (lantern like form) in a recrystallized fine groundmass. ab, q, chl, ser, pum, hem.

22.9.5. Fatachio. aphanitic pillow. aug, chl, ab, q, epi, lw, pum, tit, ser.

22.9.3. Fatachio. Porphyritic pillow with plagioclase phenocrysts. ab, q, epi, lw, pum, chl, tit.

30.8.7. Col du Cardo. Hematitic pillow with plagioclase phenocrysts and squelettic transformed olivine microphenocrysts and hollow plagioclase microlites in a recrystallized glass.

Dykes

They cut the upper gabbros (euphotide, intermediate gabbro). No magmatic mineral has been preserved. Assemblages are characteristic of a greenschist facies with occurrence of pumpellyite; sodic amphiboles are locally present.

20.9.3. Monte Piano Maggiore. Intersertal texture. act, chl, ab, q, tit.

5.9.6.A Monte Piano Maggiore. Sub-ophitic texture with plagioclase phenocrysts. act, chl, tit, ab, q, pum.

5.9.4. Inzecca massif. Sub-ophitic to intersertal texture with plagioclase phenocrysts. act, ab, epi, pum, q, hem, tit, chl, ser.

16.9.11. Monte Piano Maggiore. Sub-ophitic to intersertal texture with plagioclase and magnetite phenocrysts, numerous squelettic Fe-Ti oxides in the groundmass. act, ab, tit, epi, pum, q, ser.

Dolerites

Inzecca massif. They intrude the pillow lava carapace. The following magmatic mineral proportion is only a broad estimation due to alteration products.

27.6.1. Ophitic texture. Mode is: plagioclase (2/0.7 mm) 67%, clinopyroxene (10/.6 mm) 20%, olivine (2 mm) 11%, ilmenite 2%. Clinopyroxene is the only relictual magmatic phase. Ab, cpx, chl, pum, tr, il, tit, ser, q, cros.

27.6.3. Ophitic to nesophitic textures. Mode is: pl. (2.5/0.5 mm) 76%, Cpx. (6/4 mm) 20%, il. 3%. Cpx., ab., Chl., pum., il., tit., q.

7.9.9.1. Intersertal to sub-ophitic textures; locally graphic intergrowth of pl. and cpx. Mode is: pl. (4/1 mm) 74%, cpx. (3/0.7 mm) 21%, mt. 5. %, ap. 0.1%. Aug., ab., q., mt., leuc., act.

25.6.11.2. Intersertal texture. Close association of cpx. and mt. Clinopyroxene crystallized before pl., is sometimes bent. Mode is: pl. (5/1 mm) 62%, cpx. (3/0.5 mm) 27%, mt. 11%, ap. 0.1%. Fe and Ti rich aug., ab., mt., leuc., Fe-chl., aeg., cros., ser., q.

Ultramafics

25.9.1. Cap Corse. Ol., opx., and cpx mesocumulat (lherzolite). Diopside is fresh. opx and ol. are serpentinized.

25.9.4. Cap Corse. Ol. websterite adcumulate serpentinized.

13.9.8.c. Casaluna. Deformed dunite. Protogranular texture. Ol. (Fo_{88}) brown aluminous Cr-spinel, transformed pl. into hydrogrossular.

13.9.8.b. Casaluna. Deformed harzburgite. Transformed pl. and Cr-spinel are accessory; ol. (Fo_{90}); enstatite ($2V_Z = 76° \pm 1$). Protogranular texture; olivine shows elongate tablets bigger than orthopyroxene ones, and parallel to the plagioclase foliation.
22.7.15.2. Inzecca serpentinite. Liz, mt.
29.6.19. Inzecca. serpentinite. Liz., mt., hydrogrossular.

Troctolites

Samples come from the Casaluna massif. They are composed of plagioclase (An_{80-70}), olivine (Fo_{90-85}), and diopsidic augite ($2V_Z = 53° \pm 2$; $ZZcc = 40° \pm 2$).
11.9.15.1. Pl.-ol.-Cpx. mesocumulate (olivine gabbro). Pl. size varies from 0.2 to 1.2 cm., cpx. is poikilitic.
11.9.11. Pl.-ol.-cpx. heteradcumulate (troctolite). Pl. (1.5 cm), very abundant are locally separated from olivine by a narrow zone of chl. (0.1 mm). Secondary minerals: liz., mt., pre.
11.9.12. Ol.-pl. heteradcumulate (plagioclase dunite). Sub-automorph ol. and few chromites are inclosed in heteradcumulus pl. Other phases: liz., mt., pre.
11.9.14. Pl.-ol mesocumulate (troctolite). Locally accessory heteradcumulus cpx. Secondary minerals: liz., mt., pre., tr.

Gabbros

20.9.6. Casaluna. Pl.-cpx.-Mt. adcumulate. Aug. is fresh. act., chl., ab., q.
12.9.24. Casaluna. Pl.-cpx.-ol. heteradcumulate (olivine gabbro). Cpx. is fresh (diallage). Chl. pseudomorph ol. which are separated from pl. by corona texture. Tr., Mg-chl., ab., pre.
1.7.11. Inzecca. Pl.-cpx. heteradcumulate. Diallage is deformed (kink-bands), Tr., chl., pre.

Ferrogabbros

20.9.5. Casaluna. Pl.-cpx.-mt.-opx.-ap. mesocumulate. Cpx. and ap. are the two preserved magmatic minerals. Other phases: brown hb., act., Fe-act., chl., cros.
29.7.13.1. Inzecca metagabbro. All the magmatic minerals are transformed except ap. and zircon. (0.2–1 mm). Aeg., aegyrinic pyroxenes, cros. stil., ser., ab.
23.8.3. Inzecca massif. Metagabbro. Aegyrinic pyroxenes, aeg., cros., stil., Fe-chl., ab., leuc., tit.

Plagiogranites

23.8.6.1.A. Inzecca massif. Pl.-cpx.-Zr.-ap. mesocumlate. Cumulus structure is well preserved despite metamorphic recrystallization. Igneous lamination is marked by pl. Ab., jadeitic and aegyrinic pyroxenes, Fe-chl., zr., ap.
30.7.3.1. Inzecca. Plagiogranite. No magmatic texture is preserved; mortar texture ab., q., Fe-chl., zr., mt.
28.7.9. Inzecca. Plagiogranite. Mortar texture. Ab., q., he., stil., ZR.

Abbreviation of the minerals: *ab* = albite; *act* = actinolite; *aeg* = aegyrine; *aug* = augite; *chl* = chlorite; *cpx* = clinopyroxene; *cros* = crossite; *epi* = epidote; *hb* = hornblende; *hem* = hematite; *il* = ilmenite; *leuc* = leucoxene; *liz* = lizardite; *lw* = lawsonite; *mt* = magnetite; *ol* = olivine; *opx* = orthopyroxene; *pl* = plagioclase; *pre* = prehnite; *pum* = pumpellyite; *q* = quartz; *ser* = sericite; *stil* = stilpnomelane; *tit* = titanite; *tr* = tremolite; *zr* = zircon.

Appendix II. Trace element concentrations obtained and used for interlaboratory standards

	GSP-1	AGV-1	BCR-1	PCC-1	DTS-1	BR	BM	T-1	SY-1	MRG-1	SY-2	SY-3
Ni	8	20	12	2375	2350	–	64	–	34	203	–	–
	(12.5)	(18.5)	(15.8)	(2339)	(2269)	–	(57)	–	(43)	(190)	–	–
Co	–	16	42	113	133	–	–	–	18	96	–	–
		(14.1)	(38)	(112)	(133)				(19)	(87)		
Cr	–	–	22	2800	3950	–	150	–	55	439	–	–
			(17.6)	(2730)	(4000)		(123)		(56)	(450)		
Cu	36	64	16	13	–	–	45	53	24	126	–	–
	(33.3)	(59.7)	(18.4)	(11.3)			(45)	(47)	(23)	(135)		
Zn	94	108	131	42	–	–	112	166	240	–	258	–
	(98)	(84)	(120)	(36)			(107)	(190)	(219)		(250)	
V	54	123	398	30	–	–	182	79	88	540	–	–
	(52.9)	(125)	(399)	(30)			(180)	(96)	(89)	(520)		
Li	29	12	13	2	–	–	79	12	–	4	95	91
	(32.1)	(12)	(12.8)	(2)			(70)[1]	(8)		(4)	(86)	(85)
Rb	250	67	47	–	–	–	12	32	–	12	204	187
	(254)	(67)	(46.6)				(12)	(32)		(10)	(214)	(198)
Sr	–	662	333	–	–	–	–	396	–	267	263	302
		(657)	(330)					(380)		(270)	(270)	(310)
Y	34	26	44	–	–	–	29	–	–	–	151	874
	(30.4)	(21.3)	(37.5)				(26)				(136)	(750)
Nb	29	16	14	–	–	91	–	–	–	–	33	–
	(29)	(15)	(13.5)			(92)					(27)	–
Zr	490	221	190	–	–	–	101	–	–	103	274	323
	(500)	(225)	(190)				(105)			(110)	(270)	(320)

In brackets bibliographic averages or recommended values: GSP-1, AGV-1, BCR-1, PCC-1, DTS-1, BM, data from Flanagan (1973); BR, data from Roubault et al. (1970); T-1, data from Thomas (1963); SY-1, data from Abbey (1975); MGR-1, SY-2, SY-3, data from Abbey et al. (1975)

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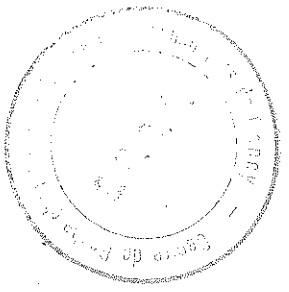

NEW INSIGHTS IN THE FRACTIONATION PROCESSES OF OPHIOLITES : EVIDENCE FROM CORSICA AND VOURINOS

M. OHNENSTETTER

INTRODUCTION

Ophiolites are one of the best control of the fractionation processes through plutonic and liquids relationships. The diversity of the ophiolitic lavas put constraints on the genesis of oceanic crust in different tectonic settings. On the basis of FeO and MgO contents, the Corsica and Vourinos ophiolite has been selected to approach the fractional crystallization processes and the coexisting liquids.

I - CORSICA

For Corsican liquids (fig. 1), fractionation occur at two main stages (OHNENSTETTER M., part. IV.4) :

A - The high pressure fractionation of :

1) Sp-Opx ± Pl occuring within the Seiland facies leeds to more alkali enriched picritic liquids which evolve towards the normal MORB.

2) Ol - Opx - Cpx - Pl ± Sp occur within the plagioclase lherzolite facies leading to picritic liquids which evolve later towards transitional MORB.

The existence in Corsica, Alps and Apennine of tholeiites of transitional and normal MORB (VENTURELLI et al., 1981) may be explain by the localization of the high pressure fractionation occuring within the rising mantle diapir. This may be also true for oceanic ridge tholeiites (OHNENSTETTER, 1978, MOREL, 1979, MOREL and HEKINIAN, 1980).

Like oceanic tholeiites (O'HARA, 1968) the Corsican basalts may be derived from liquids in equilibrium with olivine + orthopyroxene in the upper mantle.

B - The low pressure fractionation of Ol -Pl - Cpx with precipitation of troctolite and olivine gabbro. The original magma in equilibrium with olivine Fo_{92} is represented in table V . This calculated liquid show strong chemical analogy with MORB glasse analyses. It can be estimated that liquids outpourred on the surface only after 10 % of low pressure fractionation.

II - THE VOURINOS

The correlation between FeO and Zr may reveal some relationships between liquids belonging to a similar series. For example, in Krapa series the range of Zr content (BECCALUVA, et al.,1976) in andesite which encompass that of andesitic basalt (except for a sample) is an argument for no derivation of most of the former from the latter. A similar kind of evidence may be pointed out in the Asprokambo series between basalts and andesitic basalts often Zr-poor than basalts. Nevertheless it is possible to chuse samples which may define the magmatic trends linked to cumulate deposition under low pressure regime. A second step may be reached after having defined the usual fractionation by the composition of parental liquids supposed to be in equilibrium with mantle olivine (Fo_{90}).

A - THE KRAPA SERIES

In ophiolites all the geochemical line of evidence may be partly controlled by mass-balance calculations reflecting the nature and proportions of rocks. The least differentiated basalt has Mg/Mg+Fe ratio about .639 less than that of basalt (.777) thought to be in equilibrium with mantle olivine (Fo_{92}) according to ROEDER and EMSLIE (1970) like in Vourinos ultramafites (HARKINS et al., 1980). Consequently baselts are not primary magmas but the problem is how fractionation occur to obtain these derivative liquids. Fractionation may develop in cumulate dunite located in mantle harzburgite and/or in the lower ultramafic cumulates : dunites, wherlites, clinopyroxenite and olivine-melanogabbros. In this later case, the phases involved are Ol-Cpx-minor Opx and Pl. Compositions of extracts have been obtained using microprobe analyses of Cpx and Ol, or analyses of ultramafits cumulates mixed with olivine of different compositions (see fig. 2). In fig. 2 proportions of extracts varies from 22 % to 46 % using respectively dunite (Fo_{89}) or Cpx-rich cumulates. Extract E5 computed with a high proportion of Cpx (75 %) approaches those corresponding to nature samples mixed with 11 % Ol (Fo_{89}) (E2) and 13 % Ol (Fo_{87})(E3). If the lower cumulates, especially Cpx-rich cumulates, play a major role for the derivation of basaltic liquids, the first basalt which appears to the surface outpours after about 40 % fractionation. They are MgO-rich basalts, sometimes picritic, ranging from 13,65 % to 16,97 % MgO. Although not highly magnesian (MgO) 20 %) they are above the range defined by HART and DAVIS (1978) for parental liquids and slightly below the MgO content proposed by MALPAS (1978) for Bay of Islands ophiolite.

TAB. I - PARENTAL LIQUIDS FROM KRAPA

	E_1	E_2	E_3	E_4	E_5
SiO_2	40.81	48.51		47.91	
TiO_2		.11		.04	
Al_2O_3		4.54		1.02	
FeO	10.77	6.35	6.69	8.01	6.24
MgO	48.42	22.13	22.51	28.13	23.55
CaO		15.29		13.77	

	L_1	L_2	L_3	L_4	L_5
SiO_2	49.86	50.66		50.7	
TiO_2	.35	.30		.29	
Al_2O_3	12.32	10.73		10.18	
FeO	8.79	7.39	7.52	8.43	7.50
MgO	16.97	14.41	14.73	15.71	14.43
CaO	3.40	9.27		7.93	
Ni(ppm)		156			
Cr(ppm)		618			

Percentage of extract from the parental magma

E_1 = 22 % E_4 = 36 %

E_2 = 45 % E_5 = 41 %

E_3 = 46 %

E_1 = Fo_{89}

E_2 = ultramafits : 5 % Ol Fo_{89} + 95 % melagabbro calculated from the whole rock composition.

E_3 = ultramafits : 13 % Ol Fo_{87} + 87 % melagabbro calculated from the whole rock composition.

E_4 = wehrlite : 41.5 % Ol Fo_{85} + 58.5 Cpx calculated from the coexisting minerals.

E_5 = Ol clinopyroxenite : 25 % Ol Fo_{85} + 75 % Cpx calculated from the coexisting minerals.

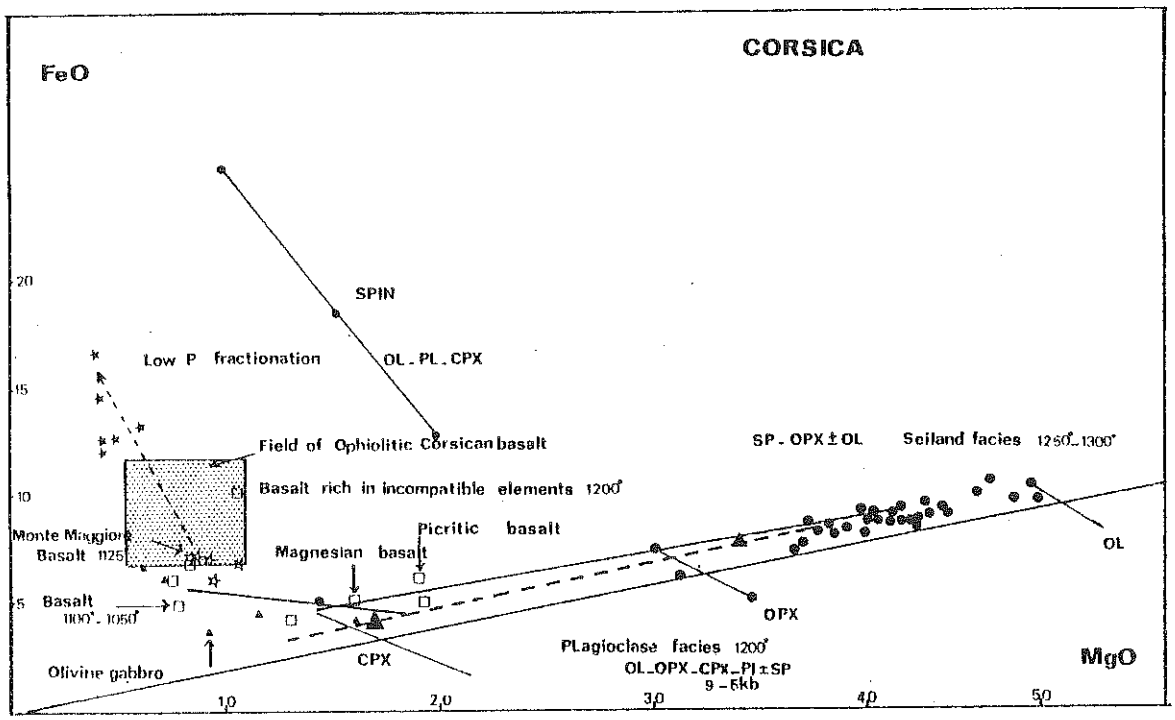

Fig. 1 - FeO-MgO diagram for Corsican sequence (see part IV).

The model proposed, based either on Cpx or Ol-rich cumulates may be tested when tempting to resolve the evolution of more differenciated liquid by fractionation of the upper cumulates. The andesites and andesitic basalts chosen are those used for trace element model. The observed and predicted liquids are reported in table 1 as well as the extract corresponding to the average of selected samples in defined proportions. Andesites may derive from andesitic basalts by 30 % fractionation of mt-bearing gabbro in agreement with the good correspondance between observed and calculated liquid. A less good fit (especially for SiO_2 and MgO) results when deriving andesitic basalts from basalts by a 55 % fractionation of noritic gabbros mixed with melanogabbros (proportion : 2/3-1/3).

The line between basalts and basaltic andesites indicate an apparent Cpx and Ol-rich extract (E7). Such a cumulate is rejected because calculated basaltic liquid has a too low Al_2O_3 content (13.43) compared to the observed composition. Therefore gabbro norite more or less mixed with melanogabbros must represent the extracted rocks between basaltic andesite and basalts. The discrepancies in the result come from different original basaltic composition linked to variation in previous fractionation (wehrlite-clinopyroxenite) or during partial fusion processes. Nevertheless it may be emphasized that Cpx fractionation from basalts may lead to FeO and Al_2O_3 richer and MgO lower basalt or andesitic basalt (consider sample AP_{10-6}).

Derivation of the successive liquids implies fractionation of 40 % of Cpx and Ol-rich cumulates to obtain basalts coming to the surface from parental magnesian basalts entering the magma chamber, 32 % of noritic gabbros and melanogabbros to get andesitic basalts and 8 % gabbro norite and Mt-bearing gabbros to give andesites. The schema proposed is compatible with an open system at least at the base in which magnesian liquid cooled and partly crystallized before mixed with the differenciated liquids and escaped towards the surface according to HUPPERT and SPARKS' model (1980). When gabbro norite precipitates, the chamber is nearly closed as the most evolved liquid fits closely the fractionation scheme.

B - THE ASPROKAMBO SERIES

Cumulates possibly playing a role in the Asprokambo series under low P regime are Pl-websterite, gabbro norite, norite and Q-diorite. Mixing calcula-

TAB. II - PARENTAL LIQUIDS FROM ASPROKAMBO

	Ol_{Fo89}	Opx	Cpx	E_1	E_2
SiO_2	40.81	57.3	54.6	53.69	52.08
TiO_2			0.01	0.03	0.03
Al_2O_3		1.02	1.11	1.24	1.09
FeO	10.77	6.7	2.39	9.04	9.26
MgO	48.42	34.4	18.6	23.41	26.54
CaO		0.82	23.50	10.48	9.17

Percentage of extract from the parental magma

	L	L'
Ol	16 %	17 %
Opx	20 %	21 %
Cpx	27 %	29 %
E_1	45 %	48 %
E_4	38 %	40 %

	L1	L'1	L2	L'2	L3	L'3	L4	L'4	L'5
SiO_2	50.64	52.59	53.47	55.48	53.07	54.88	53.04	54.37	53.83
TiO_2	0.10	0.22	0.17	0.21	0.16	0.19	0.13	0.15	0.17
Al_2O_3	13.20	11.24	12.77	10.91	11.76	9.94	9.19	7.64	0.56
FeO	7.57	7.69	6.91	6.98	5.73	5.71	7.90	8.01	7.94
MgO	14.63	14.80	13.44	13.49	11.19	11.01	15.04	15.35	15.37
CaO	7.39	6.63	7.20	6.48	12.77	12.49	9.56	9.19	8.46

Ol, Opx and Cpx : Mantle minerals composition
E_1 : pyroxenite calculated from coexisting minerals (60 % Opx - 40 % Cpx)
E_2 : Ol pyroxenite : 12.5 % Ol_{Fo89} + 87.5 % pyrox. calculated from coexisting minerals.

TAB. III - LOW PRESSURE FRACTIONATION OF KRAPA LIQUIDS

	Mela gabbro	gabbro norite	Mgt-bearing gabbro	basalt 267 observed	basalt calculated	andesitic basalt 197 observed	andesitic basalt calculated	andesite 76 observed
SiO_2	49.47	48.19	44.71	52.41	50.47	53.24	54.50	58.5
TiO_2	0.15	0.13	0.25	0.45	0.37	0.66	0.62	0.78
Al_2O_3	5.11	14.88	17.33	15.80	15.64	16.37	15.78	15.12
FeO	5.80	6.12	8.81	8.23	7.56	9.32	9.27	9.47
MnO	0.12	0.11	0.17					
MgO	18.85	10.68	7.40	8.1	8.23	5.23	5.40	4.55
CaO	17.20	16.01	12.80	4.35	10.13	2.95	5.91	2.96
Na_2O	0.53	0.76	1.49	3.56	2.60	4.85	3.53	4.41

Mela gabbro = 1/3 wehrlite, 1/3 clinopyroxenite, 1/3 melagabbro
Gabbro norite = 3/8 Ol-O, 1/4 gabbro norite, 3/8 leucogabbro norite
Mgt-bearing gabbro = 1/2 gabbro norite, 1/2 ferrogabbro norite

41 % extraction of possible parental basalt to obtain the basalts at the surface
32 % extraction of Ol-gabbro norite to obtain andesitic basalts
8 % extraction of gabbro norite and ferronorite to obtain andesites

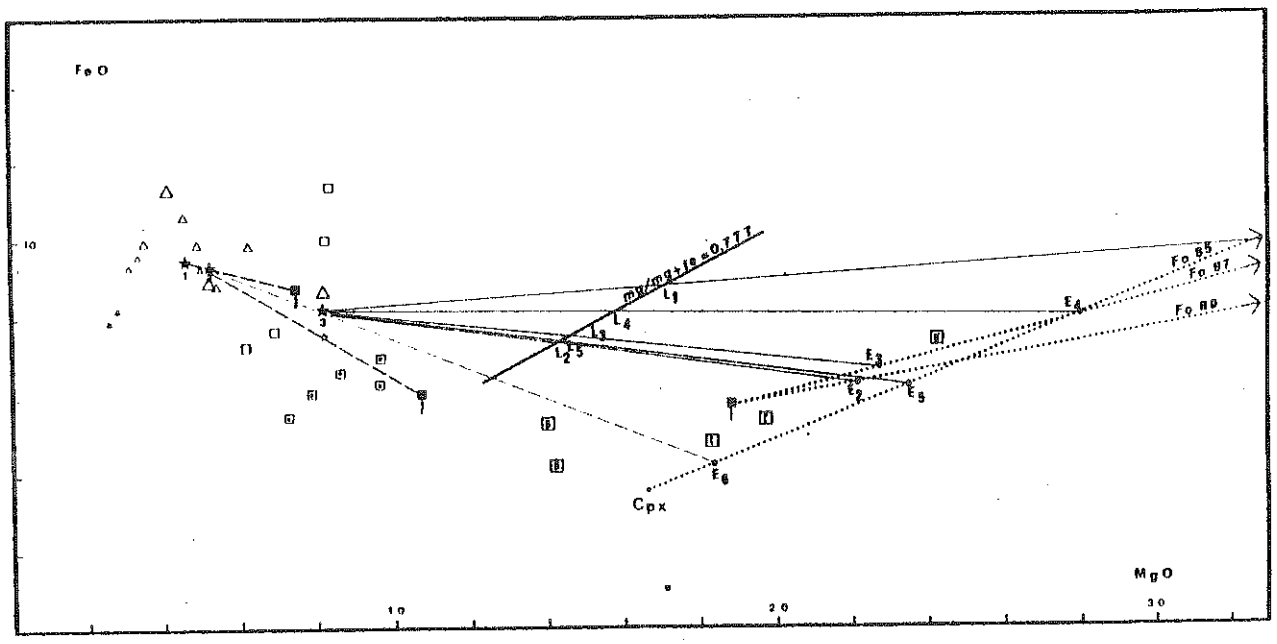

Fig. 2 -

FeO-MgO diagram for Krapa sequence : dacite, andesite, basaltic andesite basalt. Open square : observed cumulates. Letters in squares : W, wehrlite, C, clinopyroxenite ; M, melagabbro and plagiferous clinopyroxenite. Point in open square : gabbro norite ; open square : opaque-bearing gabbro norite. L1, 2, 3, 4, 5 : parental basalts calculated using extracts 1, 2, 3, 4, 5 and a basalt from Krapa. E6 is a Cpx-rich extract possibly formed during fractionation of basalt 3 to get basaltic andesite 2. Full squares : calculated extracts ; 1, ultramafite (1/3 wehrlite, 1/3 clinopyroxenite, 1/3 melagabbro) 2, mean from selected olivine-gabbro and gabbro-norite (3/8 olivine-gabbro, 1/4 gabbronorite, 3/8 leucogabbronorite ; 3, opaque-bearing gabbro-norite. Full star : liquids used for calculations ; 1, andesite ; 2, basaltic andesite ; 3, basalt. Open star : calculated basalt from an andesitic basalt.

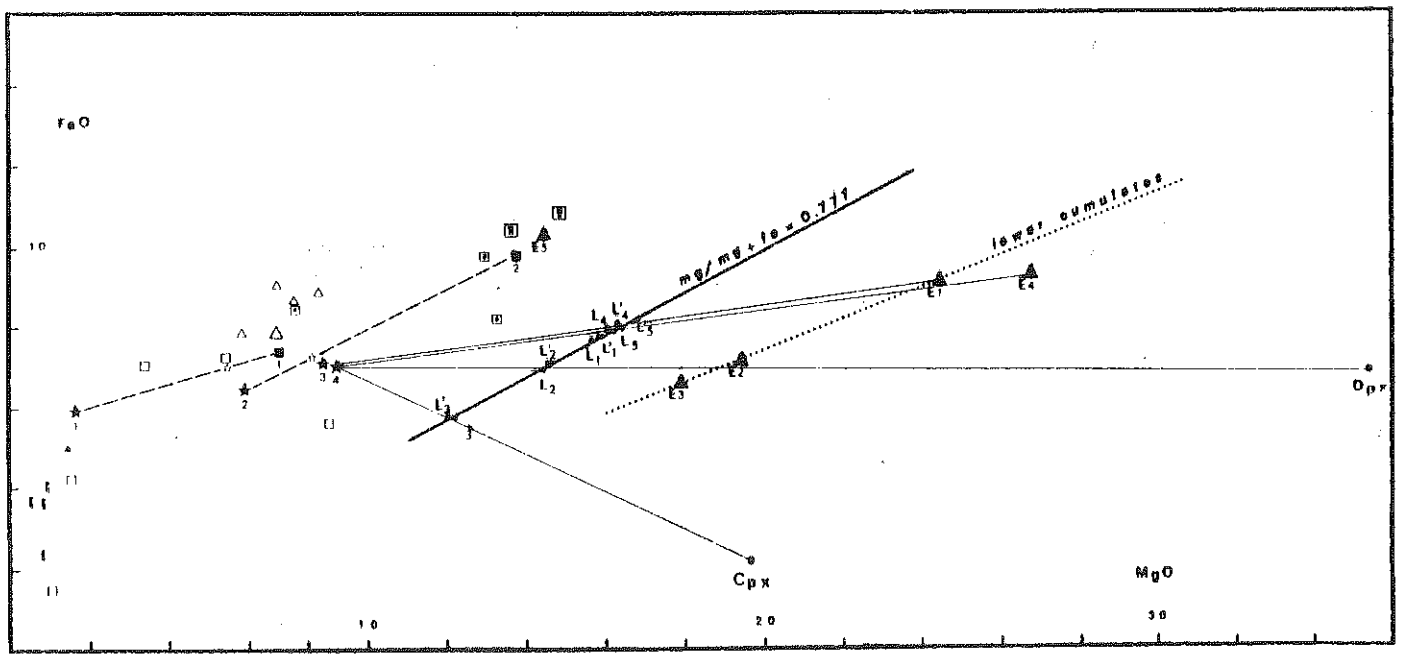

Fig. 3 -

FeO-MgO diagram for Asprokambo sequence : rhyolite ; basaltic andesite ; basalt. G, granophyre ; open square, quartz-diorite ; open square with a point, norite ; open square with W, plagiferous wehrlite. Full stars: selected liquids for calculations ; 1, dacite ; 2, evolved basaltic andesite 13 ; less evolved basaltic andesite 123 ; 4, basalt. Empty star, calculated liquids ; 1, basaltic andesite ; 2, less evolved basaltic andesite. L1, 2, 3, 4 parental basalts ; L1, 2, 3, 4 parental andesitic basalts. Cpx and Opx from mantle (HARKINS et al., 1980). Full square ; 1, norite with SiO_2 55 % ; 2, calculated cumulate from 50 % plagiferous wehrlite and 50 % gabbronorite. Full triangles, calculated cumulates from mineralogical data ; 1, pyroxenite (60 % Opx + 40 % Cpx) ; 2, plagiferous websterite (20 % plagioclase, 45 % Opx + 35 % Cpx) ; 3, plagiferous websterite (25 % Pl + 40 % Opx + 35 % Cpx) ; 4, websterite (10 % Ol_{Fo89} + 90 % extract E1) ; 5, plagiferous websterite with quartz.

TAB. IV – LOW PRESSURE FRACTIONATION OF ASPROKAMBO LIQUIDS

	Norite	Melagabbro norite	E_5	basalt 131 observed	andesitic basalt 123 observed	basalt calc E_5	basalt calc melagabbro norite	andesitic basalt 13 observed	andesitic basalt 13 calculated	Dacite
SiO_2	53.18	51.04	49.46	52.51	55.	54.14	54.54	55.70	55.73	63.38
TiO_2	0.24	0.19		0.21	6.27	0.15	0.19	0.19	0.26	0.32
Al_2O_3	16.16	9.25	7.18	15.70	13.54	12.76	13.28	14.62	15.34	12.90
FeO	7.33	9.67	10.15	6.96	7.06	7.33	7.21	6.39	6.98	5.91
MgO	6.78	12.72	13.54	8.20	7.92	7.78	7.57	5.86	5.52	1.72
CaO	8.90	13.39	12.22	8.80	7.99	10.22	10.51	9.55	7.33	2.60

E_5 : Qz bearing Pl-websterite calculated from coexisting minerals (36 % Opx, 36 % Cpx, 18 % Pl, 10 % quartz).

40 % extraction of the parental liquids to obtain the basaltic andesite

15 % extraction of melagabbro norite to obtain the evolved basalt and andesitic basalt

34 % extraction of norite to obtain dacite.

tions indicate that dacite II may derive from evolved andesitic basalts by about 75 % fractionation of Cpx-bearing norites (SiO_2 < 55 %) from basaltic andesites as suggested by similar composition between computed and observed (sample 13) liquids. Lower intervening cumulates are represented by Pl-websterites and gabbro norite. 25 % of an extract of these terms in similar proportion permits to derive sample 13 from a less evolved basaltic andesite (similar to sample 123) having also low content of incompatible element (Zr = 18). Calculations from mineral compositions give similar results. The observed basalts and low Si-basaltic andesites having lower Si and higher Al than calculated liquids using either pure wehrlite or norite extract cannot constitute parental liquids for evolved basaltic andesite (sample 13). A plagioclase phase may suppress the Si and Al discrepancy but would implies systematic lower Fe and Mg (about 6.4) as well as incompatible elements, facts usually not pointed out when the less evolved liquids are considered. Therefore, it appears that liquids more or less Si saturated may arrive at a high level in the crust to suffer Pl-websterite and gabbro norite fractionation. The composition of parental magma estimated either from basalt or andesitic basalts very mainly according to the fractionated phase. Magnesian basalts and basaltic andesites (14 < MgO < 15) are get using Ol (Fo_{89}) where as magnesian andesites (11 < MgO < 14) are obtained using pyroxenitic fractionation. The role of pyroxene, especially of orthopyroxene is suggested by the absence of Ol and the precocity of Opx in the lower cumulates as well as the presence of late pyroxenitic veins cutting the harzburgitic mantle (MOORES, 1969). In order to compare with Krapa, a theorical extract was computed using 9 % Ol and 35 % websterite computed from the most magnesian phases. It appears that Asprokambo parental magma, Si-richer, Al and incompatible elements-poorer than that of Krapa may be the partial fusion product of an impoverished mantle both in incompatible element and the more fusible phases. Earlier fractionation involves less Cpx in Asprokambo series. The degree of impoverishment in mantle and the earlier fractionation may respectively explained low Al VI and high Ca in Cpx from Asprokambo erupted liquids compared to Krapa. Parental liquids from Krapa and Asprokambo may be easily compared with Troodos ones for the second stage melting (DUNCAN and GREEN, 1980).

TAB V – COMPARISON OF PARENTAL MAGMAS

	L2 Krapa	L4 Aspro.	L'4 Aspro	Troodos 4	Tortuga NT23	Boninite C2	Average famous glass	Average MORB glass	calculated composition for Corsican parental liquid
SiO_2	50.66	53.04	54.37	52.40	47.25	54.19	50.23	50.68	48.49
TiO_2	0.30	0.13	0.15	0.30	0.79	0.18	1.02	1.49	1.05
Al_2O_3	10.73	9.19	7.64	11.7	13.64	9.57	15.52	15.60	15.35
FeO	7.39	7.90	8.01	8.4	9.77	8.40	9.17	9.85	6.33
MgO	14.41	15.04	15.35	15.80	17.61	12.79	9.11	11.44	11.44
CaO	9.27	9.56	9.19	10.7	9.58	8.54	11.85	11.44	7.96

Troodos 4 : parental magma of Qz tholeiites from the upper pillow lavas.

Tortuga NT 23 : Primary magma from the Tortuga ophiolite (ELTHON, 1979).

Boninite C2 : Boninite of Bonin Islands (KOMATSU, 1980)

Comparison of Corsican parental liquid with Mid-ocean ridge (MORB) glasses.

CONCLUSION

The calculated parental magma are different within the three ophiolitic series. In Corsica like in the oceans, a fractionation of Opx, Sp and Ol occurs at high pressure (≥ 15 Kbar) before extraction of Ol, Pl and Cpx in low pressure magma chambers. The outpourring of liquids may be expelled rapidly after their entry in the magma chamber after on as about 15 % of Ol-gabbroic-fractionation.

Then as depicted by O'HARA (1968) the oceanic magma as well as the Western mediterranean (e.g. Corsica) ophiolitic liquids are not primary liquids. In Eastern Mediterranean (e.g. Vourinos) basalts expelled from the magma chambers derived after about 40 % low pressure fractionation of mainly Ol+Cpx. At the end of these kind of ophiolite liquids are more diversified and have like Troodos (CAMERON, 1980) affinities with high magnesia andesites like boninites. The suitable source of such melts may be a Cpx poor lherzolite considered as a residue after MORB extraction (BECCALUVA et al., ms). These two stage model (DUNCAN and GREEN, 1980) may occur as outlined by CAMERON (1980) near fore arc area and may be related to hydrous melts of depleted upper mantle within a subduction zone. The production of liquids with andesitic affinities has been favoured by partial melting of spinel peridotites under wet conditions (O'HARA, 1965).

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TWO MAGMATIC SERIES WITH ISLAND ARCS AFFINITIES WITHIN THE VOURINOS OPHIOLITES.

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ABSTRACT

In the Vourinos ophiolites evidence of two episodes of magmatic activity is preserved in cognate cumulates, effusive and hypabyssal rocks, which constitute the earlier Krapa sequence and the younger Asprokambo sequence. Asprokambo dyke magmas, poor in incompatible elements and relatively Ni and Cr rich, bear some resemblance to Opx-bearing basalts erupted in island arcs. Krapa series magmas are best equated with low-K tholeiites erupted in immature island arcs. The Krapa sequence is constituted by a large volume of cumulates and overlying sills and massive and pillowed lavas, all of which display the following crystallization sequence : Ol $\pm$ Sp, Pl $\pm$ Opx, Mt. Périodic influx of fres h magma batches into the magma chamber occurred mainly during the formation of the lower cumulates (wherlite, Ol-clinopyroxenite, melagabbro adcumulates to mesocumulates). The upper cumulates, gabbro-norite, leuco-gabbronorite with minor Mt-bearing gabbronorite (meso to orthocumulates) crystallized in the upper levels of a magma chamber which became progressively smaller with time. They are cut by ortho to heteradcumulates belonging to the Asprokambo sequence which are websterites, Pl-websterite, gabbronorite, amphibole-bearing leuconorite, diorite and granophyre having Ol $\pm$ Sp, Opx, Cpx, Pl and Am as successively appearing phases. In cumulates, especially in the lower Krapa sequence, significant subsolidus reaction may have been favoured by the persistence of high geothermal gradients linked to continuous magmatism. In both sequences, the oldest basalts may be derived by fractionation of Ol-rich cumulates from a parental magnesian basalt in the Krapa sequence. Basaltic andesite may be produced by fractionation of melagabbro and gabbronorite, and andesite production involved fractionation of Mt-rich gabbronorites. In Asprokambo fractionation of Pl pyroxenites and gabbronorites give a wide range of basaltic and andesitic compositions. Precipita-

tion of Cpx-bearing norite produced dacite from andesitic basalts. Compositions of Al-and Ti-poor Cpx from lavas reinforce the similarities with island arc basalts. Asprokambo clinopyroxenes are richer in Ca and Cr than those from the Krapa sequence. Mineralogical differences between the Krapa and Asprokambo cumulates are pointed out. Importantly, the Asprokambo intrusives have igneous Mg-hornblende and chromian and V_2O_3-bearing aluminian titaniferous magnetite, wich are responsible for the calcalkaline evolutionary trend of these rocks. Major and trace element modelling necessitates a two-stage model for the petrogenesis of the Vourinos ophiolites. The primary magma parental to the Krapa sequence may have been produced by partial fusion of an impoverished lherzolite, leaving a harzburgitic residue. Asprokambo liquid compositions indicate that incongruent fusion of Opx possibly under more hydrous conditions played a greater role in controlling the chemistry of these magmas, leaving a refractory residue of dunitic composition.

I - INTRODUCTION

The compositional variations with time of some extrusive and intrusive rocks from ophiolites are generally considered to be linked to their generation in a single, or multiple geotectonic environments (PEARCE, 1975, ROCCI et al., 1975, SMEWING et al., 1975, PARROT, 1977, CHURCH and RICCIO, 1977, BECCALUVA et al., 1979, ALABASTER et al., 1980, PEARCE, 1980). In any particular geodynamic setting, the composition and evolution of successive derivative liquids following generation in the upper mantle may be partly controlled by low pressure fractionation leading to the production of cumulates (O'HARA, 1965). Based on geochemical arguements, several attempts have been made to correlate compositions of erupted liquids with underlying cumulates thought to have crystallized in either a closed or open system magma chamber under broadly equilibrium conditions (O'HARA, 1977, BECCALUVA et al., 1977, STERN, 1979, OHNENSTETTER et al., 1979, JACQUES and CHAPPELL, 1980, SERRI, 1980, PALLISTER and KNIGHT, 1981). Liquids entering the magma chamber may be of variable composition, leading to polycyclic and polygenetic cumulates (CHURCH and RICCIO, 1977, SMEWING, 1981).

An island arc origin has been proposed for the Vourinos ophiolite based largely on geochemical evidence (MIYASHIRO, 1973, 1977, BECCALUVA et al., 1979, OHNENSTETTER et al., 1979, NOIRET et al., 1981). Two main magma lineages have been recognized (PAUPY, 1976, BECCALUVA et al., 1979). Increasing complexity of the cumulate sequence in the upper levels (OHNENSTETTER et al., 1979) may be partly explained by intrusions of magma of different compositions towards the end of the differenciation history of an earlier magma chamber. Here, we attempt to follow the evolutionary course of the two main magmatic pulses undergoing low pressure fractionation, employing mass balance, major and trace element modelling. Some additional constraints regarding the development of cumulate piles are given, based on textural and mineralogical observations. The compositions of both the parental liquids which entered the magma chamber and their proposed mantle sources are estimated.

II - GENERAL GEOLOGICAL SETTING AND PREVIOUS STUDIES

The Vourinos ophiolitic complex has been tectonically emplaced over rocks of the Pelagonian Zone, from which it is separated by an amphibolitic sole

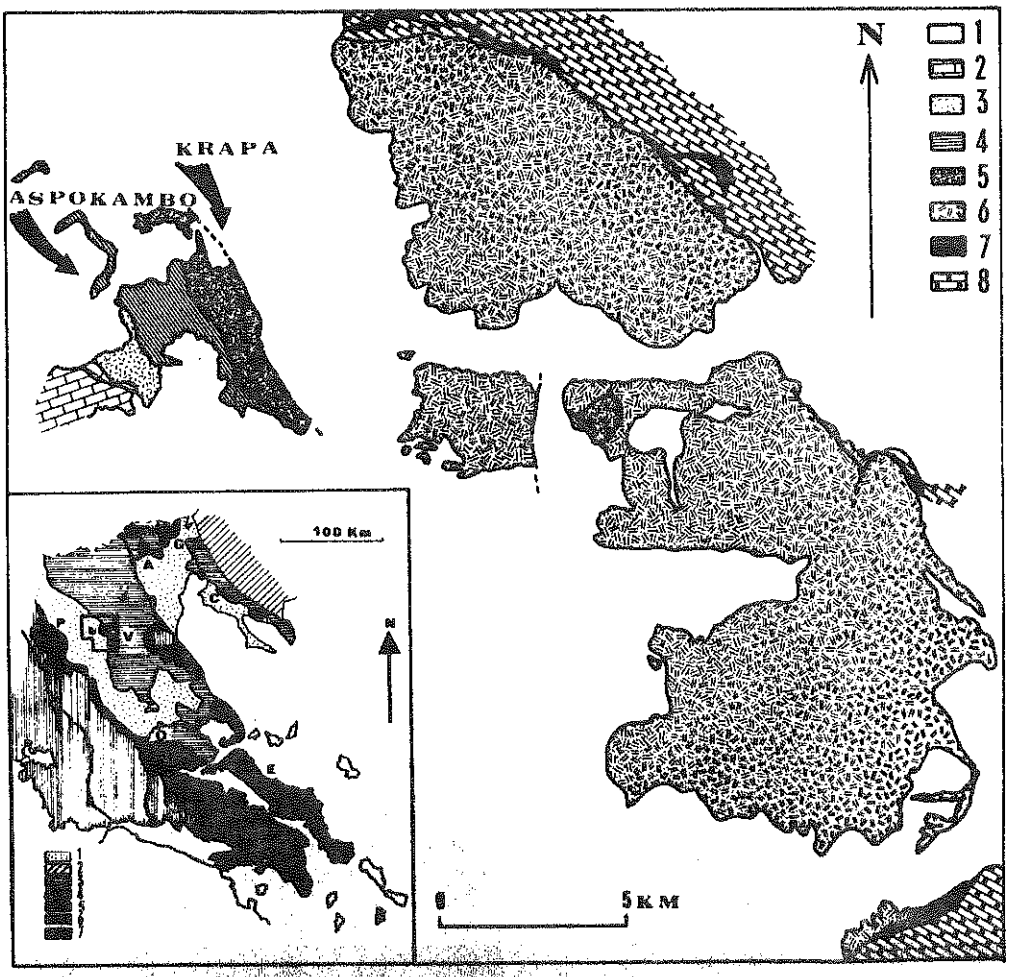


Fig.1 The Vourinos ophiolitic complex after MOORES (1969).

1. Tertiary and Quaternary
2. Radiolarian cherts and calpionellids limestones
3. Effusive and hypabyssal rocks
4. Gabbros and Quartz diorites
5. Transition zone
6. Ultramafics mainly harzburgites
7. Amphibolites and terrigeneous rocks underlying the ophiolitic complex
8. Pelagonian marbles

Sketch map of the Geology of Greece after BEBIEN et al. (1980).

1. Recent deposit
2. Serbo-Macedonian massif
3. Internal areas
4. Intermediate areas
5. External areas
6. Blueschists unit
7. Ophiolites: A=Almopias, E= Euboea, G= Guevgueli, O= Othris, P= Pindos, V= Vourinos

(BORTOLOTTI et al., 1969, PICHON and BRUNN, 1977) dated at 179 $\pm$ 4 M.Y. (SPRAY RODDICK, 1980). Radiolarian cherts of Upper Jurassic age overlie the ophiolitic lavas (BRUNN, 1956) and are themselves overlain by Calpionellid limestones (MOORES, 1969) of Tithonian to Berrasian age (PICHON, 1976, 1977 ; MAVRIDES, 1980) which contain both pelagic and reef facies. Later, an alternation of pelagic and neritic horizons suggests deposition in a continental slope environment before an Upper Cretaceous transgression and molasse deposition (PICHON and LYS, 1976).

NAYLOR and HARLE (1976) postulated that the Vourinos ophiolite was thrust from the Othris Zone toward the northeast, whereas VERGELY (1975, 1976, 1977) suggested that it was derived initially from the Vardar Zone to the northeast (Figure 1) by thrusting toward the west and southwest, and then it was thrust back toward the northeast during early Cretaceous retrothrusting.

The dominant lithology in the Vourinos ophiolite is mantle harzburgite considered by MOORES and VINE (1971) and ROSS et al. (1980) to have been generated in a diapir under an oceanic spreading centre with a high geothermal gradient (MERCIER, 1980). Coarse-grained dunite, peridotite and chromitite occur in slightly deformed isoclinal folds (AYRTON, 1968) with steep plunge and variable direction (MOORES, 1969). In the harzburgite, pyroxenite and some dunite dykes are linked to multiple magmatic events which occurred both before and after the deformation (HARKINS et al., 1980). The transition zone between ultramafic and mafic rocks comprises dunites interlayered with websterite and gabbro (MOORES, 1969).

III - TWO SEPARATE MAGMATIC SERIES IN THE VOURINOS OPHIOLITE

In the Vourinos extrusive and hypabyssal rocks, two different magma pulses, both with affinities to low-K_2O tholeiites, have been recognized. These correspond broadly to the Lower Pillow Lavas and some of the Upper Pillow Lavas of the Troodos ophiolite in Cyprus (BECCALUVA et al., 1979). On the basis of stratigraphic and petrological data, two main magmatic series may be defined ; the Krapa series forms most of the mappable units, while the Asprokambo series occurs mainly in the Aliakmon River area. In the more complex southern area, both Asprokambo and Krapa rocks occur in close proximity, but mutual relationships have not been determined.

A diagramm presenting the generalized internal relationships within the Vourinos Ophiolite is given in Figure 2. The lowermost units of the Krapa series are cyclic clinopyroxene-rich cumulates (JACKSON et al., 1975). With a progressive increase in the plagioclase content, ol gabbro and gabbro follow the ultramafic rocks and in turn are followed by noritic gabbros, ferronoritic gabbros and leucodiorites with variable amounts of interstitial quartz. The evolved, uppermost cumulates are often intruded and brecciated by doleritic sills. Massive flows and occasional dykes overlie the sills. Pillow lavas rest unconformably on these eroded flows (PAUPY, 1976).

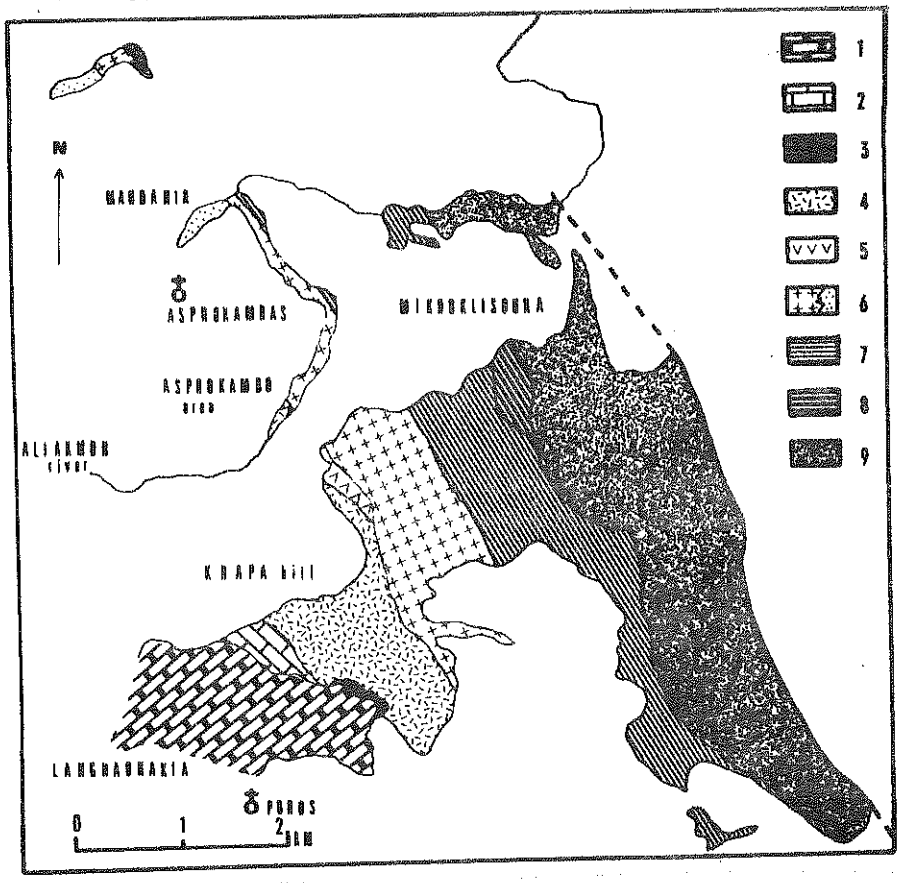


Fig. 2 Detail map (slightly modified after MOORES 1969) of the Krapa and Asprokambo area representing the upper and middle part of the Vourinos ophiolitic complex

In white : recent deposits

1. Cornstone and limestone series of Albian to Turonian age
2. Langhadhakia transgressive limestones of Tithonian-Berriasian age
3. Radiolarian cherts
4. Basaltic to andesitic pillows breccias and pillows lavas
5. Dacitic sills and dykes
6. Acid residues : crosses = diorites, stipped = Qz diorites and granophyres
7. Norites and noritic gabbros often layered
8. Interlayering of pyroxenites and norites
9. Transition zone

The Asprokambo series contains no well-developed layered cumulates such as are present in the Krapa series. The lowest exposed part of the Asprokambo series consists of websterite dykes in the Mikroklissoura-Aliakmon area (sections E, F and G in JACKSON et al., 1975). Higher in the sequence, norite and noritic gabbros follow pyroxenite, which locally forms abundant breccias with quartz gabbro and quartz diorite and granophyre, especially in the Mandania area. A sheeted complex is well-developed immediately above the breccias in the diorites. Lavas overlying the granophyre were erupted during or just after the crystallization of residual siliceous magmas (PAUPY, 1976).

IV - PETROGRAPHY OF EXTRUSIVE AND HYPABYSSAL ROCKS

1 - Krapa Series

Based on the classification of PECCERILLO and TAYLOR (1976), the Krapa lavas include basalts, basaltic andesites, andesites and dacites. Basalts display homogeneous intersertal textures with small, euhedral clinopyroxene phenocrysts and subhedral plagioclase phenocrysts ; relics of altered olivine are occasionally observed. While groundmass plagioclase microlites are generally skeletal, Cpx microlites are more stout prisms often intergrown at their margins with opaque oxides. Altered interstitial glass is usually present. Basaltic andesites are generally less glassy than basalts and contain common Cpx phenocrysts ; plagioclase occurs mainly as microlites and poikilitic laths in interstices. Both opaque oxides and quartz are more abundant than in the basalts. The degree of crystallinity increases approaching andesitic compositions and plagioclase, quartz and oxides become more volumetrically abundant ; the latter may occur as a phenocryst phase. Clinopyroxene is less abundant in andesites than the less evolved lavas, and occasional altered grains thought to have been Opx may be observed. In the andesites, abundant plagioclase define an intersertal texture. In dacites, glomerocrysts of plagioclase, orthopyroxene and opaque oxides occur scattered throught a more leucocratic groundmass.

Observations on the metamorphism are restricted to a continuous section approximately one kilometer thick. Metamorphic assemblages belong to the greenschist and lower amphibolite facies of burial metamorphism. Most of the extrusive rocks were probably strongly altered shortly after submarine eruption. Actinolite is more common in the massive flows and sills than in the pillow lavas, probably reflecting an increase in temperature with depth in the lava pile. While prehnite and calcite are the most abundant secondary calcic phases in basalts, pistacitic epidote is more frequent in the evolved rocks.

2 - Asprokambo Series

Asprokambo rocks are always dykes, and these range compositionally from basalts, through basaltic andesite and andesite to dacite and rhyolite. Basalts and most basaltic andesites are porphyritic and contain variable amounts and proportions of Cpx, Opx and plagioclase phenocrysts. Olivine may also have been present in some of the more primitive samples, but it has been destroyed during post-magmatic alteration. Textural evidence suggests that Cpx and Opx crystallized together prior to the appearance of plagioclase. Opx commonly mantles Cpx but the opposite has also been observed. Microlites of skeletal, zoned Cpx and subordinate plagioclase occur in the altered glassy groundmass of the basaltic rocks. As for the Krapa rocks, approaching andesitic compositions the degree of crystallization increases and the amount of clinopyroxene decreases as plagioclase becomes more abundant. Ca-poor pyx is a significant phenocryst phase and also occurs as microphenocrysts and coarse microlites. Granophyric intergrowths of quartz and sodic plagioclase often fill interstitial areas. In both dacite and sodic rhyolite, phenocrysts of plagioclase and quartz are common, while altered pyx phenocrysts occur in the dacites.

Asprokambo dykes are generally more altered than the Krapa samples. Adularia has developed within some plagioclase phenocrysts. Observed metamorphic parageneses are frequently indicative of disequilibrium (eg. coexistence of calcite, prehnite and epidote in single thin sections). Where secondary calcic minerals have replaced chloritized mafic phases, original textures have been destroyed leading to the production of epidosite. The relatively thin volcanic pile above dykes which display upper greenschist to lower amphibolite facies metamorphic assemblages imply metamorphism under an abnormally high geothermal gradient ; this probably occurred during hydrothermal activity associated with the creation of the ophiolite. As in the Troodos and Sarmiento ophiolites (GASS and SMEWING, 1973 ; ELTHON and STERN, 1978) a vertical metamorphic zonation linked to increasing pressure and temperature with depth in the crust exists in the Vourinos ophiolite, where metamorphic modification is considerably more pronounced in the sheeted dykes. The extent of metasomatism is controlled by P_{H_2O} of the circulating fluid and the geothermal gradient and depth in the crust. While this type of metamorphism is probably common and well-documented for oceanic crust (CANN, 1969), it may also occur in immature island arcs ; for example, the lavas dredged from the Palau-Kyushu Ridge (a remnant arc) are frequently strongly altered by hydrothermal fluid activity (ALDRICH et al., 1981, SCOTT, 1981).

V - MAGMATIC AFFINITIES OF THE VOLCANIC AND HYPABYSSAL ROCKS

Discrimination between the two magmatic series and their assignment to particular tectonic settings of eruption are based on the immobile elements Ti, Zr, Y, Nb, V and Cr (PEARCE, 1975). Other elements, including Si, Al, Fe, Mg, Ni and Co, even if they may be variably mobile during low-temperature alteration processes, have also been used where it could be shown by correlation with immobile elements that their abundances had not been significantly altered during alteration. All rocks show a sub-alkaline (qtz-tholeiitic) character with low K_2O contents and hith Y/Nb ratios. Comparison with modern volcanics using various discrimination diagrams (PEARCE and CANN, 1973, BECCALUVA et al., 1979) shows that both series fall in the island arc tholeiitic fields (Figure 3). Recent Nd isotopic studies and Hf/Th and Ta/Th ratios support this island arc assignment (NOIRET et al., 1981).

The Asprokambo dykes have higher Ni and Cr contents than the Krapa basalts ; the latter have high Ti/Zr ratios (100-135) and low Zr/Y ratios such as are typical of island arc tholeiites (EWART et al., 1977, GILL, 1976, SUN and NESBITT, 1978, BECCALUVA et al., 1980, MATTEY et al., 1981), and are distinguishable from most MORB (SUN et al., 1979, SAUNDERS et al., 1980).

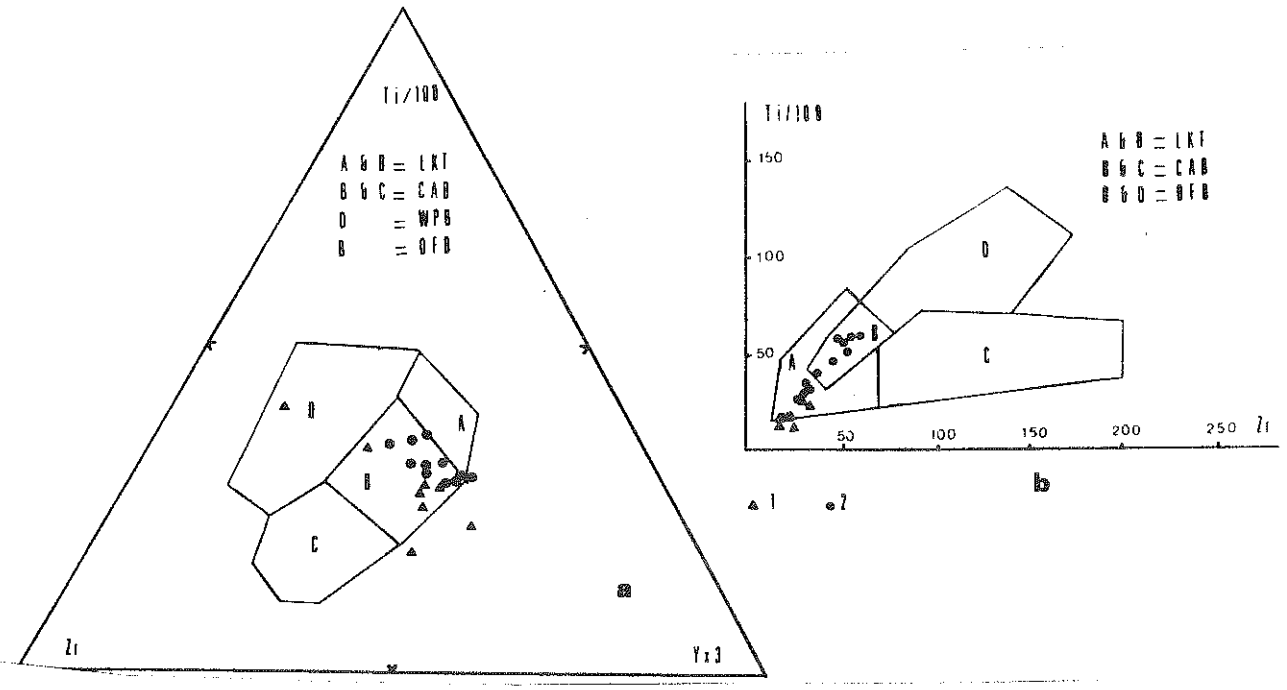
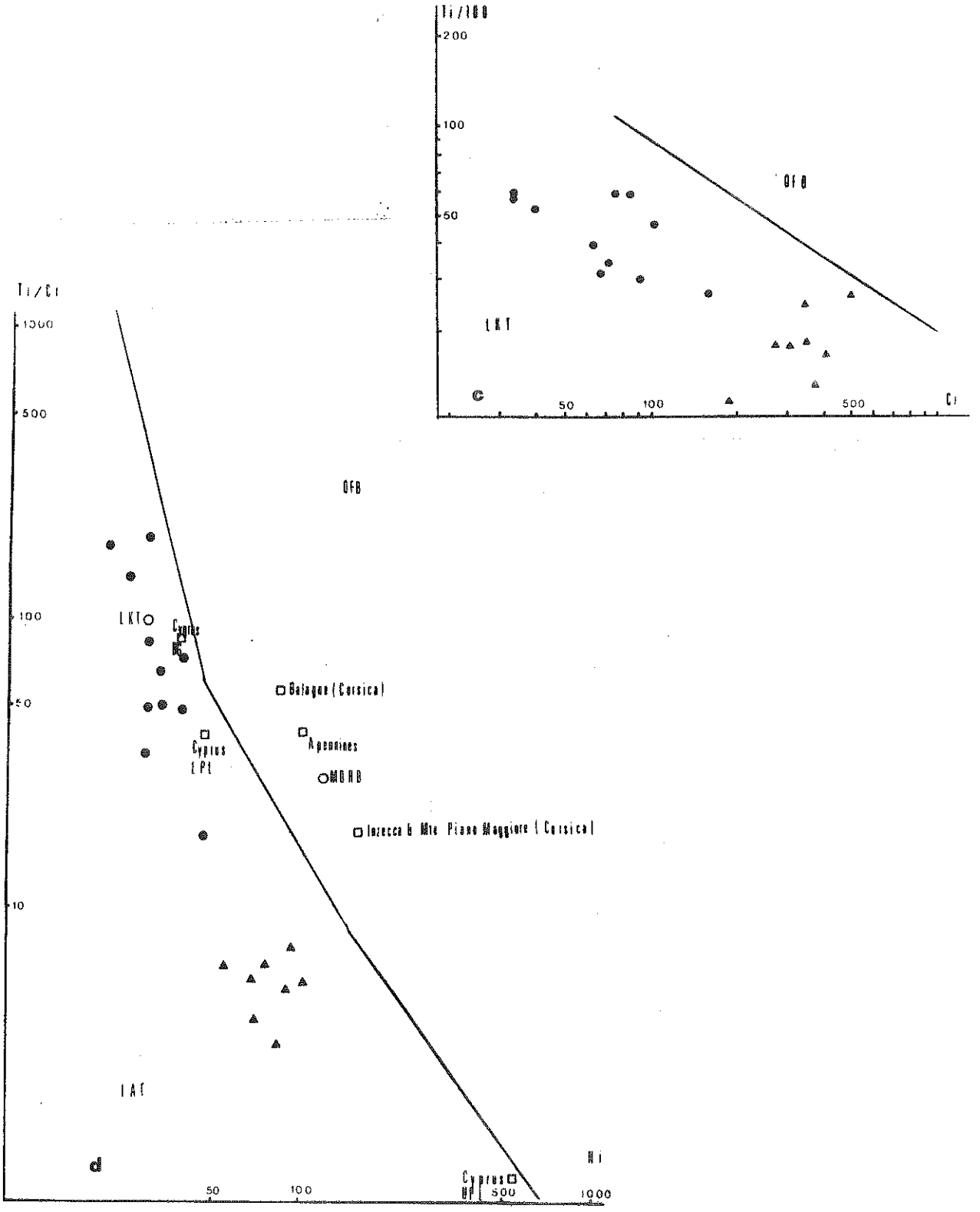


Fig. 3 Discrimination diagramms.

Dots = Krapa, triangle = Asprokambo

3 a) Ti/100; Zr; 3.Y after PEARCE and CANN (1973), LKT= low K_2O tholeiites, CAB = calcalkali basalts, WPB = within plate basalts, OFB = ocean floor basalts

3 b) Ti vs Zr (PEARCE and CANN, 1973) same symbols as in Fig 3a



3 c) Ti vs Cr (PEARCE, 1975) same symbols as in Fig. 3a

3 d) Ti/Cr vs Ni (BECCALUVA et al., 1979) LKT = low K tholeiites, Cyprus BG and UPL = mean composition of Cyprus basalts from the basal group and from the upper pillow lavas, OFB = ocean floor basalt , mean composition of the mid-ocean ridge basalts (MORB), the Western Tethys ophiolitic basalts : Balagne and Inzecca (COrsica) and the Apennines.

The Asprokambo basic dykes have lower Ti/Zr (79-99), Zr/Y (18-22) and Al_2O_3/TiO_2 (33-77) ratios than Krapa basic rocks thus showing affinities with very low-TiO_2 basalts found solely in subduction-related island arc or interarc basin settings (EWART et al., 1977, GUST and JOHNSON, 1981, MEIJER, 1980). Supporting the geochemical assignment of the Krapa and Asprokambo rocks to an island arc setting is the fact that their crystallization sequence following initial Ol + chromite separation (Cpx-plag-Opx+opaques for Krapa, Opx+Cpx-plag-qtz for Asprokambo) are both quite different from the well-known MORB crystallization sequence (Ol+chromite-plag-Cpx-opaques+low-Ca pyx : CAMERON et al., 1980).

A comparison of samples from both series with SiO_2 less than 56 % (Table 1) shows that the Krapa rocks are richer in Al_2O_3, FeO, MgO and incompatible elements and poorer in Ni and Cr than the Asprokambo lavas. The more oversaturated character of the Asprokambo relative to the Krapa samples is indicated by the distributions of the two series in Fig. 4, the early appearance of Opx, the presence of qtz-albite myrmekite in basaltic andesites and by qtz phenocrysts in Asprokambo dacites.

The most striking characteristic of the Asprokambo basaltic rocks is their unusually low contents of the incompatible elements Ti, P, Zr and Y (TiO_2 0.19-0.44 % ; P_2O_5 0.01-0.02 % ; Zr 16-31 ppm ; Y 9-15 ppm), associated with low Al_2O_3 (13.6-15.7 %) and the typically calc-alkaline lack of Fe, Ti and V enrichment during differentiation. These features suggest some similarities with the boninitic lavas found in the forearc regions of oceanic island arcs (DIETRICH et al., 1978, CAMERON et al., 1979 ; MEIJER, 1980 ; CRAWFORD et al., 1981). A high-MgO, low TiO_2 Vourinos basalt (V3) analyzed by MONTIGNY (1975) has a V-shaped REE pattern such as have been recorded for boninites (SUN and NESBITT, 1978), further reinforcing the similarities between Asprokambo rocks and boninites. However, while both boninites and Asprokambo rocks are highly porphyritic and/or vitrophyric, and display crystallization sequences marked by the early appearance of Opx, significant differences do exist between the two . Asprokambo samples contain less glass and zoned, skeletal pyx, and contain no Mg-chromite, clinoenstatite or amphibole ; moreover, they show a greater proportion of Cpx relative to Opx and contain phenocrystal plagioclase, which is never present in boninites.

Clinopyroxenes from Asprokambo and Krapa basalts, basaltic andesites and andesites evolve from endiopside to augite with only slight Fe-enrichment (Table 3 and Figure 4).

Table 1 - Chemical analyses of effusive and hypabyssal rocks from the Krepa series

	Massive lavas			lava flows			pillow lavas					Sills				Dykes	
	B AP-26-7	BA AP-7-7	A AP-28-7	BA AP-19-7	A 3-17-7	D AP-4-7	B AP-24-7	BA AP-20-7	B AP-22-7	E AP-23-7	BA AP-18-7	BA AP-13-7	A AP-9-6	B AP-10-6	D AP-1-7	BA 2-17-7	A 1-17-7
SiO_2	52.41	55.32	57.75	53.24	62.50	64.60	51.15	53.45	50.52	52.60	53.34	53.65	58.50	48.92	64.25	53.10	57.00
TiO_2	0.45	0.99	0.87	0.66	0.72	0.93	0.58	0.67	0.53	0.50	0.96	0.99	0.78	1.01	0.73	0.79	1.00
Al_2O_3	15.80	14.67	15.65	16.57	13.15	13.51	16.00	14.74	16.53	15.71	14.90	15.40	15.12	18.42	13.83	15.11	14.42
$Fe_2O_3T.$	9.15	10.96	10.30	10.35	10.40	9.16	9.70	10.84	9.95	9.86	11.76	10.95	10.52	12.52	8.77	9.75	10.37
MgO	8.10	3.50	3.10	5.23	3.38	2.83	7.15	6.18	6.16	5.15	4.50	4.92	4.55	4.12	2.63	5.37	5.01
CaO	4.35	4.00	3.82	2.95	1.69	1.30	4.61	2.28	6.88	6.19	1.83	3.03	2.86	8.15	1.22	4.32	2.55
Na_2O	3.56	5.85	4.99	4.85	4.18	5.22	3.80	4.64	3.51	3.78	4.74	5.59	4.41	3.85	5.02	5.26	5.18
K_2O	1.54	0.43	0.09	1.23	0.05	0.15	1.16	0.32	0.43	0.48	0.38	0.05	0.37	0.33	0.16	0.12	0.15
P_2O_5	0.02	0.06	0.05	0.03	0.05	0.07	0.03	0.05	0.03	0.02	0.05	0.05	0.05	0.06	0.07	0.05	0.06
PF	4.91	2.63	2.94	5.02	3.38	2.62	5.17	5.15	5.62	6.02	6.57	3.87	3.16	1.78	2.64	4.20	3.68
Tot.	100.19	98.41	99.56	100.13	99.50	100.19	99.35	98.52	100.16	99.29	99.03	98.50	100.32	99.16	99.33	98.07	99.42
Ni	47	30	28	33	22	13	34	26	30	29	22	39	27	30	20	39	20
Co	39	40	27	37	34	16	38	41	42	38	34	40	34	47	14	38	31
Cr	130	62	93	54	67	65	58	32	55	74	28	69	62	28	52	83	84
Cu	82	9	156	155	131	6	140	97	93	74	104	139	22	18	6	9	175
Zn	101	103	138	97	116	67	96	106	84	82	102	96	80	82	96	90	110
V	241	342	279	294	235	116	354	309	304	295	317	370	300	443	177	263	270
Sr	87	72	164	72	47	40	146	82	79	69	83	45	74	125	36	62	67
Li	13	3.3	5.9	8.2	3.9	4.6	15	11	12	11	11	5.7	7.9	3.4	4.6	5.2	5.6
Rb	11	6	<5	13	6	<5	10	<5	6	5	7	<5	5	7	<5	6	<5
Mo	<2	3	3	<2	3	4	<2	3	<2	<2	3	3	3	3	4	3	3
Zr	27	53	42	35	45	64	29	51	31	28	51	44	47	58	66	44	53
Y	11	25	21	16	22	32	11	25	13	11	26	19	20	24	32	21	25
Zr/Y	2.4	2.1	2.0	2.2	2.0	2.0	2.6	2.0	2.4	2.5	2.0	2.3	2.3	2.4	2.1	2.1	2.1
Ti/Zr	100	112	124	113	96	87	120	102	103	107	113	135	100	104	66	108	113
Al_2O_3/TiO_2	35	15	18	25	18	14	28	17	31	31	15	16	19	18	19	19	14
FeO^*/MgO	1.0	2.8	3.0	1.8	2.8	2.9	1.2	1.6	1.5	1.7	2.4	2.0	2.1	2.7	3.0	1.6	1.9

B = basalts BA = basaltic andesites A = andesites D = dacites R = rhyolites

Table 2 - Chemical analyses of dykes from the Asprokambo series

	B AP-13-1	BA AP-12-3	BA AP-9-3	BA AP-20-1	B AP-19-1	B AP-3-4	BA AP-4-4	BA AP-1-3	D AP-1-1	R AP-5-1	B V3*
SiO_2	52.51	55.00	55.35	54.50	52.89	51.85	53.98	55.70	68.39	72.40	47.92
TiO_2	0.21	0.27	0.30	0.28	0.28	0.44	0.41	0.19	0.32	0.25	.24
Al_2O_3	15.70	13.54	13.58	14.45	15.16	14.35	14.90	14.62	12.90	10.90	12.15
$Fe_2O_3T.$	7.73	7.85	9.80	9.57	10.00	8.68	8.70	7.10	6.57	5.50	9.94
MgO	8.20	7.92	7.76	7.22	6.75	6.75	5.92	5.86	1.72	1.45	12.29
CaO	8.80	7.99	7.22	6.49	6.25	12.06	10.97	9.55	2.60	3.10	8.42
Na_2O	2.56	3.02	2.86	3.31	4.02	1.21	0.75	2.54	4.18	3.51	=
K_2O	0.30	0.14	0.17	0.28	0.19	=	=	0.15	0.12	0.03	0.05
P_2O_5	0.01	0.01	0.01	0.01	0.01	0.02	0.02	0.01	0.03	0.04	=
PF	3.43	3.47	3.12	3.88	3.86	3.66	3.44	3.32	2.06	1.75	3.97
Tot.	99.45	99.21	100.17	99.99	99.41	99.02	99.09	99.04	98.87	98.93	94.98
Ni	86	71	92	70	78	105	96	57	31	15	
Co	43	43	40	35	36	44	42	40	25	16	
Cr	311	332	289	249	223	402	285	154	92	81	
Cu	54	104	134	97	121	20	45	23	27	22	
Zn	74	72	87	112	115	63	70	94	78	64	
V	290	255	304	304	256	278	313	215	93	24	
Sc	89	78	53	75	75	133	165	59	78	73	39.5
Li	4.4	2.6	3.7	7.9	4.5	1.9	1.7	4.1	3.3	3.4	
Rb	5	5	4	5	5	3	4	6	5	< 5	.6
Nb	<2	<2	<2	<2	<2	<2	<2	<2	3	4	
Zr	16	18	19	17	20	27	31	24	35	37	
Y	9	10	10	9	9	13	15	11	24	25	
Zr/Y	1.8	1.8	1.9	1.9	2.2	2.1	2.1	2.2	1.5	1.4	
Ti/Zr	79	90	95	99	84	98	79	48	55	42	
Al_2O_3/TiO_2	75	50	45	52	54	33	36	77	40	44	50
FeO/MgO	0.9	0.9	1.1	1.2	1.3	1.2	1.3	1.1	3.4	3.4	0.73

* Analysis from Montigny (1975) Symbols as in table 1

Krapa Cpx phenocrysts have slightly lower Ca and Cr contents than those from Asprokambo rocks. Pyroxenes from both suites have very low Ti (< 0.15) and Al^{IV} (< 0.13) contents, in agreement with the host rocks low incompatible element contents and SiO_2 saturation (Fig. 6). In figure 6a, they occupy the same compositional field as Cpx from Tongan basaltic andesites and andesites (EWART, 1976) and Cpx from arc basalts from dredge sites 1404 and 1427 on the nearshore slope of the Mariana Trench (BECCALUVA et al., 1980). As boninitic clinopyroxene usually controlled as quench microlites, their minor element chemistry is strongly controlled by cooling rate (GROVE and BENCE, 1977) and hence meaningful comparison of the compositions of Vourinos lava Cpx phenocrysts and boninitic Cpx is ruled out.

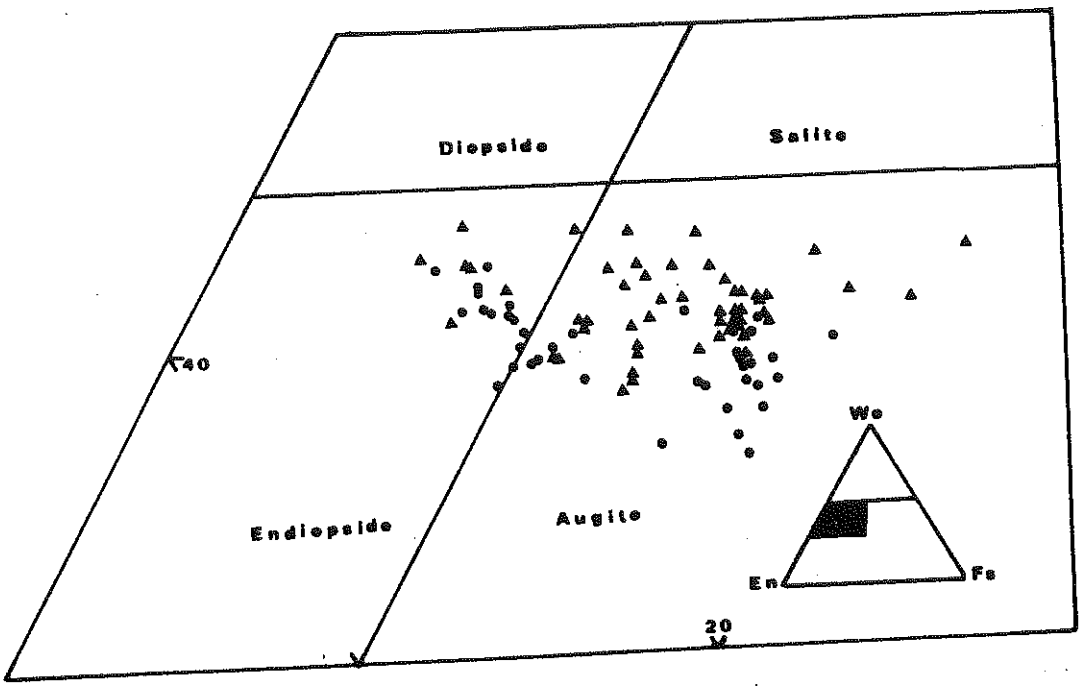


Fig. 4 Clinopyroxenes composition of the effusive and hypabyssal rocks reported in the enlarged zone of the Wo-En-Fs triangle. Dots = Krapa, triangle = Asprokambo

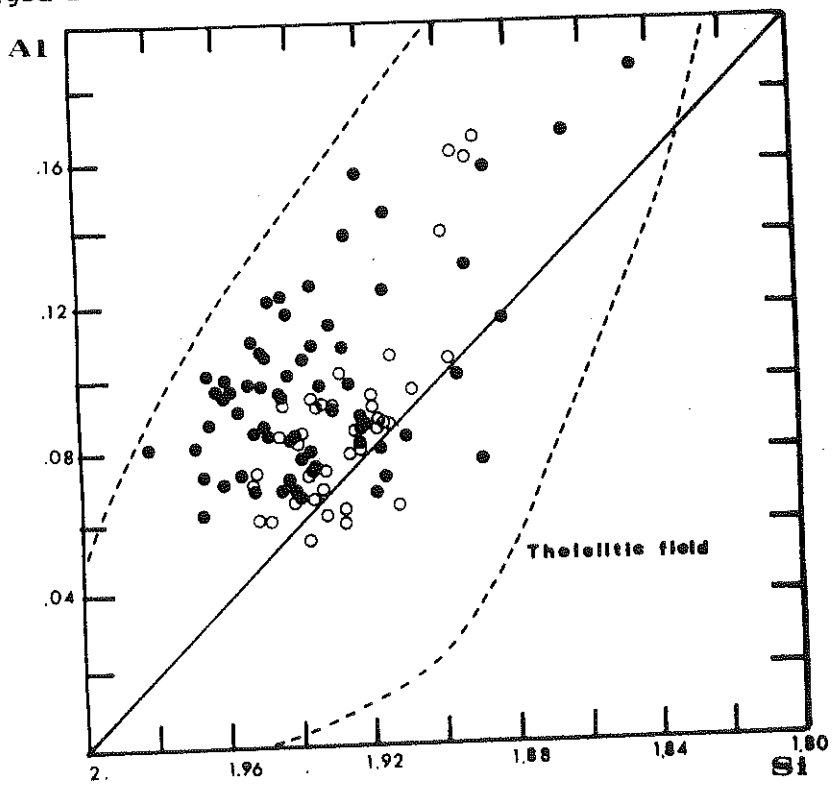


Fig. 5 Atomic proportions of Si and Al in clinopyroxenes from the effusive and hypabyssal rocks from the Vourinos . Solid circle = Krapa, open circle = Asprokambo. They plot within the tholeiitic field of KUSHIRO (1960).

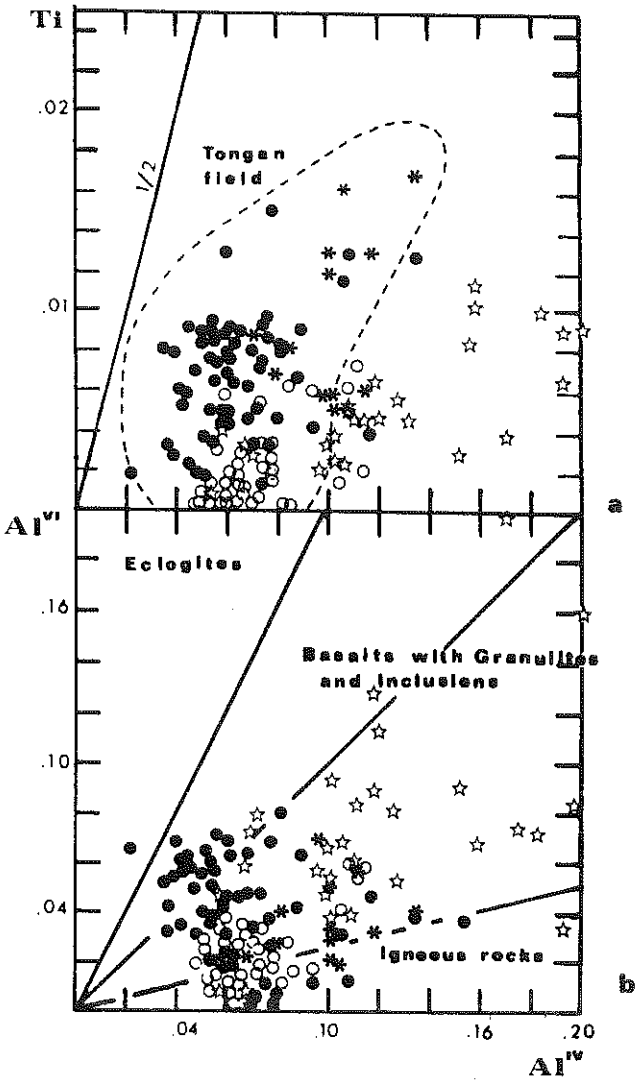


Fig. 6 a Ti vs Al IV. The clinopyroxenes from the Vourinos (same symbols as in Fig. 5) plot within the field of Tongan pyroxenes (EWART, 1976) and close to the composition of island arc tholeiitic dredged near the Mariana (BECCALUVA et al;1980) as well as boninites (SHIRAKI and KURODA, 1977; KURODA et al. ,1978; CAMERON et al., 1980; KOMATSU, 1982; SHARASKIN et al., 1980; OHNENSTETTER et al. (in prep). Note that the LKT and boninites lie relatively far away from the 1/2 ratio of the hypothetic Ca-Ti tschermak's molecule. Note also the very low Ti content of the Asprokambo pyroxenes.

Fig. 6b Al IV vs Al VI diagramm and fields defined by AOKI and SHIBA (1973). Same symbols as in Fig. 3. Note the lower Al VI content of the Asprokambo clinopyroxenes

TABLE 3 Selected analyses of clinopyroxenes of effusive and hypabyssal rocks from Krapa and Asprokambo series

KRAPA	AP 267 - B			AP 197 - BA			AP 95 - A				
	002 C	0014	008 A	0005	0024	0027	0093	0678	069 C	072 C	073 B
SiO_2	53.78	52.18	54.01	52.76	52.11	51.25	51.09	52.33	51.37	52.51	52.56
TiO_2	.07	.14	.09	.20	.30	.52	.15	.30	.29	.26	.25
Al_2O_3	2.28	2.18	1.73	2.23	1.94	3.18	1.80	1.61	1.86	1.93	1.68
FeO	4.83	9.60	6.78	9.96	10.34	11.37	10.24	9.60	9.79	9.72	10.47
MnO	.29	.25	.03	.09	.33	.26	.40	.38	.15	.21	.24
MgO	18.03	16.70	17.99	15.30	15.88	14.03	16.02	15.52	15.65	15.48	15.32
CaO	20.14	17.68	19.77	18.92	18.26	18.90	20.48	19.81	20.22	19.99	20.04
Na_2O	.06	.20	.04	.12	.16	.02	.14	.09	.10	.18	.13
K_2O	.00	.00	.01	.02	.00	.00	.00	.00	.02	.00	.00
Cr_2O_3	.24	.00	.02	.10	.00	.00	.00	.00	.00	.12	.00
Total	99.72	98.93	100.47	99.70	99.32	99.53	100.32	99.64	99.45	100.40	100.69

ASPROKAMBO	AP 131 - B					AP 123		BA			
	C 007	B 008	0012	009	A014	052 B	028 C	029 B	037	030	034 B
SiO_2	50.67	53.01	52.94	52.79	53.08	52.68	53.75	51.91	50.95	51.15	52.38
TiO_2	.06	.02	.05	.00	.00	.00	.00	.05	.21	.21	.04
Al_2O_3	3.81	1.73	1.51	1.46	2.19	2.07	1.67	2.21	2.43	3.74	1.91
FeO	8.17	7.76	7.19	7.39	3.73	5.74	4.77	8.35	10.44	9.47	9.88
MnO	.13	.04	.26	.25	.18	.22	.01	.35	.01	.00	.27
MgO	15.20	16.62	18.14	17.57	17.85	18.16	18.37	16.34	14.24	15.24	16.44
CaO	21.05	20.67	20.00	20.42	22.18	21.16	21.45	20.40	21.02	20.15	18.93
Na_2O	.01	.10	.11	.10	.06	.08	.04	.05	.05	.00	.05
K_2O	.00	.05	.00	.00	.00	.00	.00	.00	.00	.00	.01
Cr_2O_3	.04	.00	.40	.28	.69	.13	.08	.16	.05	.08	.00
Total	99.14	100.00	100.60	100.26	99.96	100.14	100.15	99.93	99.47	100.13	99.96

In summary, the Asprokambo dykes constitute a very low-TiO_2 suite bearing some ressemblance to boninites, high Mg andesites and opx-bearing basalts erupted in island arc settings associated with low K_2O tholeiites, which in Vourinos are represented by the Krapa series.

VI - PETROLOGY OF THE CUMULATES

1 - Stratigraphy of the layered sequence

Two main cumulate series may be recognized in the fossil Krapa-Mikro-Klissoura magma chamber based on textural and mineralogical variations. The lower of these is constituted by layered, coarse-grained clinopyroxenites and wehrlite adcumulates in which unit boundaries are marked by grain size and ratio contacts. Tectonic recrystallization is occasionally observed. As plagioclase becomes more abundant higher in the sequence, plag-clinopyroxenites and melagabbro adcumulates to mesocumulates are finer-grained (1-1.5 mm) than early cumulates (2-3 mm) in which lack cumulus plagioclase. In the lower cumulates, the crystallization sequence appears to have been olivine-Cpx-plag-Opw. The frequent occurrence in these rocks of interstitial finer-grained gabbronorite with subhedral minerals may testify to some entrapment of residual liquid during crystallization of the plag-poor cumulates.

Cumulates from the upper Krapa series are finer-grained and form a well-layered body marked by ratio contacts. From the base, to the top in gabbronorite, grain size decreases (1-0.3 mm) and Opx and plag become more abundant. Igneous lamination is mainly defined by plagioclase in orthocumulates high in the sequence. Opaque minerals appear before quartz and amphibole. In the uppermost gabbronorite, Cpx inverted from pigeonite occurs as subpoikilitic grains and is often altered to secondary amphibole. Small interstitial and large poikilitic amphibole also occur in heteradcumulates, the latter forming coarse-grained patches. High in the upper series, ferrogabbro norites are associated with amph-bearing gabbro, ferrodiorite and quartz-bearing leucodiorite. Towards the end of differentiation, the abundance of Cpx increases at the expense of Opx. Strong compositional zoning is only well-developed in plag from diorite and related rocks.

The Asprokambo series is constituted by dykes, and some small intrusive bodies of dioritic composition which are themselves intruded by more basic dykes. Those Asprokambo rocks with volcanic (porphyritic) textures have already been described. The remainder show holocrystalline plutonic textures which are best described by the WAGER et al. (1960) terminology for cumulate rocks, although a process of gravitative accumulation is not implied for the Asprokambo rocks to which this terminology has been applied. Cumulate-textured rocks are mainly heterad-

cumulates and orthocumulates with minor mesocumulates which range compositionally from websterite to norite and quartz diorite. The apparent crystallization sequence for these rocks is olivine + opaques - Opx - Cpx - plag amph + qtz. Less evolved members which contain Cr-rich ulvospinel have an igneous lamination defined by Opx euhedra. Websterite and norite alternate with gabbro-norite which is constituted by euhedral Opx, and subhedral Cpx and plag, with interstitial sodic plag, amph and qtz. Locally, amph-rich veins may interrupt the layering and are of broadly qtz-gabbro to dioritic composition. The coarse-grained igneous amphs contain rounded, reacted relics of Cpx and Opx and their alteration products which include plagioclase and qtz. Quartz-rich gabbro and diorite alternate with gabbronorite and plag-websterite and contain strongly-zoned plagioclase and skeletal opaques (MOORES, 1969). The granophyres from the Mandania River contain amph in addition to quartz and plagioclase (PAUPY, 1976). Narrow dykes of plag-pyroxenite composition intrude both Krapa and Asprokambo cumulates; these have chilled margins with quench textures and phenocrysts of Cr-rich ulvospinel. The interiors of these dykes are coarser-grained (1 mm) and contain microphenocrysts and elongate microlites of Opx, the latter with displaying feathery terminations. Cpx may rim Opx.

In the Bushveld, quench-textured pyroxenitic sills (CAWTHORN et al.,1981) bear some ressemblance to these cumulates. The groundmass contains ulvospinel, and plag and pyx which are partially included in Cpx, amph, plag and qtz.

Evidence for the existence in the Vourinos ophiolite of two separate magmatic events comes from field relationships, petrography (crystallization sequences), and bulkrock and mineral chemical data. Careful interpretation of the textural features of the cumulus rocks may lead to a better understanding of accumulation processes and the chemical evolution of the liquids and precipitates in the magma chamber.

The Krapa cumulates, volumetrically more important than the Asprokambo dykes and small intrusions, represent the relicts of a well-developed magma chamber. The lower part, rich in olivine cumulates, crystallized from successive magma inputs, as indicated by the presence of cyclic units (JACKSON et al., 1975). Large, unzoned minerals suggest the operation of diffusion processes (WAGER et al., 1960) and crystallization under equilibrium conditions. Crystallization may have occurred mainly near the floor and lower flanks of the temporary magma chamber, and in feeder dykes (JACKSON, 1961, OHNENSTETTER et al., 1975, CAMPBELL,

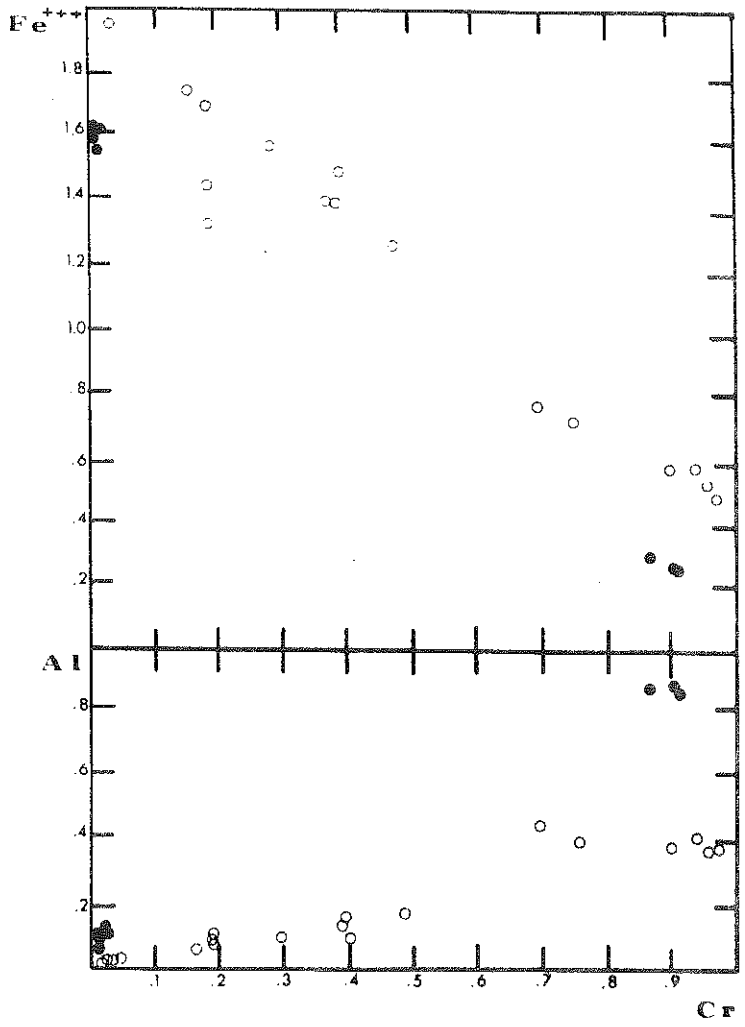
1978, IRVINE, 1980). The abrupt change in layering and the nature of cumulates between the lower and upper parts of the cumulus sequence testify to some important modification of the chemical and thermal evolution of the magma in the magma chamber (MC BIRNEY and HOYES, 1979). The gradual decrease in grainsize upward and the existence of phenomena such as graded bedding indicate current action in a progressively shrinking magma chamber in which no further major addition of liquid occurred.

The proximity of the top of the magma chamber may have induced higher cooling rates and lower temperatures, as suggested by the presence of low-temperature magmatic minerals. Thus, the upper group correspond to the virtually complete crystallization of a magma chamber in which convection cells became smaller and more short-lived with time, and in which fractionation occured close to surface (Raleigh) equilibrium conditions. The persistence of virtually identical cumulate textures from the base to the top of the Asprokambo sequence may testify to no major change in the process of accumulation during crystallization of magma pulse. Ortho to heteradcumulate textures, the presence of elongate and/or skeletal minerals and the entrappment of residual liquid of Qz-dioritic composition indiciate rapid in situ crystallization. Therefore each magma pulse seems to have crystallized rapidly in a small, closed magma chamber.

2 - Mineral chemistry

a) Spinels and olivine

Several series of spinels exist in both magma suite at Vourinos. Representative analyse (Table 4) are plotted in Figure 7, which shows major substitutional trends in the octahedral B site of spinels. Cr-spinel in the lower Krapa cumulates, and Ti Mt in the ferrogabbros are separated by a compositional gap. In wherlite 52-47, small disseminated (less than 0.1 mm) Cr-spinels mainly included in Cpx have a Cr/Cr+Al ratio of 0.5 slightly lower than in stratiform intrusion chromites (IRVINE, 1965). The Cr-Al spinels fall in the same compositional field as the spinels from the Marum ophiolite (JACQUES, 1981), and are more Cr-rich than most mid-ocean ridge basalt chromites. The $Fe^{3+}/Fe^{3+}+Cr+Al$ ratio of the Krapa spinels is around 0.2, a value typical of stratiform intrusion chromites. Spinel composition may depend of the onset of plagioclase crystallization (RIDLEY, 1977). Thus the low Al_2O_3 contents of late crystallizing Cr-spinels compared to early Krapa spinels may be explained by some plagioclase crystallization during this stage of differentiation. The occurence of interstitial to poikilitic TiMt in ferrogabbros resulted from the buildup of TiO_2 and FeO in residual Krapa liquids compositions.



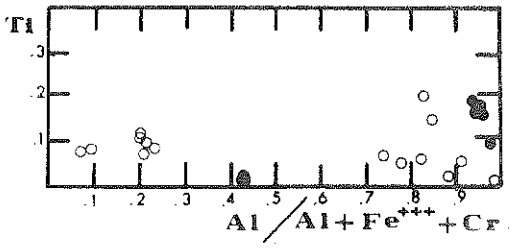


Fig. 7a Fe^{+++} vs Cr and Al vs Cr in the spinels of the cumulates from Krapa (solid circles) and Asprokambo (open circles).

Fig. 7b Ti vs Al/Al+Fe^{+++}+Cr. same symbols as in Fig. 7a

Chromian and aluminian titaniferous magnetite and titanomagnetite from the Asprokambo rocks display a wide compositional range : Ti (0.04-0.201), Cr (0-0.385), Al (0.008-0.139), Fe^{3+} (1.322-1.944). Ti spinels poor in Mt molecule are small subhedra ($<150\mu$) included in Opx, Cpx and Pl. Ti spinel richer in Mt component ($3<Cr_2O_3<16$; $1<Al_2O_3<4$) are larger, more euhedral and occasionally skeletal. They have crystallized later being in intersertal position towards the rim of igneous amphibole. Asprokambo Ti-spinels, always Mg-poor, have a lower $Mg/Mg+Fe^{++}$ (<0.1) than those from stratiform complexes (0.2-0.9) recorded by IRVINE (1967) and compositions do not correlate with the evolution of the $Mg/Mg+Fe^{++}$ ratio of coexisting silicates. Low ratios are known in evolved lavas (ARCULUS, 1978) where Fe^{++} increases strongly in late Ti-chromite (EL GORESY, 1976). Ferroan spinels from Asprokambo may also belong to class II of JOHAN (1979) characteristic of basic complexes having a metallogenic potentiality. Both Cr and Al have a sympathetic variation possibly partially influenced by Cpx and Pl fractionation, whereas Cr and Fe^{+++} and also Al and Fe^{+++} show antipathetic behaviour at a constant $Mg/Mg+Fe^{++}$ ratio (Fig. 7). Fe^{+++} increase (Fe trend) has been attributed to reaction between cumulus phases and trapped liquid (HENDERSON, 1975, MENZIES, 1975). Higher f_{O_2} possibly linked to an increase in H_2O concentration during differentiation is also suggested by igneous amphibole coexisting with Fe^{+++} poor plagioclase. Same amounts of Cr and Al are present in spinels from lunar rocks, Paricutain volcanics and the Bear Mountain igneous complex (EL GORESY, 1976, OSBORN et al., 1979, SNOKE et al., 1981). These features as well as the lack of Ti increase with increasing Fe^{+++} emphasize the calc-alkaline nature of the Asprokambo series. Like in chromian magnetite, ulvospinel evolving to aluminian chromite from pegmatites intruding the Critical zone of the Bushveld complex (CAMERON and GLOVER, 1973), substantial amounts of V_2O_3 (1-2.5 %) are present in Asprokambo spinels which may explain the regular V decrease in residual liquids. In ophiolites, spinels with such compositions are rare. For example, in the Marum ophiolite, JACQUES (1981) describes a Fe^{+++} rich spinel having low Mg/Mg+Fe ($\simeq 0.15$) from a bronzite cumulate. Similarly in Oman, spinels from a wherlite at high stratigraphic level are TiO_2 ($\simeq 3$ to 7 %)and Fe_2O_3 (30 %) rich (PALLISTER and HOPSON, 1981).

b) Pyroxenes

Pyroxenes are important cumulus phases in both sequences, crystallizing after olivine and spinel (Fig. 8). Opx crystallized at an early in Asprokambo magmas compared to the Krapa magmas in which it is common only in the upper cumulates (Tab4).

TABLE 4 Selected analyses of minerals from the intrusives from Krapa and Asprokambo series

KRAPA	52 - 47 Wr				AP 2247 Pl - Wr		AP 12307 n - G		AP 42 Mt - nG		
	OL	OPX	CPX	Cr-Sp	CPX	PL	CPX	PL	OPX	CPX	PL
	0320	0521	0116	1238	0034	0059	0385	0477	0326	116	222
SiO_2	40.14	56.14	52.71	0.00	53.06	45.19	52.82	44.47	52.03	51.36	44.24
TiO_2	.02	.12	.01	.42	.13	.00	.06	.00	.08	.40	.00
Al_2O_3	.00	1.63	2.02	22.57	1.72	34.66	.50	34.89	1.35	1.89	34.88
FeO	14.66	9.09	3.62	31.69	4.48	.37	7.66	.66	23.68	10.37	0.47
MnO	.21	.23	.00	.22	.20	.00	.44	.00	.36	.22	.00
MgO	44.95	31.44	17.26	9.11	15.57	.02	14.29	.01	20.92	13.50	.04
CaO	.05	1.06	23.75	0.00	23.58	18.75	23.28	19.39	1.26	20.37	18.82
Na_2O	.00	.00	.19	0.14	.23	.95	.05	.56	.02	.11	.61
K_2O	.02	.00	.00	.00	.00	.00	.00	.00	.00	.00	.04
Cr_2O_3	.00	.14	.59	34.37	.17	=	.00	=	.00	.00	=
NiO	.26	.00	.00	0.00	.00	=	.00	=	.00	.00	=
Total	100.31	99.85	100.15	98.52	99.14	99.94	99.10	99.98	99.71	98.23	99.11

ASPROKAMBO	AP 2714 n G					AP 71				Pl Wb	
	OPX	CPX	PL	AMPH	SP(1)	OPX	CPX	PL	PL	AMPH	M t
	0266	0574	0030	0720	1223	0123	0024	0217	0059	0319	0851
SiO_2	53.76	53.13	43.31	50.60	.00	52.99	52.13	45.26	59.27	49.04	.06
TiO_2	.00	.00	.00	.27	2.18	.10	.42	.00	.00	.27	36.33
Al_2O_3	1.18	1.71	35.03	4.29	3.67	1.08	1.35	34.45	25.71	7.38	.00
FeO	17.67	3.61	.62	11.12	76.26	19.34	8.73	.49	.33	12.57	59.10
MnO	.21	.30	.00	.33	.28	.33	.26	.09	.10	.31	.52
MgO	25.48	16.55	.00	17.23	.12	23.83	14.98	.07	.00	14.84	.58
CaO	1.25	23.88	19.27	10.91	.00	1.51	21.67	19.06	7.76	11.54	.00
Na_2O	.00	.11	.46	.63	.00	.06	.13	.93	6.78	.64	.16
K_2O	.00	.00	.00	.04	.00	.03	.00	.00	.32	.00	.00
Cr_2O_3	.25	.50	.00	.12	12.17	.05	.00	=	=	.00	.31
NiO	.10	.00	.00	.00	.00	.02	.00	=	=	.00	.00
Total	99.90	99.79	98.69	95.53	93.67	99.34	99.67	100.35	100.27	96.59	97.05

OL = olivine; OPX = orthopyroxene; CPX = clinopyroxene; Cr-Sp= chrome -spinel; Pl = plagioclase; AMPH = amphibole; Mt = Magnetite Other symbols same as in table. (1) Included 1.03 % V_2O_5

α) Orthopyroxenes

Opx from both series are poor in Al, especially in Al^{IV} (0.015-0.050) ; such compositions are typical of Opx precipitating from a Qz-tholeiitic magma at low P. With structural height or magmatic differentiation, Mn contents increase ; Ti is always low and constant. The Cr contents or early cumulate Opx are low (0.009-0.014) and decrease regularly range from 1.3 to 3.3 % Wo (Fig.6) outside the field of "solidus" Opx : > 4 % (HUEBNER, 1980). Their compositions may indicate contimination by a calcic phase in fine exsolution lamellae (HUEBNER, 1980) or incomplete reequilibration (JACQUES, 1981). Ca is generally constant (0.06) for slight Si variation. A regular Mg decrease is observed from ultramafites to ferrogabbros for M1 and M2 sites (Fig. 9) linked to upwards $act^{Opx}_{Mg_2Si_2O_6}$ decrease in the stratigraphic pile. Rapidely cooled cumulates from Asprokambo exhibit large and antipathetic variation of $act^{Opx}_{Mg_2Si_2O_6}$ and $Fe^{++}/Fe^{++}+Mg$ in single rocks (Fig.10)

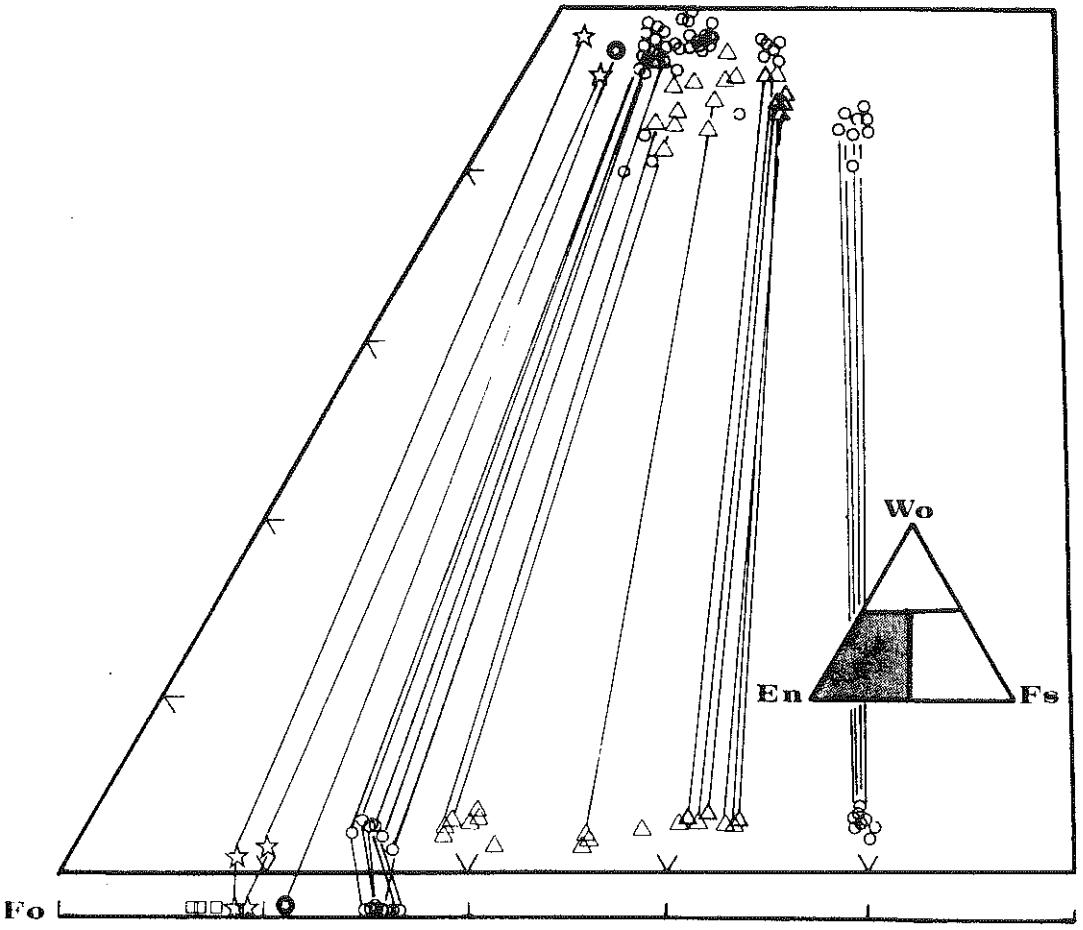


Fig. 8 Coexisting Ol, Opx and Cpx in the plutonic rocks from Krapa (circles) and Asprokambo (triangles) compared to dunites (squares) harzburgites (stars) and ultramafic cumulate (star enclosed in circle) from the mantle sequence (HARKINS et al., 1980).

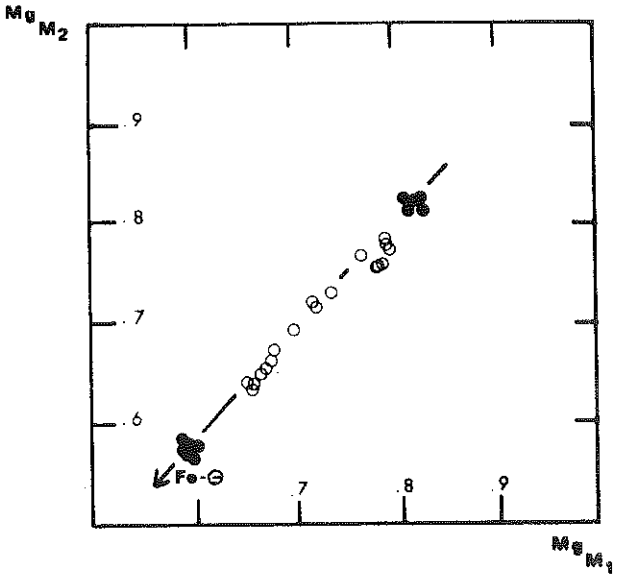


Fig. 9 Mg concentration in M1 and M2 sites from the orthopyroxene. Solid circle = Krapa; open circle= Asprokambo. The trend indicates the magmatic differentiation.

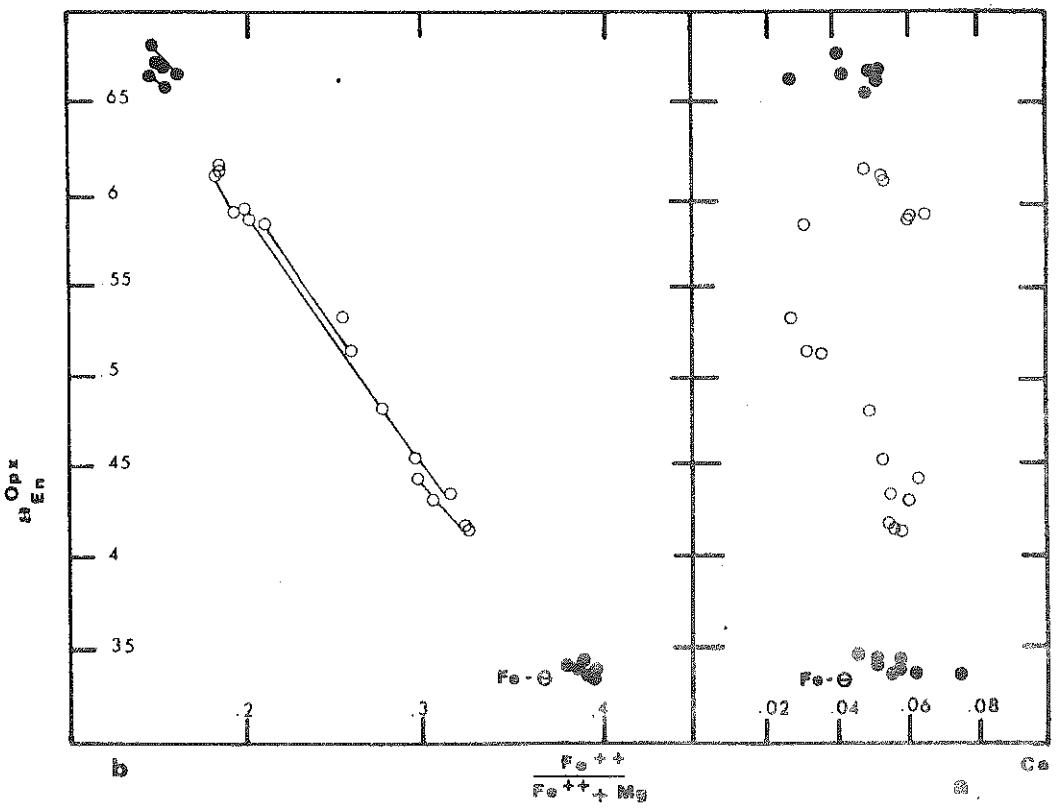


Fig. 10a $Act^{Opx}_{Mg_2Si_2O_6}$ vs Ca content in Opx. Same symbols as in Fig. 9

Fig. 10b $Act^{Opx}_{Mg_2Si_2O_6}$ vs $Fe^{++}/Fe^{++}+Mg$ in Opx. Same symbols as in Fig. 9. The lines indicate evolution in some single rocks. Note the large variation in rapidly cooled rocks.

β) Clinopyroxenes

As for Opx, Cpx of both series have Ti (0.10) and Al nearly constant during differentiation whereas Mn increases and Cr decreases (0.016-0)(fig.11). At low pressures, the low Al contents of these pyroxenes indicate that they have crystallized from a Si-rich liquid (CAWTHORN and O'HARA, 1976). Variation in element concentration, especially Cr, with changing $Fe^{++}/Fe^{++}+Mg$ ratio may testify to addition of fresh magmas pulses during accumulation of the lower cumulates. Al is mainly restricted to octahedral sites, Si occupying most of the tetrahedral sites. The Ca-Si relationsphip is more pronounced than in Opx, with Ca varying from 0.82 to 0.69 with only limited Si variation (1.92-1.96). It may also be noted that Krapa cumulate clinopyroxene are slightly richer in Ca and Al and poorer in Cr and Ti than in Asprokambo clinopyroxenes (Fig. 8 and 11).

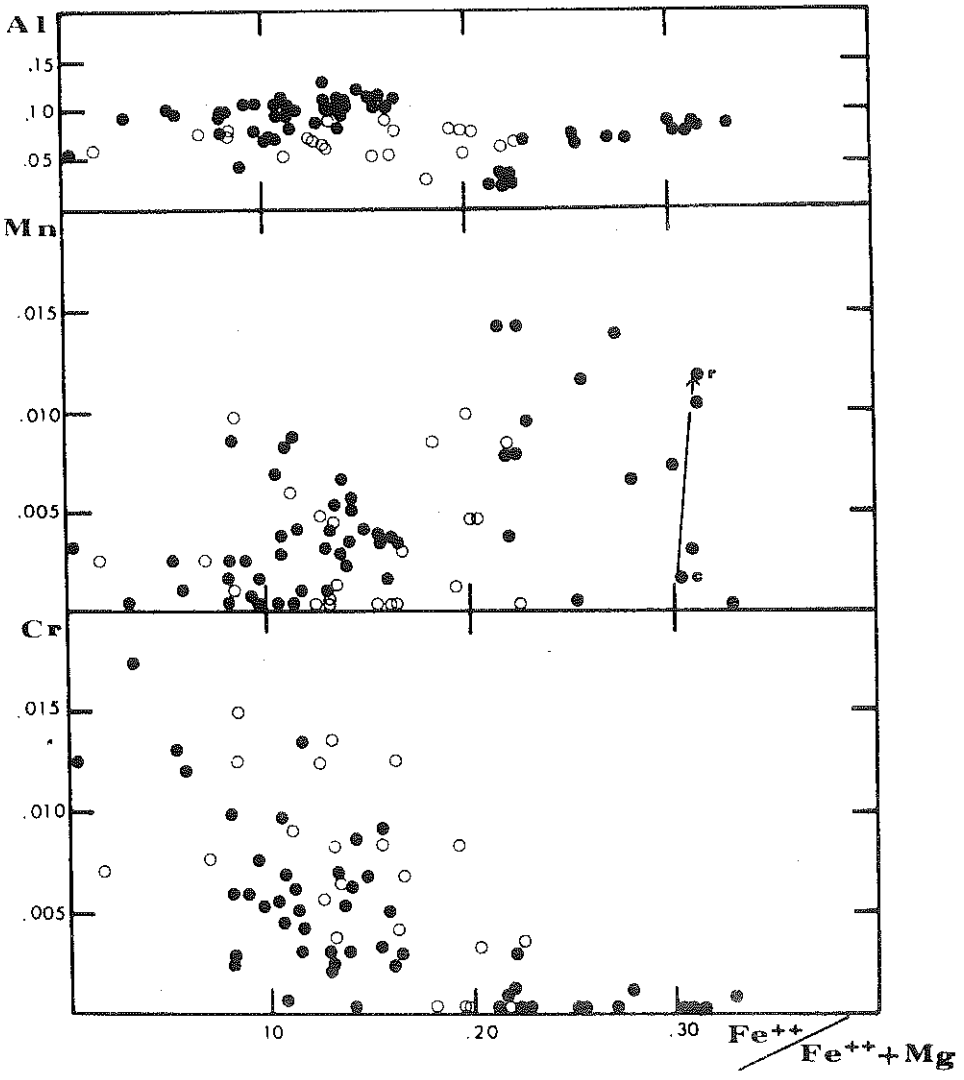


Fig. 11 Mn, Al, Cr evolution against $Fe^{++}/Fe^{++}+Mg$ in clinopyroxenes. Same symbols as in Fig 9.

Most of Cpx from both series lie in the field between solidus and subsolidus Cpx (HUEBNER, 1980) due to incomplete reequilibration, as do pyroxenes from the Harum ophiolite (JACQUES, 1981). En-rich Cpx with 40 % Wo as well as short tie-lines in ferrogabbros may represent magmatic relicts. In fig. 12A, the $act^{Cpx}_{Mg_2Si_2O_6}$ decreases regularly with increasing calcium content, due to subsolidus reequilibration. Ferrogabbros and a sample from Asprokambo deviate from this trend. No obvious correlation is apparent between this activity and $Fe^{++}/Fe^{++}+Mg$ (fig. 12b). In single rocks, especially from the lower cumulates steep trends indicate a slight increase of Mg with decreasing activities, that is during subsolidus recrystallization. From ultramafics to ferrogabbros, a line with a smooth slope passing at high activity values may represent the solidus trend. Consequently, trend in ferrogabbros may indicate preserved magmatic features.

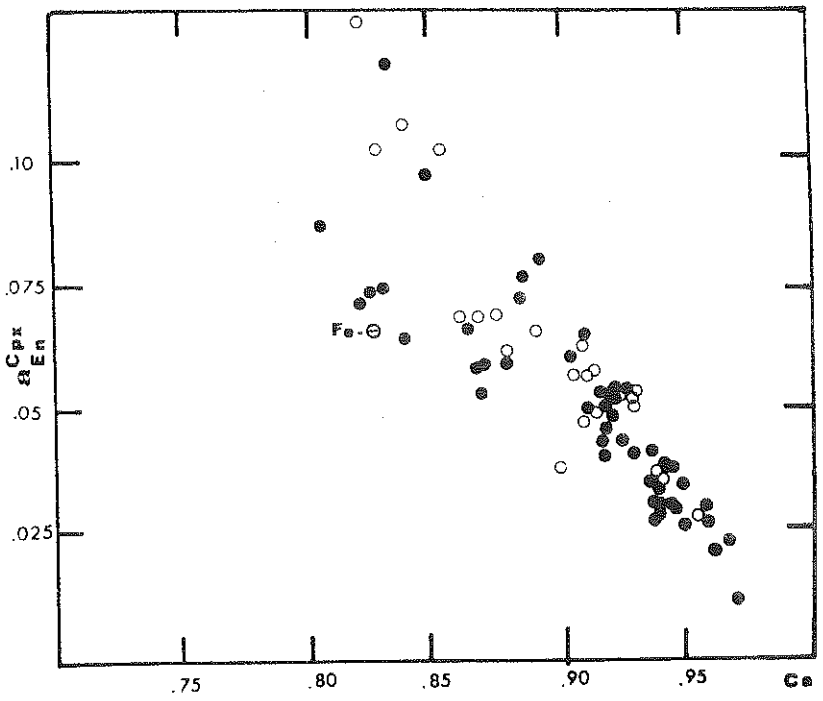


Fig. 12a $Act^{Cpx}_{Mg_2Si_2O_6}$ against Ca content in clinopyroxenes. Same symbols as in Fig. 9

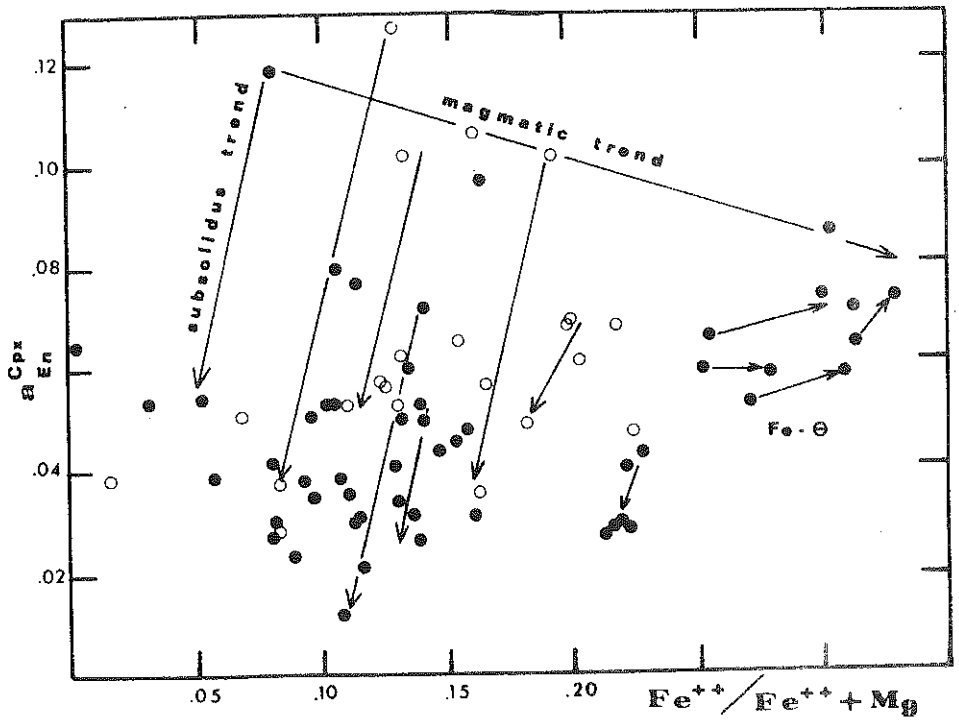


Fig. 12b $Act^{Cpx}_{Mg_2Si_2O_6}$ against $Fe^{++}/Fe^{++}+Mg$ in Cpx. Same symbols as in Fig. 9

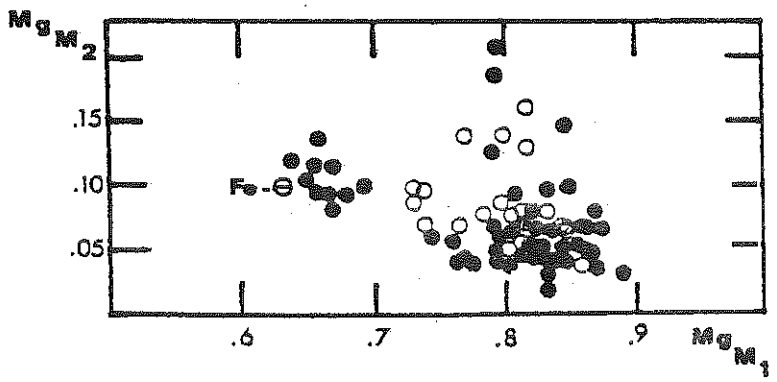


Fig. 12c Mg concentrations in M1 and M2 sites from Cpx. Same symbols as in Fig. 9. The das ed line indicate possible magmatic differentiation.

γ) Thermometry

Mg partitioning between M1 and M2 sites for both pyroxenes reveals that the best correlation is exhibited by the M1 sites (fig. 13). The sympathetic igneous correlation between $M2^{Opx}$ and $M2^{Cpx}$ has been destroyed by variable decrease of Mg^{Cpx}_{M2}, resulting from Ca increase during subsolidus reequilibration. The WOOD and BANNO (1973) geothermometer has been used to estimate temperature of equilibration of the Vourinos samples. As expected calculated temperatures decrease as $Fe^{++}/Fe^{++}+Mg$ of the pyroxene increase. In fig. 14 a line passing through the higher temperature points of the cumulates broadly defines the tem-

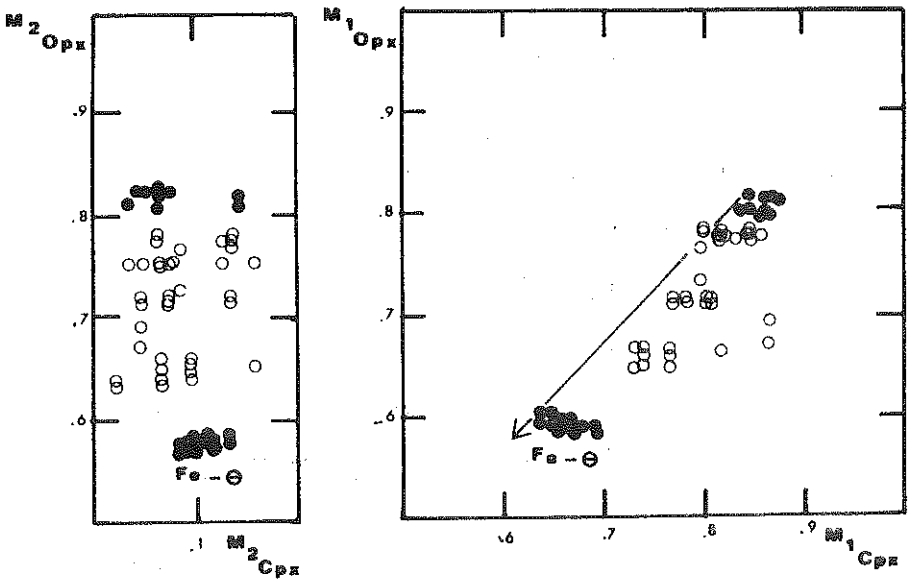


Fig. 13 Repartition of Mg in M1 site (Fig.13a) and M2 site (FIg.13b) from coexisting pyroxenes.

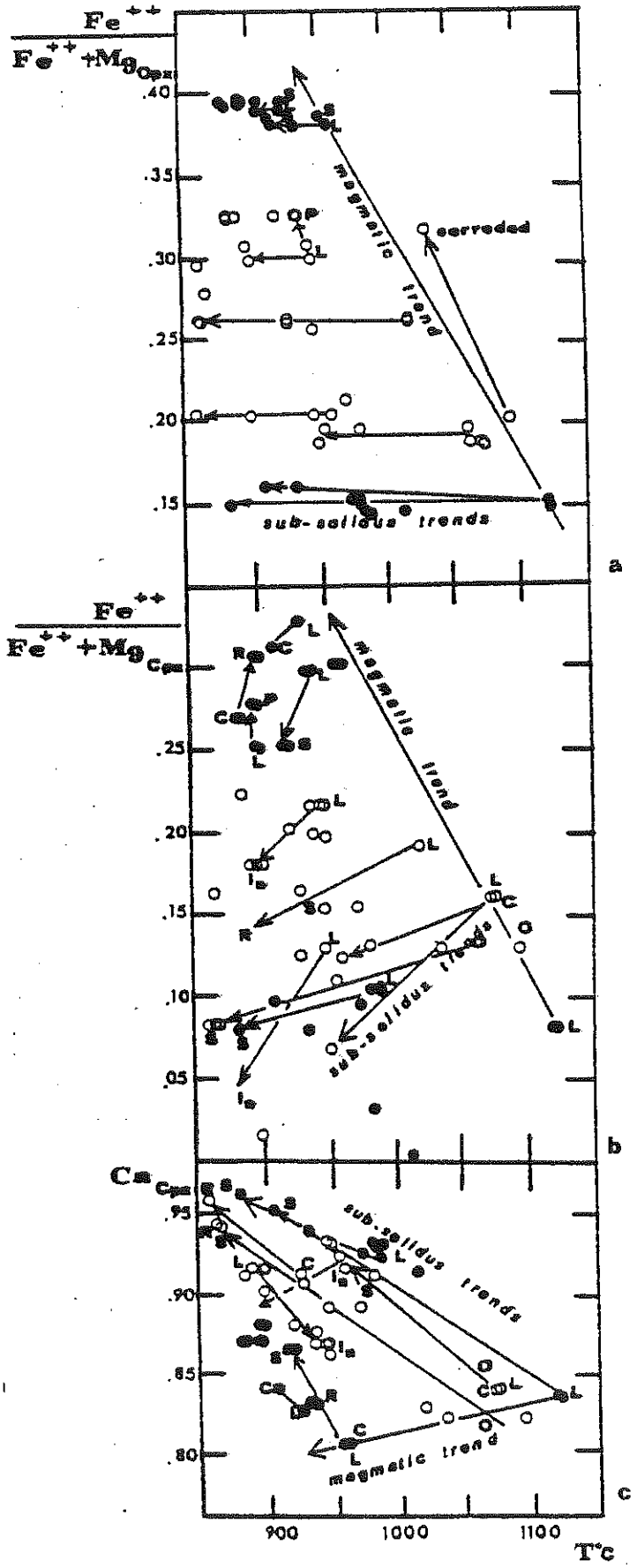


Fig.14 $Fe^{++}/Fe^{++}+Mg$ in Opx (Fig.14a), and Cpx (Fig.14b) and Ca in Cpx (Fig. 14c) versus temperatures calculated with the WOOD and BANNO (1973) method in coexisting pyroxenes. Same symbols as in Fig. 9. C= core, R= rim, L=large, S= small, P= poikilitic, I_A= pyroxene included in amphiboles, O= oikocryst

perature variation during magmatic differentiation (1125-925ºC). Note also that the temperature evaluated by Fe/Mg partitioning (ROEDER and EMSLIE, 1970) between olivine from the lower cumulate and the least fractionated basalt of Krapa is about 1225ºC. In single rocks, the magmatic trend is found only in some rapidly-crystallized Asprokambo cumulates and Krapa ferrogabbros. Large temperature range has been found for every sample at constant $Fe^{++}/Fe^{++}+Mg$. The temperature difference determined for any single sample decreases from the base to the top (a 200ºC interval in ultramafic cumulates and a 50ºC or less (20ºC) in ferrogabbros). As noted above (fig. 14), a line broadly corresponding to the temperature decrease with differentiation may be drawn. In most of the samples, the temperature drop is accompanied by a Mg increase. In single rocks from Asprokambo the highest Mg concentration is found in Cpx included in amphiboles. In the Ca vs temperature diagram (fig. 14C) a line passing through the most Ca-poor Cpx of the various cumulates decreases smoothly during the magmatic differentiation. In single rocks from the lower cumulates, the large temperature interval (210ºC) is linked to core to rim Ca variation and/or form large cumulus Cpx to small interstitial Cpx. In the Asprokambo cumulates, corroded Cpx surrounded by igneous Am have variable Ca content (Ca decreasing). Consequently, from the base to the top of the stratigraphic cumulate pile, the temperature decreases as $Fe^{++}/Fe^{++}+$ Mg increases, with a slight Ca decrease. Most calculated temperatures are due to subsolidus reequilibration, leading to strong Ca and moderate Mg increases in the pyroxenes. Cores of some large Cpx preserve high (magmatic) temperature cation distributions, while smaller and intercumulus pyroxenes give lower temperature.

Fig. 14a In the Asprokambo sample, temperatures decreases from euhedral to corroded cumulus Opx (50°C) or from large to poikilitic interstitial Opx (15°C). In ferrogabbros the more reduced temperature interval (10°C) is linked to variation from large to small cumulus Opx.

Fig. 14b In each sample note the Mg increase with temperature drop. The strongest slope is got from Asprokambo sample having igneous amphibole. Ferrogabbro sample represents slight $Fe^{++}/Fe^{++}+Mg$ increase coupled with a Ca decrease during magmatic differentiation of the Krapa bodies.

Fig. 14c In the least evolved cumulates large temperature interval (210°C) shown from large cumulus Cpx and/or oikocrysts to rim or small interstitial Cpx. A minor Ca decrease from core to rim is recorded in evolved cumulates, otherwise having higher Ca concentration.

Rapidly-cooled ferrogabbros which crystallized at the closing stages of magma in the chamber in an upper level of the magma chamber may have escaped extensive subsolidus reequilibration. For similar reasons and because of the small size of the Asprokambo magmatic chambers or sills, chemical variations linked to igneous processes may also have been preserved.

c) Amphiboles

In ophiolites, especially in upper cumulates both magmatic and metamorphic amphiboles may coexist. Textural relationship indicate that Am from Krapa are mainly secondary products after Cpx whereas those from Asprokambo may be both primary and secondary. Using the nomenclature of LEAKE (1978), amphibole evolve from Mg-hornblende to actinolitic hornblende. In the Krapa lower cumulates Mg-hornblende occur in blebs and patches replacing Cpx or in interstitial

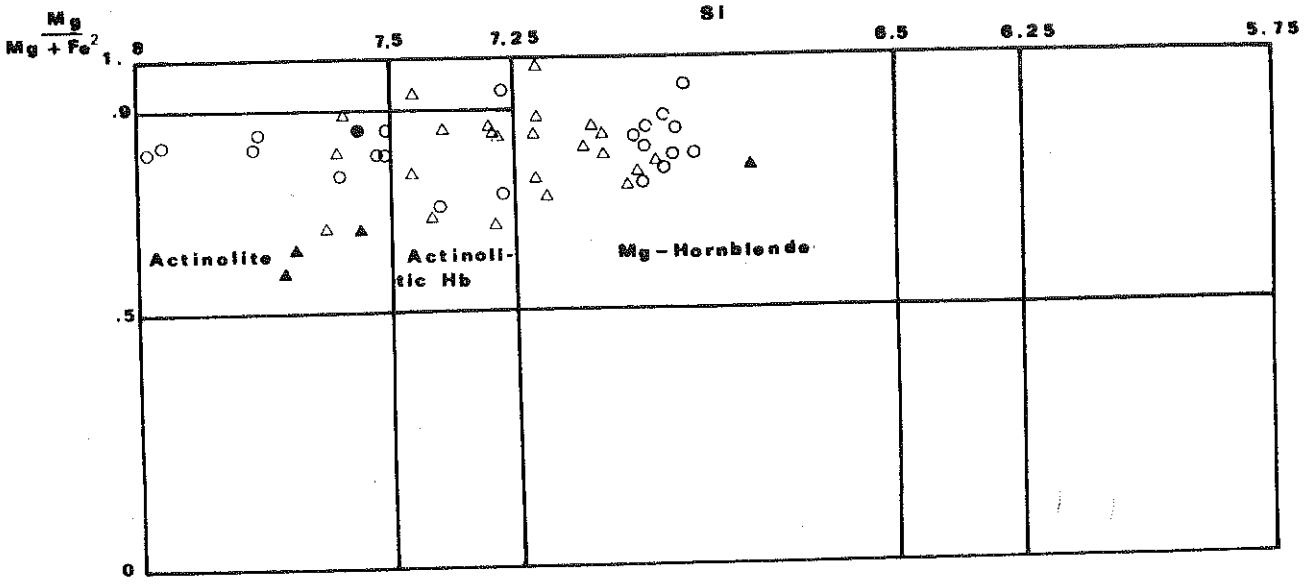


Fig. 15 Amphiboles from the Vourinos ophiolite. Solid circle Krapa effusive and hypabyssal rocks, open circle= Krapa cumulates; triangle= Asprokambo effusives, open triangle= Asprokambo cumulates.

areas betwwen them. Actinolite and actinolitic hornblende are mainly developed after Opx in close association with chlorite, and form needles sub-parallel to cleavages or criss-cross patterns. In the Asprokambo sequence most of the analyzed hornblendes may be of magmatic origin. These form postcumulus poikilitic blades, often subhedral having a brownish to greenish color. Their size and abundance increase with stratigraphic height from = 200μ in medium grained pyroxenites up to 3 mm in Am-bearing gabbronorites and Qz-diorites. Their scar-

city in granophyres may be related to the low CaO or the magma (2 %, CAWTHORN, 1967a). From projections in basalt tetrahedron, the association Pl-Opx-Am-Qz cannot coexist (CAWTHORN, 1976a). The observed reaction : corroded Opx in a Qz Am matrix may correspond to successive assemblages : Opx-Pl → Am-Qz. The local corrosion of both pyroxenes may result from reaction with liquid when f_{H_2O} increases (f_{H_2O} = 0.6 P total) towards the end of differentiation (CAWTHORN, 1976b). This may explain the disappearance of Opx and the local scarcity of Cpx in evolved cumulates, especially in Qz-diorite. TiO_2 (0.17-1.42) ; Na_2O (0.41-1.35) and Al_2O_3 (E.29-7.38) increase slightly with $Mg/Mg+Fe^{++}$ (0.65-0.95). Metamorphic Mg-hornblendes from Krapa are in the range of the Asprokambo amphibole compositions, being locally richer in TiO_2 (0.38-1.42) ; Na_2O (1.24-1.60) and Al_2O_3 (6.42-8.59).

In Asprokambo samples, Mg-hornblende may grade to actinolites varieties towards their border. Fibrous actinolite or actinolitic-hornblende in most of the cumulates and tremolitic amphiboles in the lower cumulates are clearly of metamorphic origin, replacing Mg-hornblende and pyroxenes. Ferro-magnesian amphibole cummingtonite and Mg-cummingtonite) are secondary products after Opx especially in evolved cumulates. Late crystallization of amphibole has been recorded from some ophiolitic cumulates (HERZ, 1951, BAILEY and BLAKE, 1974, PALLISTER and HOPSON, 1981) together with metamorphic greenschist facies varieties. Pargasite and tschermakite have also been recorded as primary amphiboles (HUTCHISON, 1978).

d) Plagioclase

The high An content of cumulus plagioclase from most ophiolites is well known (PARROT and RICOU, 1976, COLEMAN, 1977, JUTEAU and WHITECHURCH, 1980, JACQUES, 1981, HOPSON and PALLISTER, 1981). Generally high An contents are usual from plagioclases in calc-alkaline rocks (CAWTHORN, 1976c), for example in cumulates ($An_{90.5-95.8}$) from Granada, Lesser Antilles (ARCULUS, 1978) and from Alaskan-type complexes (SNOKE et al., 1981). In plagioclase from Vourinos cumulates, An is nearly constant and high (90.5-95.8). No important An decrease may be correlated with the stratigraphic height as is shown for instance by the Marum ophiolitic cumulates (JACQUES, 1981). Generally, in the Vourinos cumulates, only a slight variation in An occurs in each rock at a constant $Fe^{++}/Fe^{++}+Mg$. The behaviour of Fe^{+++} versus An content in both series is different. In the Krapa sequence Fe^{+++} in Pl tends to increase with bulk-rock Fe^{+++}. In Pl from

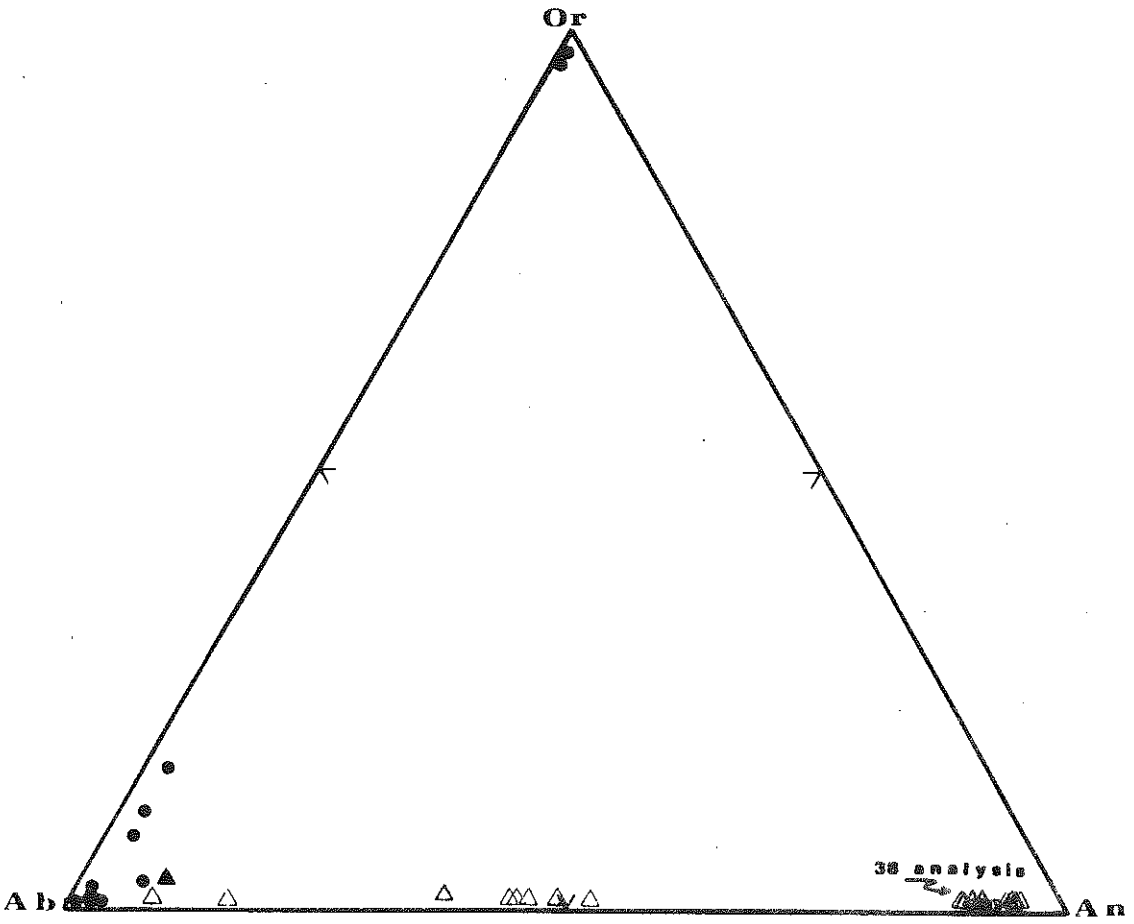


Fig. 16 Plagioclase composition of the Vourinos ophiolite in the Ab-An-Or triangle. Same symbols as in Fig. 15

the lower cumulates (e.g. in Pl - clinopyroxenite). An decreases slightly (95.8-91.6) and Fe^{+++} (0.11-0.029) increases from coarse and partly included crystals in Cpx to interstitial plagioclase. Similar core to rim trends ($An_{95.6-93.9}$) and Fe^{+++} (0.043-0.061) emphasize the positive correlation between Fe^{+++} and Na. In contrast, for the upper cumulates, especially in ferrogabbros, a Fe^{+++} decrease is coupled with a slight Na increase from cumulus to interstitial smaller Pl or from core to rim. The opposite behaviour of Fe^{+++} and Na in lower cumulates and ferrogabbros may respectively reflect the early iron increase at the begenning of the magmatic evolution later followed by a Fe^{+++} decreases (0.038-0). Abrupt sodic rims at the margins of highly calcic Pl exist in effusive rocks from Tonga (EWART, 1976). In cross-cutting micropyroxenite, xenocryts show a peculiar reaction with a coupled Ca and Fe^{+++} increase from core to rim, possibly linked to partial reequilibration with the magma. Homogeneous and rapidly-crystallized micropyroxenites show $Fe^{++}/Fe^{++}+Mg$ increasing in Opx at constant An in poikilitic Pl. Therefore in Krapa and Asprokambo plagioclases, An and Fe^{+++} compositions seem to result mainly from magmatic processes.

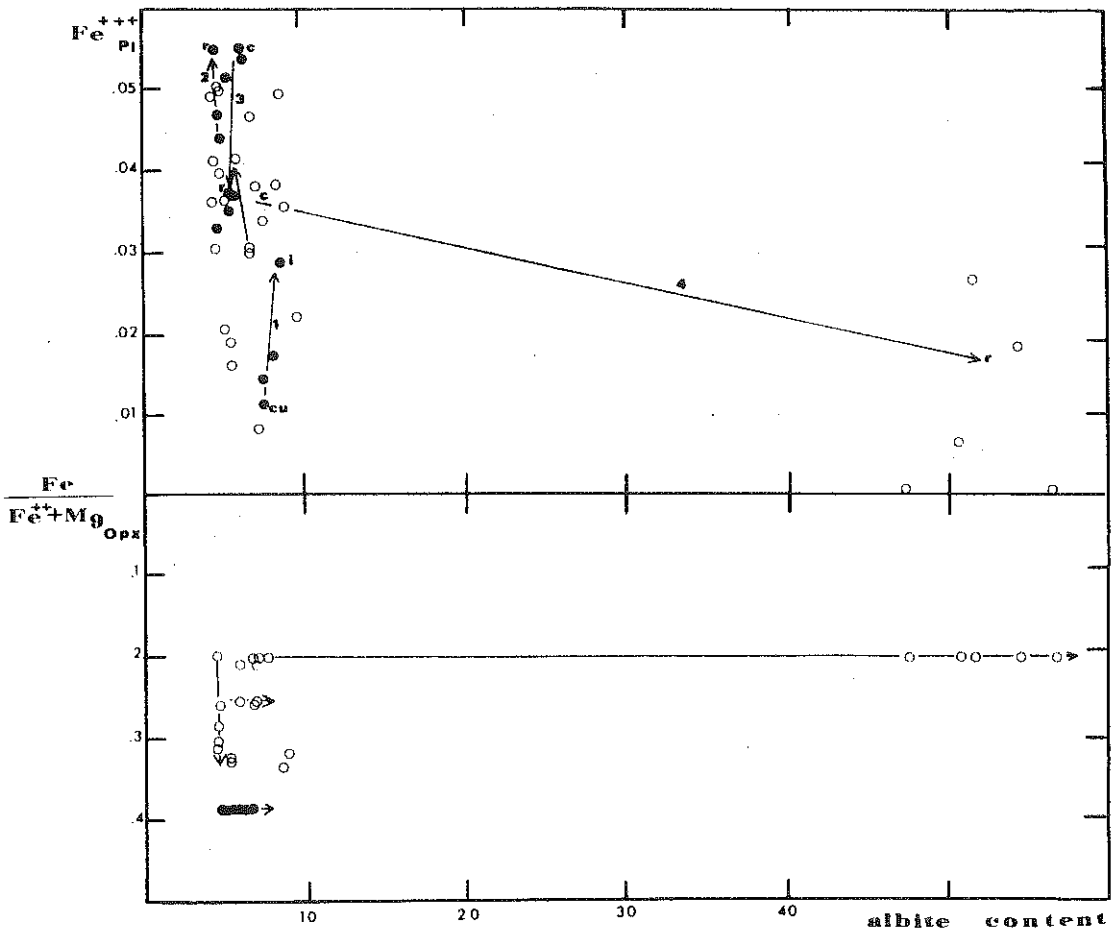


Fig. 17A Correlation between Fe^{+++} and Ab content in feldspar. For clarity only some trends have been drawn 1) Pl clinopyroxenite, 2) melagabbro, 3) ferrogabbro, 4) rapidly cooled pl pyroxenite. Same symbols as in Fig. 9 c=core, r= rim, cu= cumulus, i= intercumulus

Fig. 17B Ab content vs $Fe^{++}/Fe^{++}+Mg$ ratio in Opx. same symbols as in Fig.9. Lines drawn from single rock indicate the major evolution from core to rim or from cumulus to intercumulus minerals

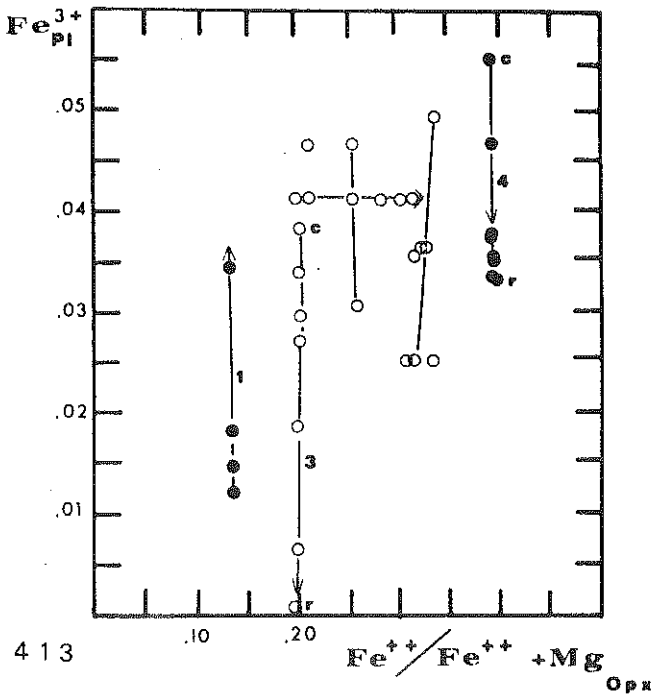


Fig. 17c - Fe^{+++} in Pl vs Vs $Fe^{++}/Fe^{++}+Mg$ ratio in coexisting Opx. Same symbols and trends as in fig. 17a.

3 - Whole rock chemistry

The chemistry of the intrusive rocks is strongly controlled by presence and relative abundance of the various cumulus phases. Thus, compositions of the plutonic rocks do not correspond to liquid compositions, but represent mixtures of the compositions of cumulus phases plus trapped interstitial liquid. Although cyclic layering occurs (Krapa), some broad geochemical trends have been noted.

In the Krapa cumulate sequence, FeO^*, TiO_2 and V increase from the base to the top, while Mg, Ni and Cr all decrease. These chemical variations correspond to the transition from clinopyroxenites and wehrlites through noritic gabbros to Mt-rich noritic gabbros. Al reaches a maximum concentration in the upper part of the noritic gabbros.

The constant very low concentrations of Zr and P, which are "incompatible" in the major cumulus phases reflect the low contents of trapped intercumulus liquid (except in the uppermost part of the sequence) in these rocks, in concert with their adcumulate textures.

The chemical features of the Asprokambo intrusives reflect different cumulus processes and crystallization histories which occured during formation of these rocks in a shallow chamber. Fe, V, Mg, Cr, Co and Ni decrease upward through the basal part of sequence and Al and Si increase, according to the progressively increasing abundance of cumulus plagioclase accompagnying the transition from pyroxenites to melanoritic gabbros. Once again, the low-Zr and P contents in these basal rocks indicate very low contents of trapped interstitiel liquid, reflecting the efficency of the heteradcumulus crystallization processes. In the middle and upper parts of the Asprokambo sequence, rock compositions approach those of the extrusive and hypabyssal rocks (of similar Fe^*O/MgO ratios), since orthocumulus and isotropic crystallization are dominant during these later stages of fractionation.

4) Fractionation of the two magmatic series in separate low pressure magma chambers

In the following section we discuss the distinctive features of the two fractionation systems. Variations diagrams of Zr versus other elements (fig. 18) permit definition of several stages of fractionation related to the separation of successive different cumulate assemblages. Open versus closes system and perfect equilibrium versus Rayleigh fractionation models have been used to test possible liquid lines of descent by consideration of Zr vs FeO^*/MgO ratio (fig.19). F represents the portion of liquid remaining in the system after separation of 1-F solids from the parental liquid.

Table 5 - Chemical analyses of intrusive rocks from the Krapa series

	Wr	Cp	Cp	pl-Wr	ol-nG	n-G	ln-G	ln-G	ln-G	n-G	mt-nG	n-G	n-G	mt-nG	mt-nG
	5-24-7	4-24-7	4-24-8	2-24-7	1-24-7	12-30-7	4-4-8	1-4-8	2-4-8	AP 4-6	AP 1-6	15-27-7	AP-12-6	AP34 11	AP 42
SiO_2	47.9	50.1	51.54	49.45	48.96	46.35	48.11	47.08	46.57	49.12	45.70	41.81	48.25	46.81	45.09
TiO_2	0.17	0.18	0.07	0.21	0.13	0.09	0.05	0.09	0.09	0.12	0.19	0.18	0.44	0.4	0.61
Al_2O_3	2.41	2.74	1.96	8.60	10.95	17.38	18.21	19.66	20.13	17.85	17.07	20.79	15.43	16.04	15.35
Fe_2O_3T.	8.20	6.05	5.30	4.62	5.84	7.88	7.26	6.78	6.14	7.04	11.15	8.53	8.11	12.63	17.58
MnO	0.16	0.16	0.10	0.10	0.11	0.10	0.11	0.09	0.10	=	=	0.15	=	0.20	0.21
MgO	24.20	19.71	18.35	14.20	14.00	9.56	8.57	7.85	7.24	9.58	8.25	6.87	6.15	8.34	7.09
CaO	12.63	17.62	20.51	19.94	16.46	13.62	13.64	14.05	14.78	12.00	12.61	14.74	10.19	13.66	12.69
Na_2O	0.13	0.20	=	0.23	0.94	1.21	0.90	1.09	1.25	0.91	1.23	0.90	3.30	0.53	0.68
K_2O	0.02	0.02	=	=	0.02	0.01	=	0.05	0.04	0.07	0.11	0.16	0.46	0.03	0.04
P_2O_5	<0.01	<0.01	<0.01		<0.01	<0.01	<0.01			<0.01	0.01	0.01	0.03		
PF	3.69	1.28	1.88	1.81	2.66	3.55	2.65	3.20	3.52	2.48	3.09	5.12	7.90	1.64	2.37
Tot.	99.51	98.08	99.71	99.16	100.08	99.76	99.51	99.94	99.86	99.17	99.41	99.27	100.26	100.06	99.71
Ni	476	277	212		179	82	89			54	76	128	22	51	43
Co	125	104	47		44	37	53			52	77	71			
Cr	1699	2174	2483		793	199	439			94	51	120		23	17
Cu	<10	< 10	10		8	107	10			137	261	48			
Zn	62	44	30		31	65	67			67	64	60			
V	136	201	175		176	201	110			195	302	398	264	260	625
Sr	24	30	4		76	98	111			121	173	197	92		
Li			8		2.5	4.8	1.8			2.1	=				
Rb	6	5	< 5		6	< 5	5			5	< 5	7	7	7	6
Nb	<2	< 2	< 2		< 2	< 2	< 2			< 2	< 2	< 2	2	2	2
Zr	9	11	11		11	10	12			10	13	12	22	19	21
Y	8	7	5		7	5	6			5	5	5	14	10	12
Zr/Y	1.1	1.5	2.2		2.2	2.0	2.0			2.0	2.2	2.4	1.6	1.9	1.75
Ti/Zr	113	98	38		71	54	25				88	90	120	126	174
Al_2O_3/TiO_2	14	15	28	41	84	193	364	218	224	149	90	115	35	89	25
FeO^*/MgO	0.3	0.3	0.3	0.3 -	0.4	0.7	0.8	0.8	0.8	0.7	1.2	1.1	1.2	1.36	2.37

Wr = Wherlites, Cp= Clinopyroxenites, pl-Wr= Plagioclase wherlites, ol-n G =Olivine noritic gabbros, ln-G = leuconoritic gabbros, n-G= Noritic gabbros, mt-nG = Magnetite - rich noritic gabbros

Table 6 - Chemical analyses of intrusive rocks from the Asprokambo series

	pl-W AP-7-1	pl-W AP-7-1	n-G 135	n-G 131A	q-inG AP-11-1	n-G AP-2-3	q-inG AP-3-3	q-D AP-8-3	q-D 108	q-D 136	Gr 132	Gr 2310	Gr 2312	Gr 2311
SiO_2	50.15	49.60	52.43	51.98	52.86	53.58	53.50	62.68	68.82	65.39	76.09	73.23	73.84	75.29
TiO_2	0.27	0.23	0.17	0.10	0.16	0.24	0.32	0.53	0.29	0.29	0.40	0.26	0.23	0.19
Al_2O_3	7.50	8.90	9.88	10.72	16.37	10.87	15.95	13.47	12.52	17.54	10.94	11.58	11.75	11.54
$Fe_2O_2T.$	11.88	11.41	8.96	10.72	6.07	9.32	10.21	7.75	4.70	1.71	2.66	5.25	4.60	4.17
MnO	nd	nd	0.14	0.15	nd	nd	nd	nd	0.12	0.10	0.08	0.10	0.09	0.11
MgO	13.90	12.71	12.31	11.86	8.10	7.28	5.45	3.50	1.56	1.07	0.94	0.98	0.97	0.59
CaO	14.30	14.74	11.78	12.74	9.62	14.38	8.17	8.14	6.80	7.16	3.70	1.86	1.47	0.68
Na_2O	0.46	0.45	1.36	0.45	2.58	=	3.33	=	2.43	3.12	2.89	3.53	3.37	4.58
K_2O	0.13	0.06	0.31	0.03	0.56	0.05	0.34	0.08	0.14	0.71	0.24	0.38	0.61	0.36
P_2O_5	0.02	0.03	=	=	=	0.03	0.02	=	0.05	=	0.06	=	0.05	0.05
PF	1.40	1.69	2.32	1.19	2.93	3.01	2.82	3.15	2.20	2.08	1.45	1.95	2.15	1.83
Tot.	100.01	99.83	99.66	100.04	99.25	98.76	100.11	99.30	99.63	99.17	99.45	99.12	99.12	99.39
Ni	95	106				61	51		29		23		19	17
Co							33		26		14		n.d.	n.d.
Cr	315	575				510	97		283		201		148	169
Cu							71		15		54		5	9
Zn							78		43		121		62	65
V	270	271				256	245		139		80		n.d.	n.d.
Sr	21	39				99	70		74		87		59	45
Li							3.1		3.2		1.6		5.5	4.0
Rb	< 5	< 5			< 5	< 5	7		4		4		16	7
Mo	< 2	< 2			< 2	< 2	< 2		4		4		4	4
Zr	9	9			8	16	20		24		28		44	47
Y	7	6			5	8	10		12		14		28	32
Zr/Y	1.3	1.5			1.6	2.0	2.0		2.0		2.0		1.6	1.5
Ti/Zr	180	153			120	90	96		72		86		31	24
Al_2O_3/TiO_2	28	39	58	107	102	45	50	25	43	66	27	41	51	61
FeO^*/MgO	0.8	0.8	0.7	0.8	0.7	1.2	1.7	2.0	1.4	1.4	2.6	4.8	4.3	6.4

pl-W= Plagioclase websterites, n-G=noritic gabbros, q-in G=Quartz ieuconoritic gabbros, q-D=Quarz diorites, Gr = Granophyres

a) Krapa

In Fig. 18a, two basalts (AP 26-7), one andesite (AP 9-6) and one dacite have been selected as representative of the liquid evolution of the series. The least differentiated basalt (AP 26-7) has an $Mg/Mg+Fe^{++}$ ratio of 0.64 and is in equilibrium (K_D 0.32, ROEDER and EMSLIE, 1970, BENDER et al., 1978) with olivine Fo_{85} as occurs in the lower cumulates. Therefore, this basalt does not represent a primary magma. The composition of hypothetical primary magmas in equilibrium with mantle olivine (Fo_{90-92}, as in Vourinos harzburgites, HARKINS et al., 1979) may be approximately determined by addition of variable amounts of progressively more magnesian olivine. For example, addition of about 4 % of each of Fo_{92} ; Fo_{89} ; Fo_{86} to basalt AP 26-7 would produce a liquid with an $Mg/Mg+Fe^{++}$ ratio of 0.72 which may be considered a reasonable value for melts in equilibrium with mantle olivine (GREEN, 1976). Thus in the following discussion, the parental primary magma of the Krapa series is calculated to have approximately 12.5 % MgO; 8.5 % FeO^* and 24 ppm Zr. Low pressure fractionation, F = 0.9 - 0.7 from basalts to basaltic andesites appears to have been mainly controlled by Ol and Cpx (and subordinate Pl) which produced wehrlite, clinopyroxenite and melagabbros. During the third stage of differentiation (F = 0.7-0.4) in which liquids evolved through basaltic andesite aoompositions , corresponding cumulates were the gabbronorites. Fe-Ti oxides began precipitating when liquids broadly andesitic in compositions, and corresponding cumulates include Mt-rich gabbro-norites ; Mt- and Qz bearing leucogabbronorites and diorites. The most dacitic liquid correspond to F = 0.2.

Thr pronounced departure of many less differentiated samples from the two theorical curves of fractional crystallization (perfect fractional and perfect equilibrium fractionation) of the selected parent magma in a closed system may be an indication of occasional replenishments of the magma chamber in open-system fractionation. These pulses may occur more commonly at the begining of the magmatic evolution in agreement with the HUPPER and SPARKS's model (1980) as suggested by cyclic units (JACKSON et al., 1975) in the lower cumulates. The increase of incompatible element abundances at a relatively low FeO^*/MgO ratio indicates that FeO and MgO were buffered during early stages of crystallization as supported by the absence of zoning and nearly constant mineral compositions in the early cumulates. Therefore the evolution of the Krapa series is probably best explained by partially fractional and partially equilibrium crystallization conditions in a magma chamber which changed from open system during its early evolutionary stages eventually to a closed system.

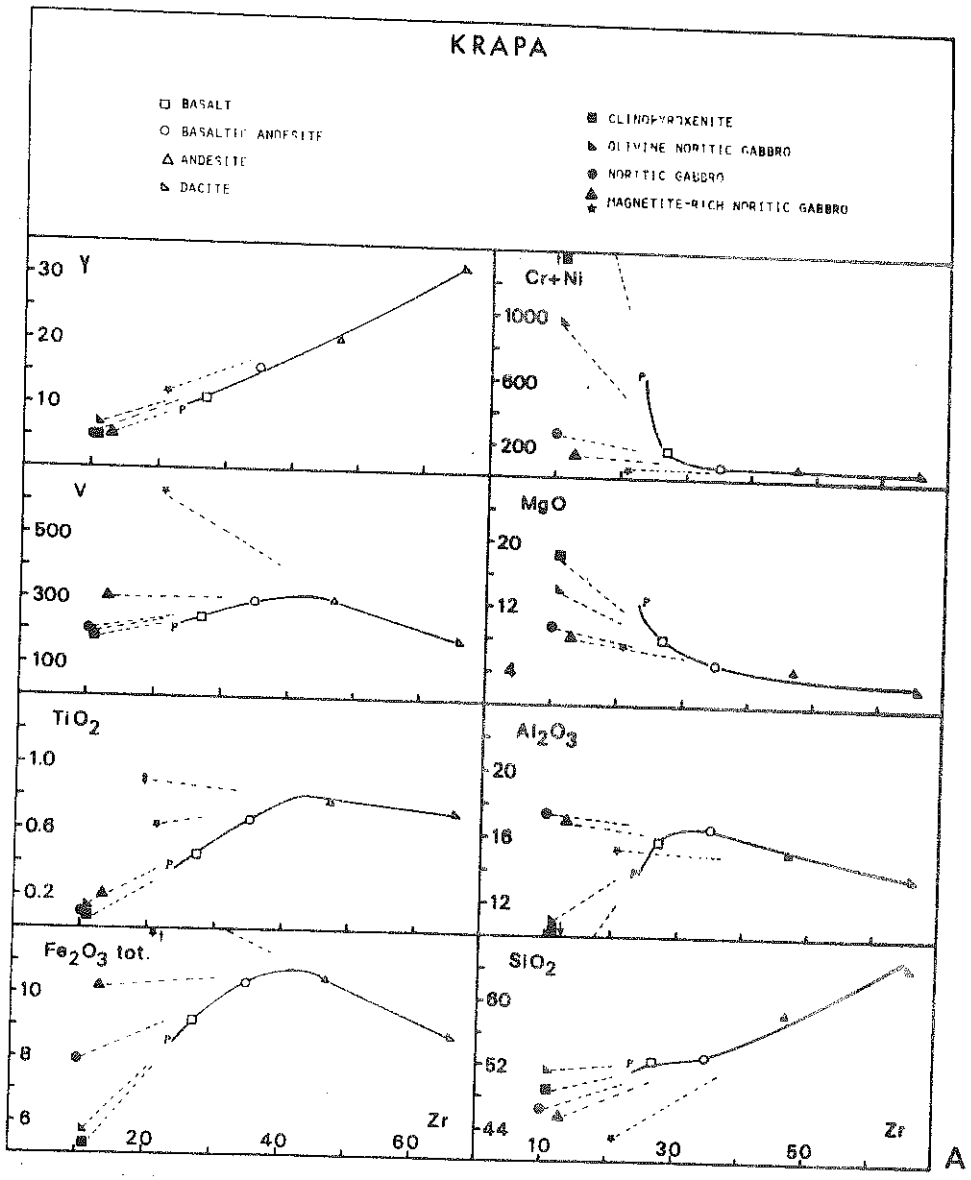
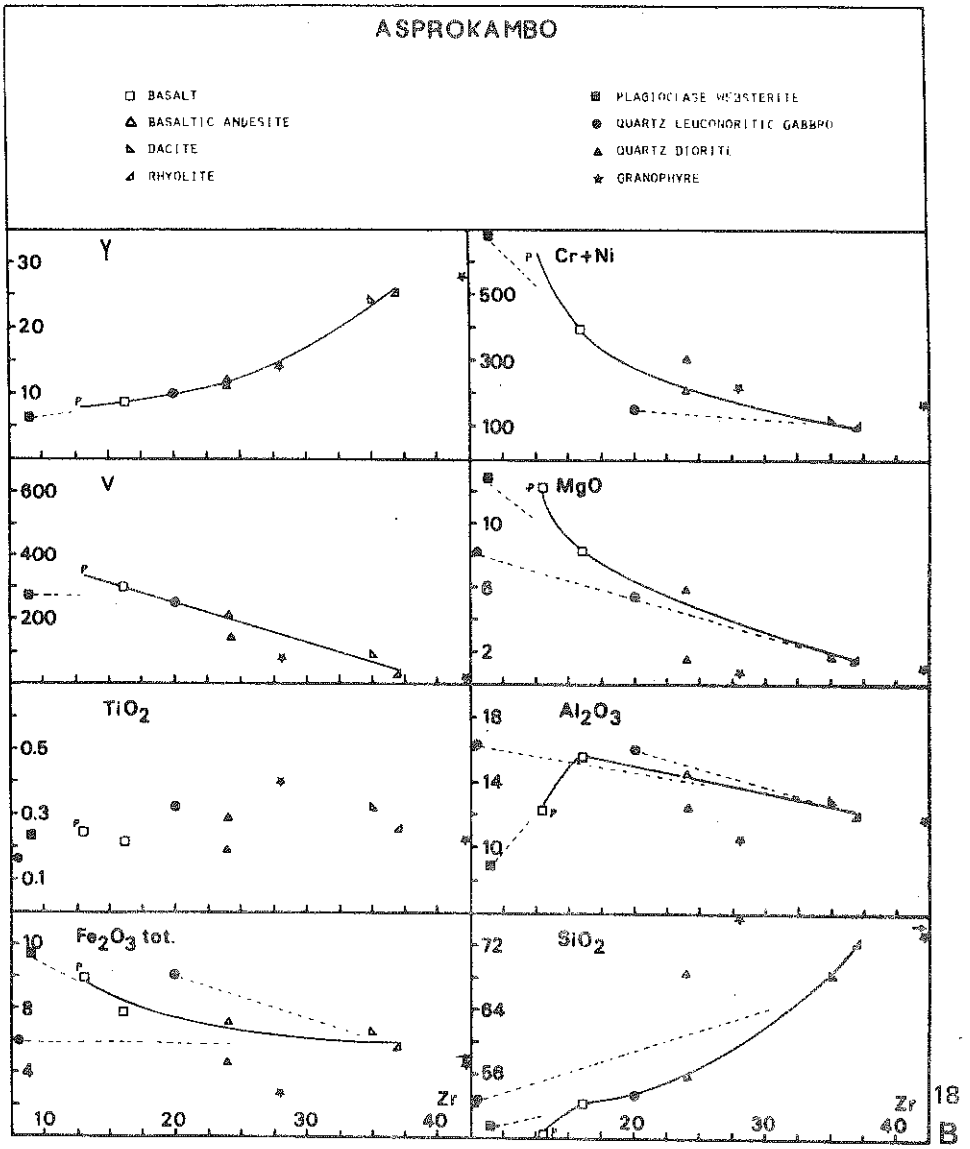


Fig. 18 - Variations of some major and trace elements vs Zr for the Krapa (A) and Asprokambo (B) sequences. Dashed light lines join representative liquid compositions (heavy lines) and the corresponding solids subtracted during fractionation. Zr of sample V3 estimated from Sm abundance (MONTIGNY, L975).

b) Asprokambo

The Asprokambo series is characterized by high silica saturation and a lack of Fe, Ti and V enrichment. In fig. 16b two basalts (V3, AP 13-1), a basaltic andesite (AP 1-3), a dacite (AP 1-1) and a rhyolite (AP 5-1) represent the evolution of erupted liquid compositions. The parental magma chosen is basalt V3 (MONTIGNY, 1975) which has $Mg/Mg+Fe^{++}$ atomic ratio of 0.71. Compared to the Krapa sequence, fractionation within the basaltic compositional range (up to F = 0.7) is



controlled mainly initially by olivine, followed soon after by two pyroxenes + Cr magnetite, these last phases forming the Pl-websterites from the lower cumulates. The other liquids, basaltic andesite to dacite and rhyolite (F = 0.7-0.2) result from the separation of gabbronorites and norites. Compared to the Krapa series, the Asprokambo more primitive compositions display significant variation in Zr^{D}/Zr ratio (fig. 19) which may be related to different degrees of partial melting of a single source or dynamic melting of a progressively more refractory source as in Troodos (SMEWING and POTTS, 1977). Commencing with V3 as a parental liquid, the most evolved samples compositions may be derived by perfect fractional (Rayleigh) crystallization in a closed system. This conclusion is supported by the occurence in the Asprokambo series of abundant orthocumulates, zoned minerals and numerous veins of evolved intrusives with high proportions of trapped liquid. Such rocks (Qz-norites, Qz-diorites, granophyres) plot around the line of liquid descent in variation diagrams (fig.18b).

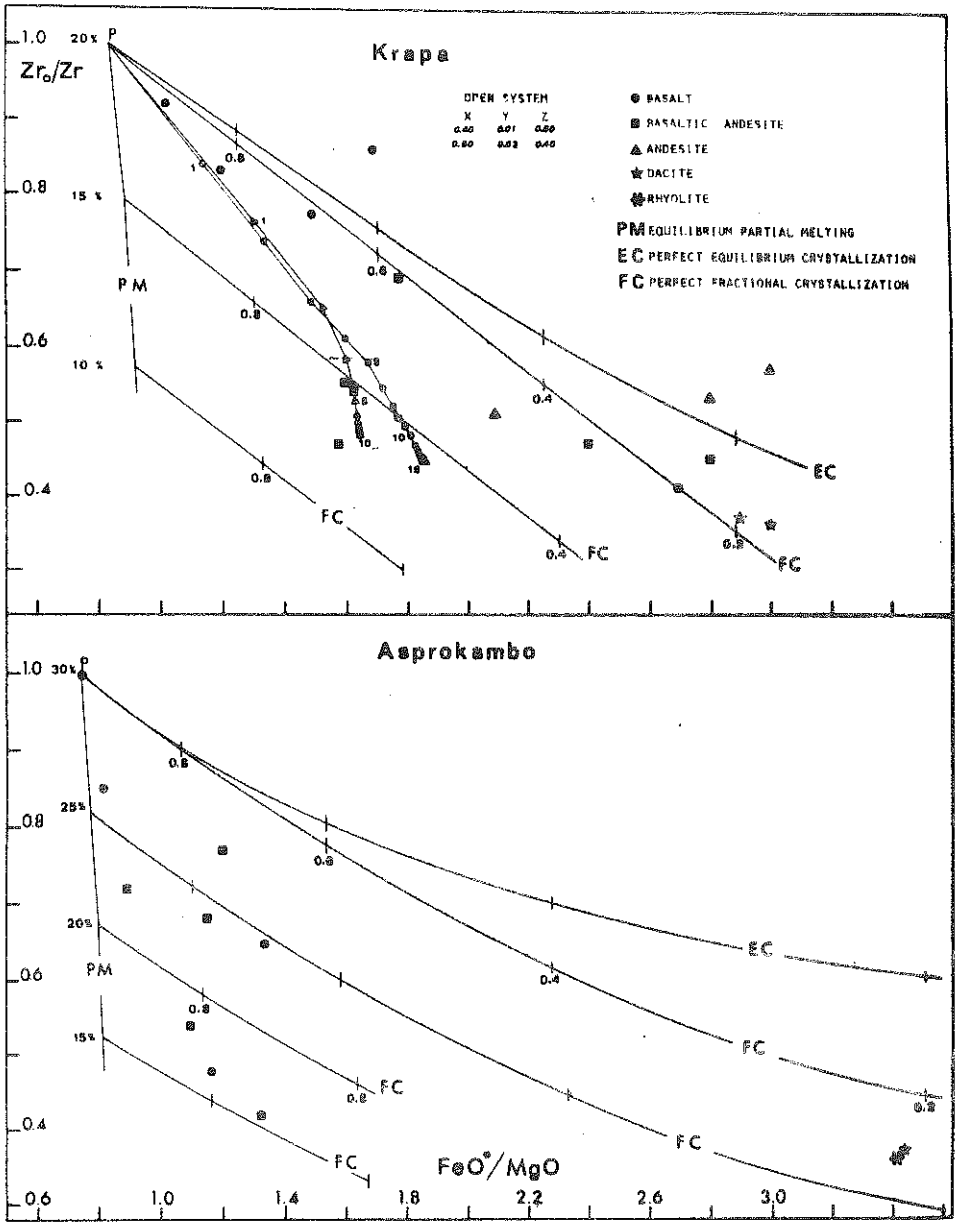


Fig. 19 - FeO^*/MgO vs. Zr_o (initial content of zirconium in the parental magmas, 24 and 13 ppm respectively for Krapa and Asprokambo sequences)/Zr. Liquid lines of descent by perfect fractional (FC) and perfect equilibrium (EC) crystallization in closed system are modelled (GAST, 1968, ARTH, 1976) from the assumed parental magmas (P) for decreasing values of the residual liquid fraction (figures). FeO^* and MgO mass balance calculations and $D_{Zr}^{s/l}$ = 0.35 and 0.5 (for Krapa and Asprokambo respectively) have been evaluated from the gabbroic solids plus trapped liquid subtracted and isolated from the main body of magma during fractionation (ALBAREDE, 1976) (see also fig. 16). Equilibrium melting trends (EM) (SHAW, 1970) indicate the liquid compositions which would derive by various partiel fusion degrees (figures) of the mantle source which produced the Krapa and Asprokambo parental magmas respectively by 20 % and 30 % of melting. To be noted that many least fractionated magmas from Krapa may have been produced by perfect fractional crystallization in an open system by periodically refilling of magma chamber employing bulk gabbroic distribution coefficients D_{Fe}=0.63, D_{Mg}= 1.9 and D_{Zr} = 0.35 and assuming : X = mass fraction of cumulate solid subtracted from the liquid in each cycle (figure), Y = mass fraction of parental magma added in each cycle (O'HARA, 1977 ; STERN, (1979).

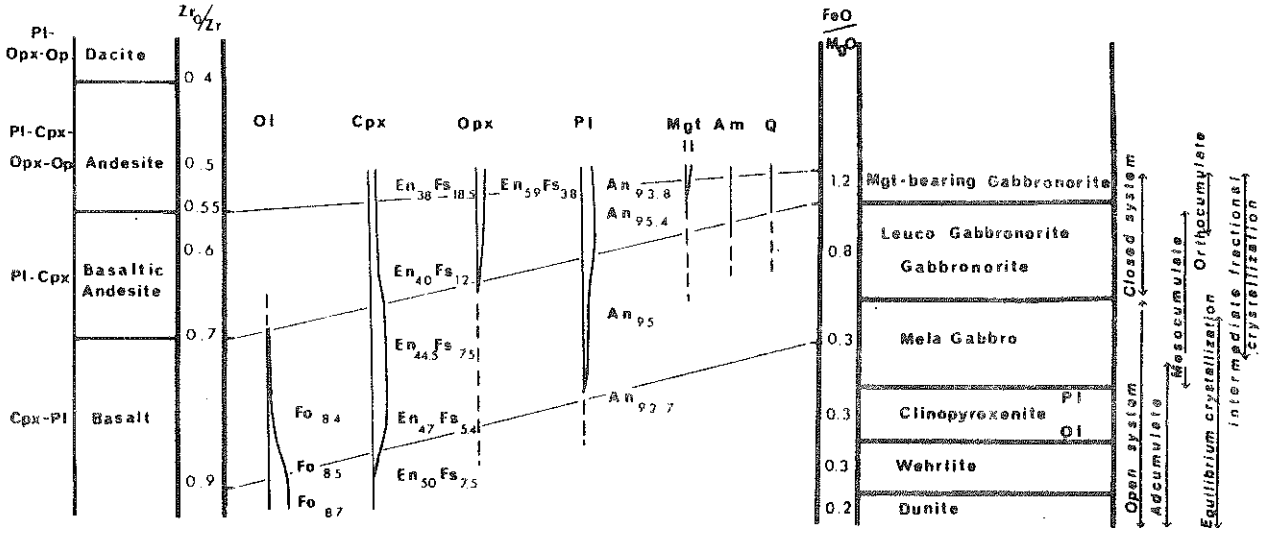


Fig. 20a- Crystallization order of the Krapa ophiolitic sequence. Ol= olivine, cpx= clinopyroxene, opx= orthopyroxene, pl= plagioclase, mgt= magnetite, Am= amphibole, Q= quartz, En = enstatite, Fs= ferrosilite, Fo= forsterite, An = anorthite

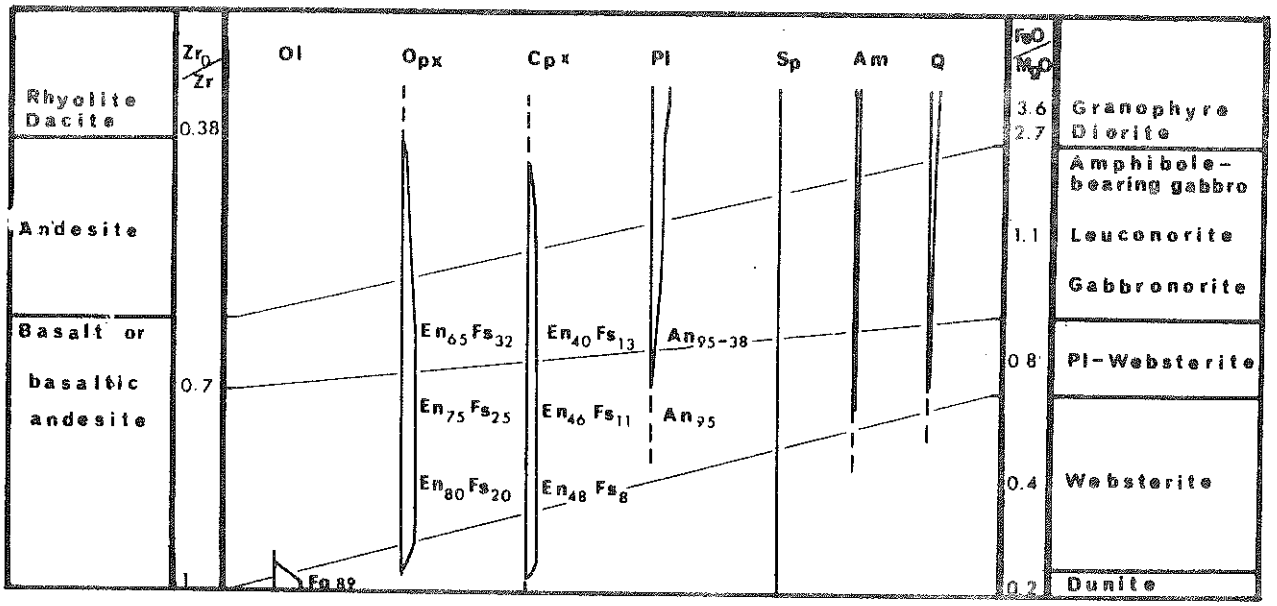


Fig. 20b-Crystallization order of the Asprokambo sequence. Same abreviation as in Fig. 20a

VII - MELTS AND MANTLE SOURCES RELATIONSHIPS

The least differentiated liquids of both series are characterized by low high field strengh elements (HFS), especially Asprokambo, and moderate to high silica saturation. Moreover a V shaped REE pattern at about 4 X chondritic level (fig. 22) emphasizes the very low incompatible element content. Genesis of such magmas requires strongly an impoverished mantle source and relatively necessitate a two-stage model source (GREEN, 1976 ; SUN and NESBITT, 1978). A suitable source for the genesis of such melts may be represented by a clinopyroxene-poor lherzolite such as are considered residue after MORB extraction (HAMLYN and BONATTI, 1980, KORNPROBST et al., 1981, BECCALUVA et al., submetted). These have compositions with : TiO_2 = 0.25-0.05, Al_2O_3/TiO_2 = 30-15 ; Zr =12-6 ; Zr/Y = 3-2 (fig.20) and have the modal mineralogy indicated in Fig.21. Equilibrium partial melting calculations (fig. 21) suggest that the primary magmas of the Krapa and Asprokambo series may be derived respectively by 20 and 30 % partial fusion of a depleted lherzolite (e.g. Appenines lherzolite S3-S4 from OTTONELLO and PICCARDO, 1979) leaving a residum of harzburgitic to dunitic composition. Basalt V3 (MONTIGNY, 1975) has REE levels from Eu to Lu indicating it may be derived by 30 % partial melting of a depleted lherzolite (samples GE 7-8 of OTTONELLO and PICCARDO, 1979) (fig. 22). However, the strong enrichment in REE in V3 implies that this depleted source must have been enriched in LREE prior to or during melting. The LREE enrichment required for this source (GE 7-8) ranges from 1.5 for Nd to 10 for La. The calculated residum shows a pattern very similar to the mantle harzburgite from Vourinos (V209 of MONTIGNY, 1975). Such LREE metasomatism probably occurs during a LIL enriching event associated with the influx of subducted slab derived hydrous fluids into overlying mantle.

Note that sources S_2-S_3-S_4 and residua r_1-r_2 fall in the field of mantle lherzolites from the Western Mediterranean ophilites (data from ERNST and PICCARDO, 1979, BOUDIER, 1976, OHNENSTETTER et al., 1975, unpublished data), whereas residua r_3-r_4 fall in the field of harzburgites and dunites from the Eastern Mediterranean ophiolites (data from WILSON, 1959, MOORES, 1969, JUTEAU, 1975, CAKIR, 1978, BINGOL, 1978, unpublished data).
For comparison parental magmas and fractionation trends (arrows) of lavas from Inzecca and Balagne ophiolites (Corsica) are also reported (BECCALUVA et al., 1977, GLOM, 1977).
Asterisk indicates chondrite composition according to SUN and NESBITT(1977).

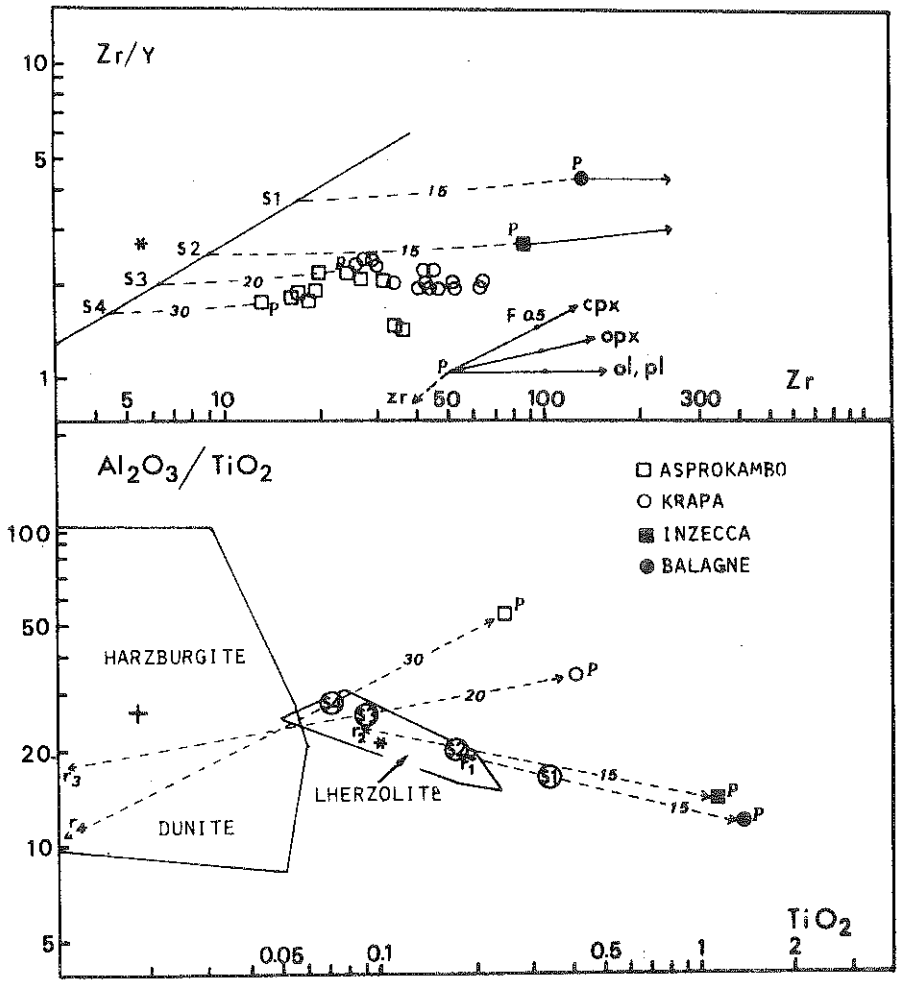


Fig. 21 - Zr/Y vs. Zr and Al_2O_3/TiO_2 vs. TiO_2 plot for lavas and hypabyssal rocks from Vourinos ophiolite.

Dashed lines join parental melts (P) and residual ($r_{1,2,3,4}$) produced by different degrees of partial fusion (figures indicate percentage) from hypotetical sources modelled as follows :

	Ol	Opx	Cpx	Sp	
S_1	0,55	0,20	0,20	0,05	
S_2	0,60	0,20	0,16	0,04	and the assumed eutectic composition
$E_{1,2}$	0,15	0,15	0,40	0,30	
S_3	0,65	0,20	0,12	0,03	
S_4	0,70	0,19	0,08	0,03	and the assumed eutectic composition
$E_{3,4}$	0,25	0,10	0,65	0,00	

Partition coefficients for Ti, Zr and Y are after PEARCE and NORRY (1979) Mass balance calculations for Al_2O_3 are based on the mineral chemistry of lherzolites from the Northern Apennines ophiolites (BECCALUVA et al., preprint). The modelled vectors indicate fractionation of olivine (Ol),plagioclase (Pl), orthopyroxene (Opx), clinopyroxene (Cpx) and zircon(Zr) under surface equilibrium conditions.

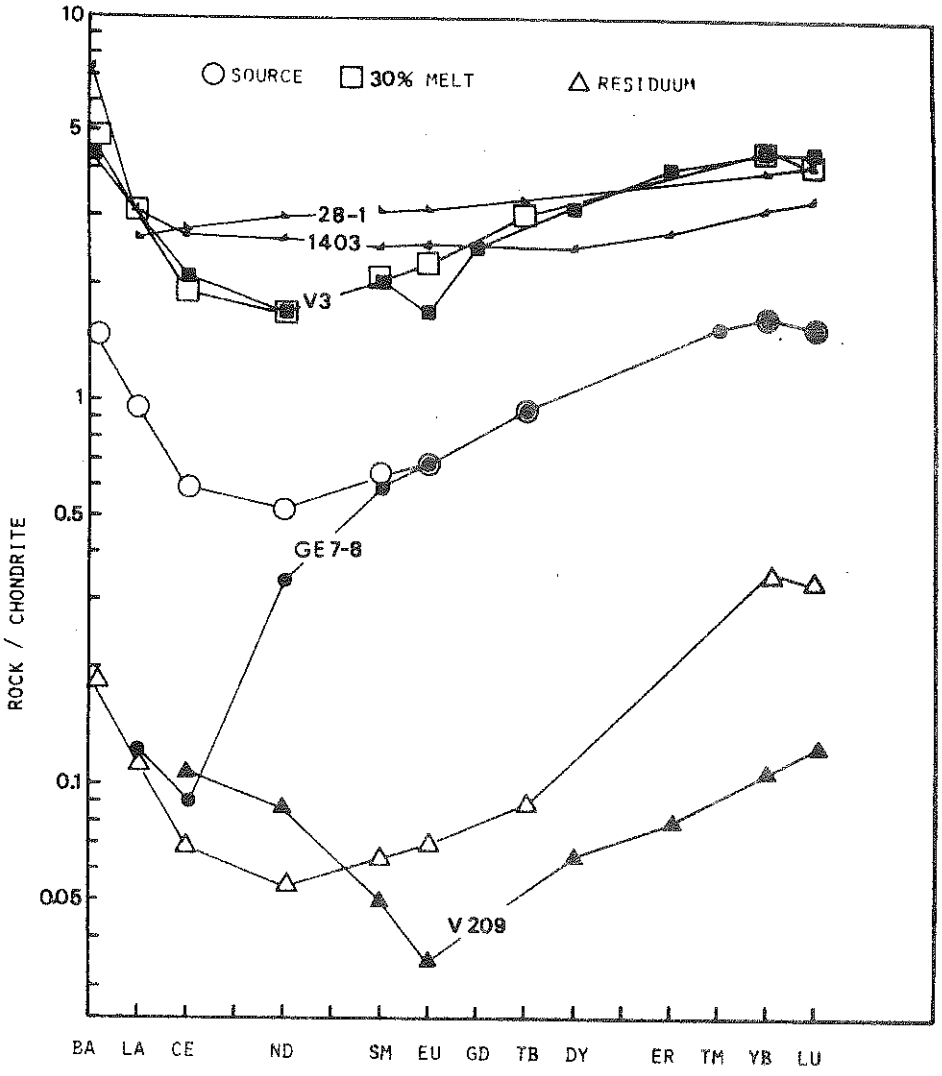


Fig. 22 - Chondrite-normalized (normalizing values according to NAKAMURA, 1974) REE patterns for basalt V3 and harzburgite V 209 from Vourinos ophiolite (analyses after MONTIGNY, 1975). These patterns approximate respectively melt and residuum resulting from 30 % equilibrium partial fusion of an hypotetical source (modelled as S4 in fig. 19) with HREE like lherzolite from Ligurian ophiolites (sample GE7-8 after OTTONELLO et al., 1979) and enriched in LREE and LILE.

Partition coefficients are after ARTH,(1976), and LEEMAN (1976) for La, Tb and spinel. For comparison REE patterns of boninites have also been reported (sample 1403 after SUN and NESBITT, 1978, sample 28-1 after HICKEY and FREY, 1980).

VIII CONCLUSION

From field and petrological data it appears that at least two main volcano-plutonic complexes exist in the Upper Jurassic Vourinos ophiolite, namely the Krapa and the Asprokambo complexes, the later younger than the former. The Krapa sequence shows clear island arc tholeiitic affinity whereas the Asprokambo sequence shows affinities with very low TiO_2 lavas found in some oceanic iceland arcs. In the Asprokambo cumulates, continuous crystallization of V_2O_3-rich chromian and aluminian titaniferrous magnetite is largely responsible for the regular Fe, Ti and V decrease in residual liquids, and the precipitation of amphibole and Qz suggest similarities with calc-alkaline magmas. Both the Krapa and Asprokambo series may be assigned to evolution in an island arc setting, and the presence of very low TiO_2 series in dykes later than the IAT sequence suggested a fore-arc setting (BECCALUVA et al., 1979). These evidence, as well as the existence of pelagic, reefal and neritic sedimentation deposited shortly after the ophiolite formation are compatible with the generation of Vourinos ophiolites and other Greek ones in multiple oceanic basins and related island arc developped between micro-continent sheared during the general convergence between Apulian and European plates since the Trias (BEBIEN et al., 1980).

All the petrological features indicate that the evolution of the Krapa series is more compatible with an intermediate fractional equilibrium of crystallization conditions in a partially open system. In the Asprokambo sequence, more or less silica saturated parental liquid may suffer Pl-websterite and gabbronorite fractionation to give a wide range of basalt and andesitic basalt compositions. Dacite may derive from evolved andesitic basalt through Cpx bearing norite extraction.

Trace-element data indicate that nearly pure conditions of fractional crystallization may have prevailed in the small Asprokambo chambers.

Petrologic studies permit to put some constraints about the evolution of magmatic chamber in Vourinos ophiolites. The development of the lower cumulates from Krapa occur in a zone pf persistant high thermal activity as indicated by the extent of subsolidus reaction in the coarse grained adcumulates. Heat and mass transfer may have been favored by the successive replenishments in the open chamber. Liquids may be expelled to the surface after extensive fractionation. The more abundant evolved liquids correspond to the vanished stage of the magma chamber in which upper cumulate crystallise.

There the frequency and volume of magma entries decrease as testify by the late Asprokambo sequence. Moreover, this later magmatism indicate that composition of the imput pulse varies with time. Cross-cutting relationships, squelettic minerals, the presence of orthocumulates and the geochemical evolution of liquids argue in favor of limited chamber cooling rapidely.

The least fractionnated lavas of both Krapa and Asprokambo area have geochemical features compatible with a high degree of partial melting (20-30 %) of different depleted mantle source in high field strengh element. The silica saturation of the magma may have been favoured by high water pressure easily provindly in a subduction zone in a 10-20 Kbar range. The mantle source may be represented by clinopyroxene poor lherzolite (like those associated to Western ophiolites) of a downgoing slab previously depleted by MORB type magmatism. Asprokambo primary melts probably result from a more depleted mantle source than Krapa and necessitate more effective inconguent melting of orthopyroxene giving a dunitic-harzburgitic residue.

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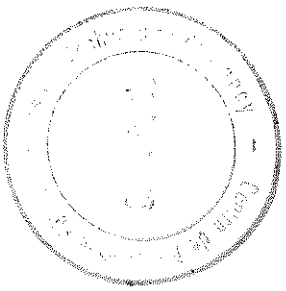

QUATRIÈME PARTIE

LES MANTEAUX OPHIOLITIQUES.

LEUR CARACTÉRISTIQUES PÉTROLOGIQUES ET STRUCTURALES

IMPORTANCE DE LA REMONTÉE DIAPIRIQUE.

"Although comagmatism of gabbro with peridotite in harzburgite-type complexes is accepted by many petrologists, some do not believe that similar relations occur with so-called "high-temperature" peridotites of the lherzolite type".

T.P. THAYER, 1972.

QUATRIÈME PARTIE

LES MANTEAUX OPHIOLITIQUES. LEURS CARACTÉRISTIQUES PÉTROLOGIQUES ET STRUCTURALES IMPORTANCE DE LA REMONTÉE DIAPIRIQUE.

I - INTRODUCTION

II - LES STRUCTURES DANS LES PERIDOTITES

1 - Relations entre déformation, fluage et anisotropie sismique

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IV - LES CARACTERISTIQUES PETROGRAPHIQUES DES PERIDOTITES OPHIOLITIQUES

1 - L'environnement géotectonique

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V - CONCLUSION

QUATRIÈME PARTIE

LES MANTEAUX OPHIOLITIQUES. LEURS CARACTÉRISTIQUES PÉTROLOGIQUES ET STRUCTURALES IMPORTANCE DE LA REMONTÉE DIAPIRIQUE.

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IV.3 - JACKSON M.D. and OHNENSTETTER M. - Peridotite and gabbroic structures in the Monte Maggiore massif, Alpine Corsica. J. of Geol., 89, 703-719.

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I - INTRODUCTION

Le manteau est séparé des croûtes océanique et continentale par la discontinuité sismique de MOHOROVICIC située à des distances variables de la surface (5 à 40 km des océans aux continents). On admet que le manteau rubané caractérisé par la présence d'une fabrique le plus souvent tectonique, est constitué de péridotites au moins vers sa partie supérieure. Quelle que soit la profondeur d'équilibration supposée d'après la nature de la phase alumineuse associée (grenat, spinelle ou plagioclase), des processus pétrologiques comme fusion partielle, fractionnement et contamination ont pu l'affecter à plusieurs reprises. Ces différentes conditions pétrogénétiques doivent conduire à une zonalité du manteau qui reflète l'évolution de la lithosphère avec le temps.

La présence de fragments de manteau dans les formations superficielles est due à des remontées tectoniques et/ou magmatiques. On peut l'échantillonner dans les massifs de type alpin liés à des fragments de croûte continentale et/ou océanique, dans les nodules des basaltes alcalins et kimberlites et par dragage et forage dans les océans actuels.

Leur remontée accompagnée ou non de fusions partielles serait amorcée dans des zones distensives tel que le laissent supposer les formations volcaniques affleurant à proximité des péridotites. Dans ce contexte, les ophiolites constituent un champ d'investigation remarquable pour deux raisons. La première est que les déformations mantellaires peuvent être replacées par rapport aux plans structuraux et lithologiques principaux de la croûte, de même que par rapport aux zones de migration des magmas. La seconde est que la composition calculée des magmas parentaux peut être comparée avec la nature des liquides en équilibre avec le résidu observé.

Les parties précédentes ont mis en évidence entre les ceintures ophiolitiques des discontinuités pétrologiques au niveau de la croûte que des mécanismes de basses pressions ne suffisent pas à expliquer. Cette quatrième partie va montrer que des discontinuités géochimiques et minéralogiques existent dans le manteau en fonction des types d'ophiolites désignées précédemment (KORNPROBST et al., 1981 ; 1982, IV 1 et 2). Ces ophiolites seraient formées dans des sites géotectoniques différents où la nature du manteau liée au renouvellement de l'asthénosphère permet d'interpréter les distinctions observées dans les péridotites

et les laves et de suivre l'évolution du magmatisme avec le temps.

Par ailleurs, des discontinuités dégagées au sein des ophiolites du type II selon la classification pétrographique de ROCCI et al. (1975) ont permis d'aboutir à une classification en deux groupes (OHNENSTETTER M., 1980, I.2): dans les ophiolites à structure verticalisée, les péridotites présentent des textures relictuelles de diapirs à la différence des massifs à plans lithologiques et structuraux sub-horizontaux. Les divergences constatées doivent être expliquées en mettant en oeuvre des processus tectono-magmatiques depuis la formation des diapirs jusqu'à leur emplacement actuel. Cette démarche permettra d'illustrer certains aspects de la mise en place des ophiolites mais aussi de préciser la chronologie des phénomènes magmatiques : fusion et cristallisation par rapport aux structures observées. Dans ce cadre, les études sur une écaille de péridotites du Cap Corse dont les grands traits pétrographiques et structuraux ont été décrits par ailleurs (OHNENSTETTER et OHNENSTETTER, CROCHEMORE, 1975 ; OHNENSTETTER et al., 1975 ; JACKSON et OHNENSTETTER, 1981 ; IV.3) m'ont amenée à analyser une suite d'évènements tectoniques et magmatiques dans un diapir s'élevant dans une zone d'accrétion (OHNENSTETTER, manuscrit, IV.4). Ces processus correspondent aux premières étapes de la structuration d'un massif de péridotites. Les textures qui en résultent sont souvent oblitérées postérieurement.

II - LES STRUCTURES DANS LES PERIDOTITES

Les péridotites qui proviennent de zones à hautes pressions et températures élevées peuvent être considérées comme des roches métamorphiques qui ont été affectées par des déformations successives accompagnées ou non de fusion et cristallisation. L'interprétation des conditions de mise en place des péridotites et leur importance vis à vis de l'assemblage ophiolitique ne peut se faire sans une certaine compréhension des mécanismes de fluage dans le manteau par rapport aux structures observées. C'est pourquoi, on décrira d'abord la position des axes cinématiques par rapport aux plans remarquables, puis les principales déformations des massifs ophiolitiques afin de proposer une chronologie des mécanismes tectoniques. Celle-ci sera ensuite précisée par rapport à la migration des liquides dont les filons plus ou moins sécants en constituent le témoignage évident. Ces dernières considérations nous conduiront naturellement à discuter de la tectonophysique de la fusion partielle bien que les liquides successifs issus de ce processus aient disparu.

1 - Relations entre déformation, fluage et anisotropie sismique

1.1 - Déformation et fluage

Les roches issues du manteau présentent le plus souvent un plan de variation lithologique et un plan d'aplatissement S1 que l'on peut suivre cartographiquement dans les massifs de type alpin. Elles possèdent aussi une fabrique obtenue grâce à la position des axes cinématiques des minéraux par rapport aux plans remarquables considérés. Leur position permet d'apprécier la nature des processus possibles à l'origine des roches (DEN TEX, 1969). Quatre mécanismes principaux peuvent être retenus selon H. AVE LALLEMANT (1976) : chute de minéraux, fluage dans une bouillie de minéraux, glissement intra-cristallin et recristallisation syntectonique. Un cinquième mécanisme peut être envisagé si les déformations ont lieu en présence d'un liquide (GEORGE, 1977).

Pour les minéraux accumulés à partir d'un magma, le plan (010) de l'olivine serait à un angle faible des lignes de flux ou magma proches du plan de rubanement. Cette fabrique est courante dans les massifs alpinotypes tandis que la fabrique Y est exceptionnelle (DEN TEX, 1969) sauf dans le massif d'Alpe Arami (MOCKEL, 1969). Elle est généralement interprétée comme étant d'origine tectonique, liée soit au glissement intracristallin, soit à des recristallisations syntectoniques (AVE LALLEMANT et CARTER, 1970 ; NICOLAS et al., 1973).

En présence d'un fluide intermédiaire des dissolutions et recristallisations de l'olivine sous contraintes uniaxiales, caractéristiques d'une piezocristallisation donneraient une même fabrique (DEN TEX, 1969). Ce processus est à rapprocher des phénomènes de transfert de matière sous contrainte (fig. IV.1) lors de fusion partielle (GEORGE, 1975, 1978) à des températures au-dessus du solidus (hypersolidus creep of peridotites). Il permettrait d'expliquer aussi la formation du rubanement (DICK et SINTON, 1979).

Deux types d'évolution des structures métamorphiques peuvent être distinguées, l'une à symétrie orthorhombique, l'autre à symétrie monoclinique. La première la plus couramment observée (BOUDIER, 1976) s'accompagne d'une fabrique planaire avec $[010]_{01}$ perpendiculaire à S_1 et $[100]_{01}$ parallèle à la linéation. Par exemple, elle est observée dans les roches à gros grain de Lanzo (BOUDIER, 1976) et dans ce[illegible] du Cap Corse (JACKSON et DHNENSTETTER, 1981 ; IV.3) ou encore dans les péridotites du Vourinos (ROSS et al., 1980). Elle correspond à un glissement dans la zone [100] sur un plan (010) à de hautes températures qui serait

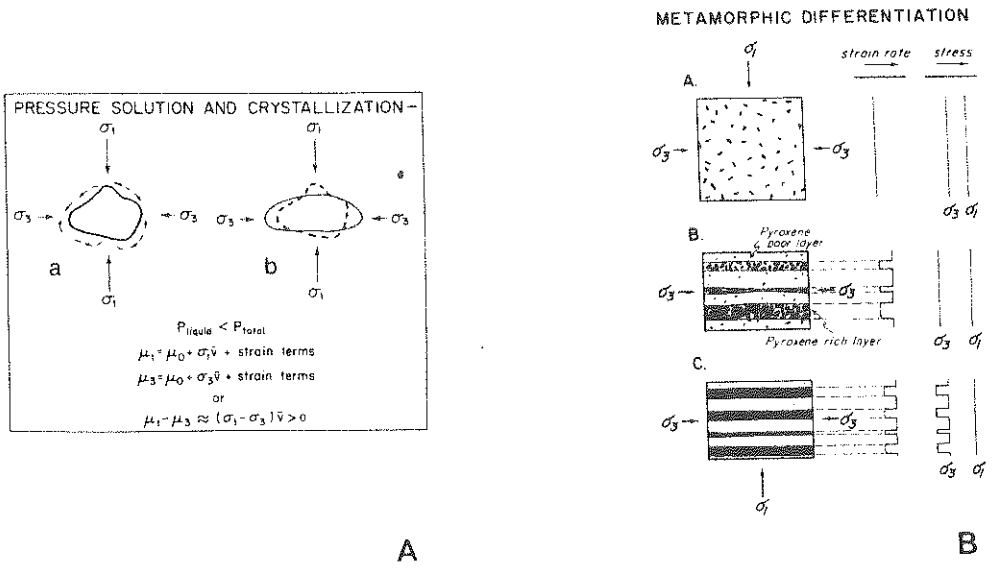


Fig. IV.1 - A, Transfert de masse sous contrainte grâce à une solution ; B, Formation du rubanement pyroxénitique dans les péridotites lors d'une fusion sous contrainte, d'après DICK et SINTON (1979).

le plus important pour le manteau (CARTER et AVE LALLEMANT, 1970). La seconde symétrie entraîne une fabrique linéaire avec $[100]_{01}$ oblique sur la linéation et les deux autres axes dont $[010]_{01}$ en couronne autour de la linéation. On la rencontrerait dans les roches à texture porphyroclastique ayant été soumises à une déformation rotationnelle.

La fabrique de l'olivine permet de connaître les plans de glissement et de les replacer par rapport à l'ellipsoïde des déformations et des contraintes principales. Les relations angulaires entre axes et plans cinétiques et ceux de la déformation sont représentées sur la fig. IV.2 (voir aussi la légende) dans le cas d'une déformation rotationnelle. La contrainte maximum tendrait vers $[010]_{01}$ (NICOLAS et al., 1973) ce qui est compatible avec les données expérimentales qui, pour des températures supérieures à 500º et des pressions de 5 à 30 Kb, montrent que $[010]_{01} = \sigma_1$ tandis que les deux autres axes sont dans le plan $\sigma_2 - \sigma_3$ (AVE LALLEMANT et CARTER, 1970).

D'une manière générale, d'après les données de terrain et les expérimentations, les relations entre axes cinématiques, plan de fluage et foliation soulignent que lors d'une déformation rotationnelle le plan de cisaillement tourne pour se paralléliser avec le plan contenant les contraintes minimums.

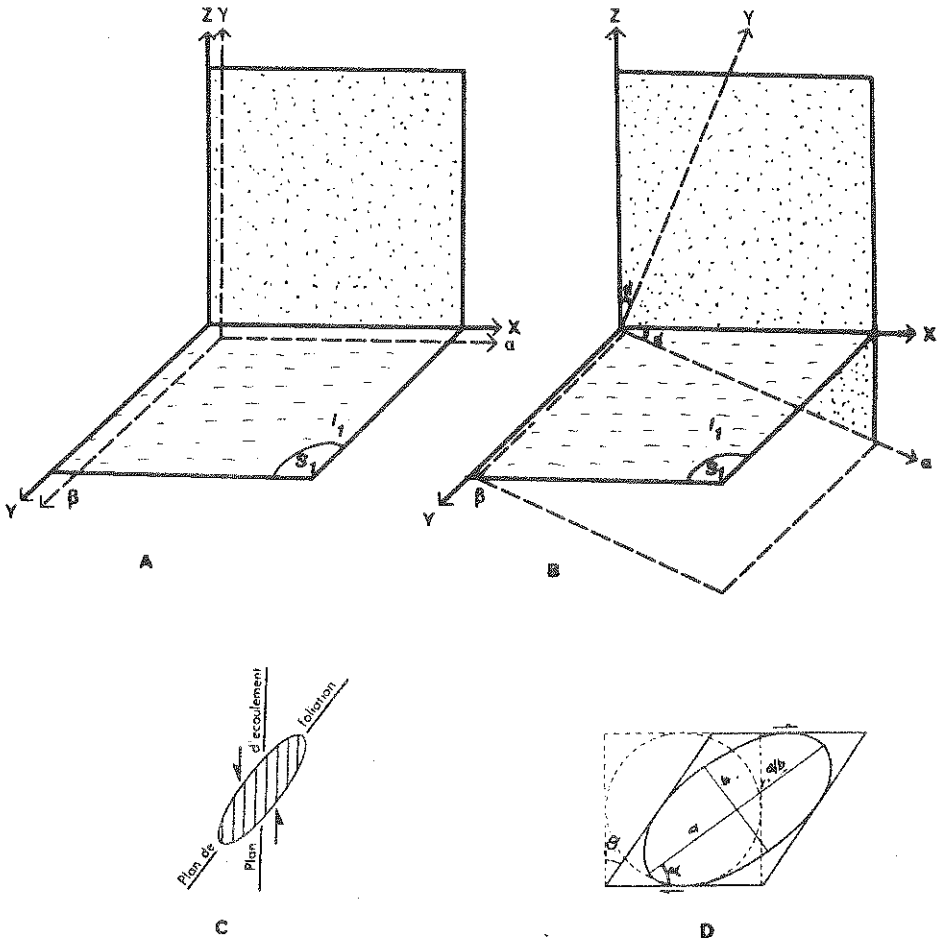


Fig. IV.2 - Position des axes principaux de la déformation et des axes cinématiques dessinés d'après BOUDIER (1976) et BOUDIER et NICOLAS (1977) : A, cas d'un cisaillement pur ; B, cas d'un cisaillement simple. X, élongation du minéral ou des agrégats de minéraux l_1 ; Y, axe intermédiaire, perpendiculaire à X dans le plan de foliation S_1 ; Z, axe perpendiculaire au plan d'aplatissement XY. Les axes cinématiques : α , ligne de flux qui correspond dans les projections au maximum des orientations préférentielles des directions de glissement parallèles aux [100] de l'olivine et [001] de l'enstatite. β , est perpendiculaire à α dans le plan de fluage. β est toujours parallèle à Y et à l'intersection entre plans de foliation et de fluage. Y est perpendiculaire au plan de fluage. Le maximum correspond aux plans (010) et (001) de l'olivine (DEN TEX, 1969), respectivement à hautes et basses températures et au plan (100) de l'enstatite. L'angle entre X et α : $\propto$ est relié à l'angle de cisaillement par la relation suivante cot $2\alpha = 1/2 \tan\theta$ (fig. C) ; tiré de NICOLAS et al., 1973). Les plans de foliation (S1) et de fluage (en blanc) sont indiqués de même que la section propre aux observations (en pointillé). En D, les traces des axes X et $\propto$ sont soulignées à partir d'un minéral déformé dans une section XZ ; tiré de F. BOUDIER (1976).

1.2 - Anisotropie sismique

Le manteau tout comme la croûte océanique est souvent connu d'un point de vue géophysique mettant en évidence des discontinuités sismiques. Les différentes vitesses mesurées, en particulier dans le manteau océanique (RAITT et al., 1969) peuvent correspondre aux orientations cristallographiques préférentielles des grains d'olivine, minéral le plus abondant, comme cela semble se vérifier d'après des études expérimentales sur l'olivine et sur le terrain (CHRISTENSEN, 1966 ; PESELNICK et al., 1974). Dans un océan, il en résulte que si $[100]_{01}$ est parallèle à la vitesse maximum et que celle-ci soit perpendiculaire à la ride, on peut s'attendre à trouver un plan de cisaillement important sub-horizontal à l'écart de la ride (FRANCIS, 1969) et σ_1 sub-vertical (AVE LALLEMANT et CARTER, 1970). Ces considérations sont à l'origine des discussions sur la rotation des lignes de flux d'un diapir mantellaire qui permet, quel que soit l'environnement géotectonique, la remontée des péridotites depuis les zones profondes (MOORES et MAC GREGOR, 1972 ; OXBURG et PARMENTIER, 1977). Dans un diapir, la foliation, guidée par les parois du diapir et les cisaillements qui le recoupent est subverticale. Cette disposition implique que $[010]_{01}$ soit horizontal de même que σ_1, tandis que σ_3 est vertical, parallèle à $[100]_{01}$ (AVE LALLEMANT, 1976 ; ROSS et al., 1980). La fabrique d'un diapir n'est donc pas conforme à celle supposée pour le manteau supérieur à l'écart de la ride d'après les vitesses sismiques. Elle nécessite une rotation des lignes de flux pour que γ en position verticale dans le diapir devienne horizontal (fig. IV.3A). Cependant, à l'aplomb d'une ride médio-océanique, la disposition des lignes de flux qui est parallèle à celle des isothermes est encore très discutée (OXBURG, 1980). Sur la fig. IV.3B, on peut en effet considérer la géométrie des zones ascensionnelles, étroites dans le modèle de MACKENZIE (1967), plus larges, dans celui de PARKER et OLDENBURG (1973). Dans les ophiolites, selon que l'on admette ou non la rotation des lignes de flux, deux dispositifs structuraux peuvent être envisagés (NICOLAS et VIOLETTE, 1982). Dans l'un, la lithosphère avec des structures relictuelles de diapir est entraînée passivement à l'écart de la ride en accord avec les modèles de MACKENZIE (1967) et de CANN (1974) (voir aussi à ce propos, les fig. II.1A et B). Dans l'autre, les lignes de flux tournent à angle droit avant de supporter les cumulats ophiolitiques (JUTEAU et al., 1977). Il est donc possible de concevoir plusieurs types de mise en place primaire des ophiolites qui seront appréciés d'après la disposition du plan de fluage dans les péridotites par rap-

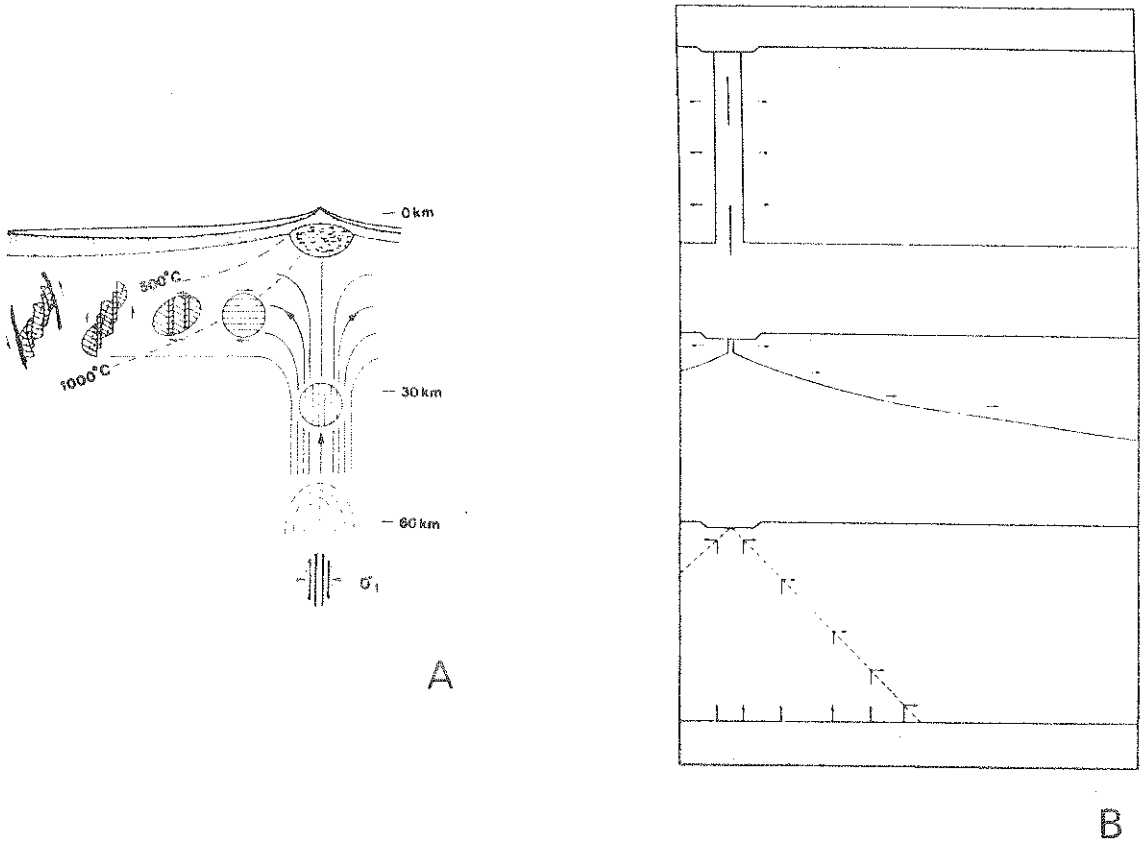


Fig. IV.3 - Sections perpendiculaires à la ride montrant : A, la rotation de Y (parallèle aux lignes de flux) à 30 km de la ride et les déformations mylonitiques postérieures ; tiré de ROSS et al. (1980). B, les différents modèles de remontée asthénosphériques, tiré de BOTTINGA et ALLEGRE (1976) ; en haut, le modèle d'intrusion en dike de MACKENZIE (1967) ; au centre, celui de PARKER et OLDENBURG (1973) ; en bas, la remontée asthénosphérique de LANGSETH et al. (1966).

port à la bordure de la chambre magmatique, au rubanement des cumulats et au champ de roches filoniennes grenues qui recoupent le manteau.

2 - Le gisement des péridotites ophiolitiques

2.1 - Nomenclature des roches et des textures

Les péridotites ophiolitiques sont constituées de lherzolites plus ou moins riches en Cpx et de harzburgites auxquelles on peut appliquer la classification de STRECKEISEN (1974). Le rubanement est souligné par des variations dans la proportion de Cpx et/ou Opx par rapport à l'olivine d'ou résultent des dunites, wehrlites, orthopyroxénolites, clinopyroxénolites et websterites. Elles sont équilibrées dans le faciès à spinelle ou à plagioclase (O'HARA, 1967). La dénomi-

nation des textures nécessite aussi une grande rigueur dans la mesure où elles permettent de caractériser certains états de la déformation en liaison avec les conditions de gisement. L'usage a progressivement entériné l'emploi de quelques termes (MERCIER et NICOLAS, 1975, HARTE, 1977, PIKE et SCHWARZMAN, 1977) : 1, texture protogranulaire (MERCIER et NICOLAS, 1975) ou texture à gros grain (coarse grained texture) de HARTE et al. (1975) équante ou tabulaire (HARTE, 1977). 2, texture porphyroclastique ; 3, texture granulaire tabulaire ou en mosaïque (MERCIER et NICOLAS, 1975) ou granuloblastique tabulaire ou équante (HARTE, 1977) ; localement des roches à grain très fin laminées résulteraient d'un processus de superplasticité (BOULLIER et NICOLAS, 1975). La réduction de grain des textures de 1 à 3 traduit une augmentation des contraintes différentielles (MERCIER, 1977); 4, texture poeciloblastique (MERCIER, 1977). Elle s'obtiendrait lors d'une réduction des contraintes différentielles par une coalescence des olivines autour des autres minéraux dont la croissance est néanmoins orientée. Le grain de la roche augmenterait alors. Pour PIKE et SCHWARZMAN (1977), les roches présentant cette texture auraient été en équilibre avec un liquide : texture pyrométamorphique (pyrometamorphic texture).

2.2 - La chronologie des déformations dans les péridotites ophiolitiques

La mise en place des péridotites ophiolitiques conduit à plusieurs stades de déformation dont les derniers, les mieux préservés ont totalement ou en partie, oblitérésles textures initiales. Si dans chaque massif ophiolitique, il est possible d'établir une chronologie des déformations (on reconnaît en moyenne deux à trois phases), il apparaît de plus en plus difficile à la lecture des travaux les plus récents de placer une coupure nette entre elles qui correspondrait à un laps de temps important entre mise en place primaire, dans la zone d'accrétion, et une mise en place secondaire sur le continent et les formations de marge (voir à ce propos la deuxième partie sur les questions de durée). La chronologie proposée pour un massif ne correspond, en général, qu'à une partie de l'évolution du manteau. Pour reconstituer l'histoire d'une écaille mantellaire depuis le stade diapir jusqu'au lieu d'incorporation final, il apparaît nécessaire de juxtaposer les états successifs de la déformation à partir de l'étude de plusieurs massifs de textures différentes.

221 - Les déformations tardives

Quel que soit le massif envisagé, des déformations précoces aux déformations tardives, la pression lithostatique décroît tandis que les contraintes différentielles augmentent conduisant en fin d'évolution à des couloirs de mylonites. Celles-ci peuvent se former à des profondeurs inférieures à 30 km lorsque les lignes de flux du diapir tournent (ROSS et al., 1980, fig. IV.3A), lors d'une subduction intraocéanique (PARROT et WHITECHURCH, 1978, NICOLAS et LE PICHON, 1980, BOUDIER et al., 1982) ou encore dans des zones transformantes (SALEEBY, 1977, PRINZOFFER et NICOLAS, 1979, REUBER, 1982). Elles détruisent soit les textures de cumulats équilibrées sous des températures de 1200 à 900°, à 15-40 km sous des contraintes différentielles de 10-100 bars (ROSS et al., 1980), soit les textures porphyroclastiques obtenues à l'écart de la ride (NICOLAS et LE PICHON, 1980). On estime que les mylonites se forment à de plus faibles températures (700 à 900°).

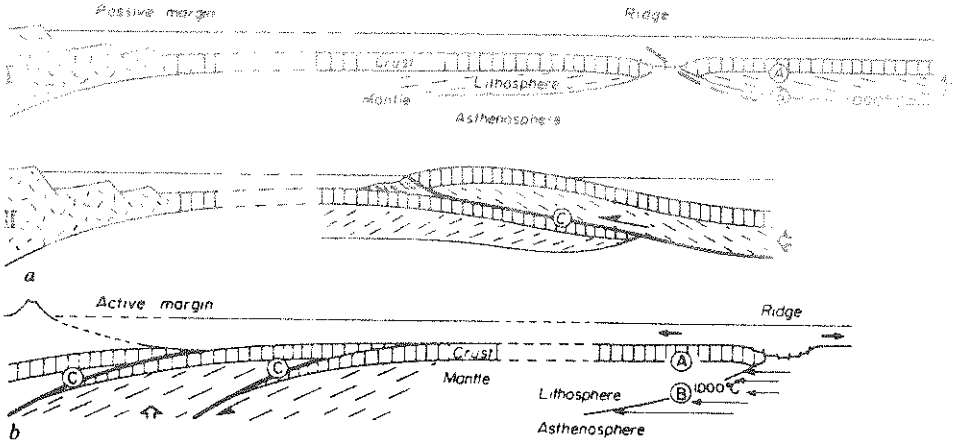


Fig. IV.4 - Site géotectonique des charriages intra-océaniques ; a, près d'un axe d'extension pour une ride assez rapide aux environs de 1000° ; b, dans une zone de subduction proche d'une marge active, modifié d'après NICOLAS et LE PICHON (1980). Ⓐ, Ⓑ, Ⓒ zones de déformation lithosphériques et asthénosphériques des péridotites. Tiré de BOUDIER et al. (1982).

Dans le cas d'une subduction intra-océanique, la mylonitisation des péridotites est contemporaine du dynamométamorphisme qui affecte les amphibolites en position infra-péridotitique (NICOLAS et LE PICHON, 1980, BOUDIER et al., 1982). La figure IV.4 indique les lieux théoriques possibles où le métamorphisme thermique peut avoir lieu avant que les péridotites ne soient char-

riées sur le continent. Ces modèles impliquent que la série volcano-sédimentaire qui appartient à la partie supérieure du cortège ophiolitique soit métamorphisée à proximité de la ride, en raison des conditions thermodynamiques (NICOLAS et LE PICHON, 1980, BOUDIER et al., 1982). Nous avons souligné (BEBIEN et al.,1980, II.2, et deuxième partie) que,compte tenu du chimisme des amphibolites fort différent de celui des laves associées, l'accrétion ophiolitique,responsable indirect du dynamométamorphisme,intervenait à proximité du continent. Ces faits s'opposent à la création d'une croûte océanique assez étendue qui aurait progressivement disparue pour ne laisser que la dernière portion proche de la ride.

222 - Les déformations les plus précoces reconnues

Pour la plupart des auteurs, les déformations les plus anciennes reconnues ont eu lieu dans le manteau, soit dans un diapir comme pour les massifs du Vourinos (MOORES, 1969, ROSS et al., 1980), de Canyon Mountain (AVE LALLEMANT, 1976), Acoje (VIOLETTE, 1980) et du Troodos (MOORES et VINE, 1971, GEORGE,1975, 1978), soit à l'écart de la ride comme pour les massifs d'Antalya (JUTEAU et al., 1977, d'Oman (BOUDIER et COLEMAN, 1981), de Terre Neuve (GIRARDEAU, 1979, GIRARDEAU et NICOLAS, 1981) et de Nouvelle Calédonie (PRINZHOFFER et al., 1980). Ces hypothèses ont été émises d'après la disposition des structures dans les péridotites que l'on peut classer en deux groupes (OHNENSTETTER, 1980, NICOLAS et VIOLETTE, 1982). Leur signification dépend aussi de la position originelle des ophiolites : les plans structuraux et lithologiques remarquables étaient-ils en position horizontale ou verticale, ou bien suivant les massifs l'un ou l'autre des deux cas est-il possible ? Rappelons que la classification structurale proposée concerne des ophiolites, en particulier celles de Méditerranée orientale dont le magmatisme est différent de celui des dorsales médio-océaniques actuelles. C'est pourquoi une analogie trop étroite entre les mécanismes supposés se dérouler sous les dorsales et ceux déduits d'après l'étude des ophiolites nous apparaît parfois maladroite. Cette restriction sur la signification des modèles est également valable dans le cas où la pétrologie des ophiolites (celles de Méditerranée occidentale) et de la croûte océanique est proche : en effet, les dorsales n'atteignent leur équilibre qu'après 40 MA (ROOTS et al., 1979), une durée fort longue comparée à celle de l'accrétion ophiolitique (voir les discussions à ce propos dans la deuxième partie). Néanmoins, l'étude du massif de Monte Maggiore (OHNENSTETTER, manuscrit, IV.4) qui serait représentatif d'une

écaille de manteau océanique pour des raisons géologiques, minéralogiques et géochimiques, m'a permis de souligner que le rubanement relictuel dans les péridotites est toujours fortement incliné lorsque les derniers dikes de gabbros et de basaltes le recoupent (fig. IV.5). La conservation des textures de diapirs à un niveau superficiel: aux environs de 10 km de profondeur, c.a.d. dans une zone où les lignes de flux auraient déjà dues tourner, laisse envisager que le modèle de MACKENZIE (1967) ou de CANN (1974) peut exister à l'aplomb des dorsales.

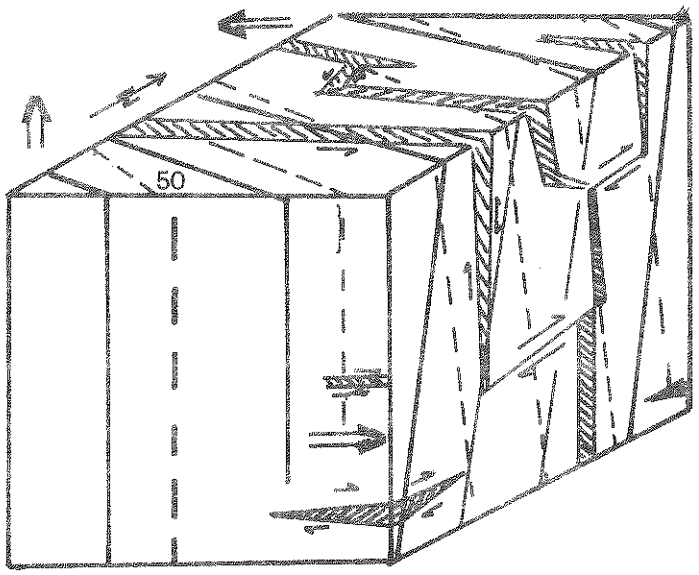


Fig. IV.5 - Relations entre les filons tardifs et les plans lithologiques et structuraux dans la péridotite du Monte Maggiore, Corse.

2.3 - Mise en place des péridotites

Dès lors, compte tenu des dispositifs observés dans les péridotites, on peut concevoir un modèle évolutif des péridotites ophiolitiques. Quel que soit le site géotectonique envisagé, l'ascension des diapirs ophiolitiques a lieu dans des zones d'ouverture liées soit à des systèmes en extension, soit à des systèmes en décrochement. Dans ce dernier cas, l'inclinaison des lignes de flux dépend des mouvements horizontaux et verticaux le long des parois (SALEEBY, 1977, OHNENSTETTER, 1980, I.2). Les diapirs sont de forme arrondie à elliptique (AVE LALLEMANT, 1976). Les couloirs cisaillants qui les bordent ou les recoupent expliquent la dissymétrie de la déformation en fonction des murs et de la superposition des déformations selon les blocs de diapirs juxtaposés (fig. IV.6). Les plis sont isoclinaux, puis plus ouverts ensuite, avec des axes toujours subverticaux. Les roches caractéristiques des diapirs, à gros grain, ont des textures

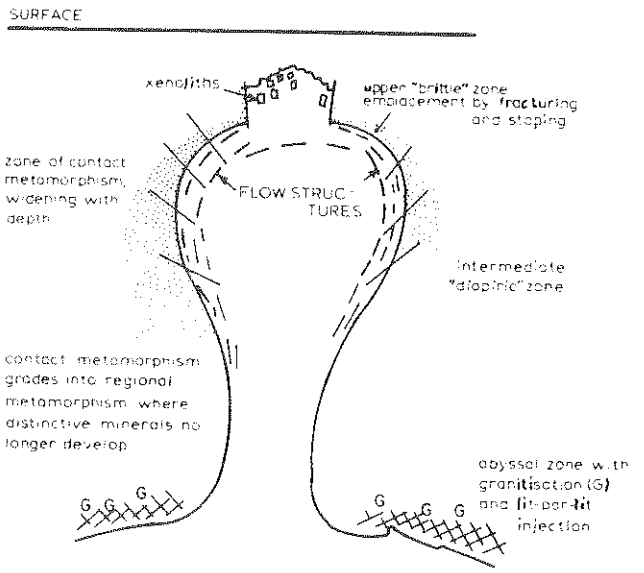


Fig. IV.6 - Aspect d'une intrusion diapirique magmatique ; tiré de FYFE et al. (1978).

protogranulaire et poeciloblastique. Cette dernière texture est rarement préservée dans les complexes de type alpin sauf au Vourinos (ROSS et al., 1980) et en Corse (OHNENSTETTER, manuscrit, IV.4). Elle traduit une diminution des contraintes déviationnelles dans le diapir contemporaine ou non de fusion ou de cristallisation : cristallisations de dunites et chromitites au Vourinos (ROSS et al., 1980) de dunites, harzburgites et lherzolites dans le Cap Corse (OHNENSTETTER, manuscrit, IV.4). Le fluage est alors lié soit à des processus pyrométamorphiques: dissolution-recristallisation sous contraintes (GEORGE, 1977) ou à des recristallisations syntectoniques (AVE LALLEMANT, 1976, ROSS, 1977, ROCCI et al., 1980). On peut souligner que dans le cas où la roche a été en contact avec un liquide, la fabrique tectonique obtenue à hautes températures ne permet pas de juger de l'origine pétrologique de l'échantillon : résidu ou cumulat. Ce n'est que dans les couloirs de cisaillement où lorsque le diapir atteint la surface que les glissements intracristallins dominent (ROSS et al., 1980, Fig. IV.3A).

2.4 - Relations cumulats-péridotites

Si une comparaison entre les massifs de péridotites ophiolitiques permet de définir une chronologie des déformations qui se déroulent sur une période assez brève (40 MA), celle-ci doit également être précisée par rapport à la

cristallisation des cumulats, c.a.d. à la partie superieure de la croûte ophiolitique. Le schéma de la fig. IV.7 résume quelques uns des cas théoriquement envisageables pour les ophiolites.

structuraux d'ophiolites (OHNENSTETTER, 1980, I.2) de même que de nombreux sous-types selon les relations chronologiques entre la rotation du diapir et la cristallisation des cumulats:en profondeur par fluage ou dans des chambres magmatiques plus superficielles.

Dans les ophiolites à structures verticalisées qui à priori traduiraient des extensions plus limitées (cf. deuxième partie),l'angle de dépôt entre cumulats formés dans des chambres à faible profondeur et la foliation du diapir peut être élevé (OHNENSTETTER, 1980, I.2) (fig. IV.8A). Cependant, si les roches grenues cristallisent en profondeur par fluage lorsque le diapir s'élève, les contacts lithologiques et structuraux obtenus seront voisins de celui du diapir (fig.IV.7B) ou tout au moins subverticaux et d'orientation variable (OHNENSTETTER, manuscrit, I.3). Dans les ophiolites à structure verticalisée, la précocité des cumulats est attestée par l'existence d'une déformation intra-mantellaire tardive (S_2) qui les affecte de même que la péridotite associée comme c'est le cas à Chypre (GEORGE, 1978). On peut noter que si l'ensemble est basculé, on obtiendra une disposition conforme au groupe d'ophiolites à structure horizontale (fig. IV.7, cas 3). Tout dépend quand a lieu la rotation des péridotites mantellaires et des premiers cumulats par rapport à la cristallisation des dernières roches grenues (II A et II b de la fig. I.2) (fig. IV.7, cas 4, 5, 6). Si les cumulats se sont déposés une fois la rotation des lignes de flux achevée, l'angle entre la limite des cumulats et la foliation du manteau sera faible (fig. IV.7, cas 4). Pour une rotation moins accentuée (45º) la foliation la plus ancienne sera encore fortement pentée vers la ride comme dans les péridotites du sud de l'Oman (BOUDIER et COLEMAN, 1981) ou à Antalya (REUBER, 1982). Le dispositif final dépend aussi de la subsidence à l'écart de la ride (DEWEY et KIDD, 1977, GIRARDEAU et NICOLAS, 1981) (fig. IV.8C). La rotation plus ou moins prononcée autour d'un axe horizontal est à relier au degré d'expansion qui dépend des conditions tectoniques d'ouverture (cf. deuxième partie). Celle-ci est d'autant plus accusée que l'expansion est plus importante. L'accrétion ophiolitique est alors caracterisée par des séries cumulatives puissantes qui ont été en équilibre avec des liquides moins appauvris que ceux des ophiolites à structures verticalisées. Des extensions de faible amplitude empêcheraient un renouvellement de l'asthénosphère au voisinage du site d'accrétion ce qui expliquerait l'appauvrissement progressif du manteau (cf. première partie).

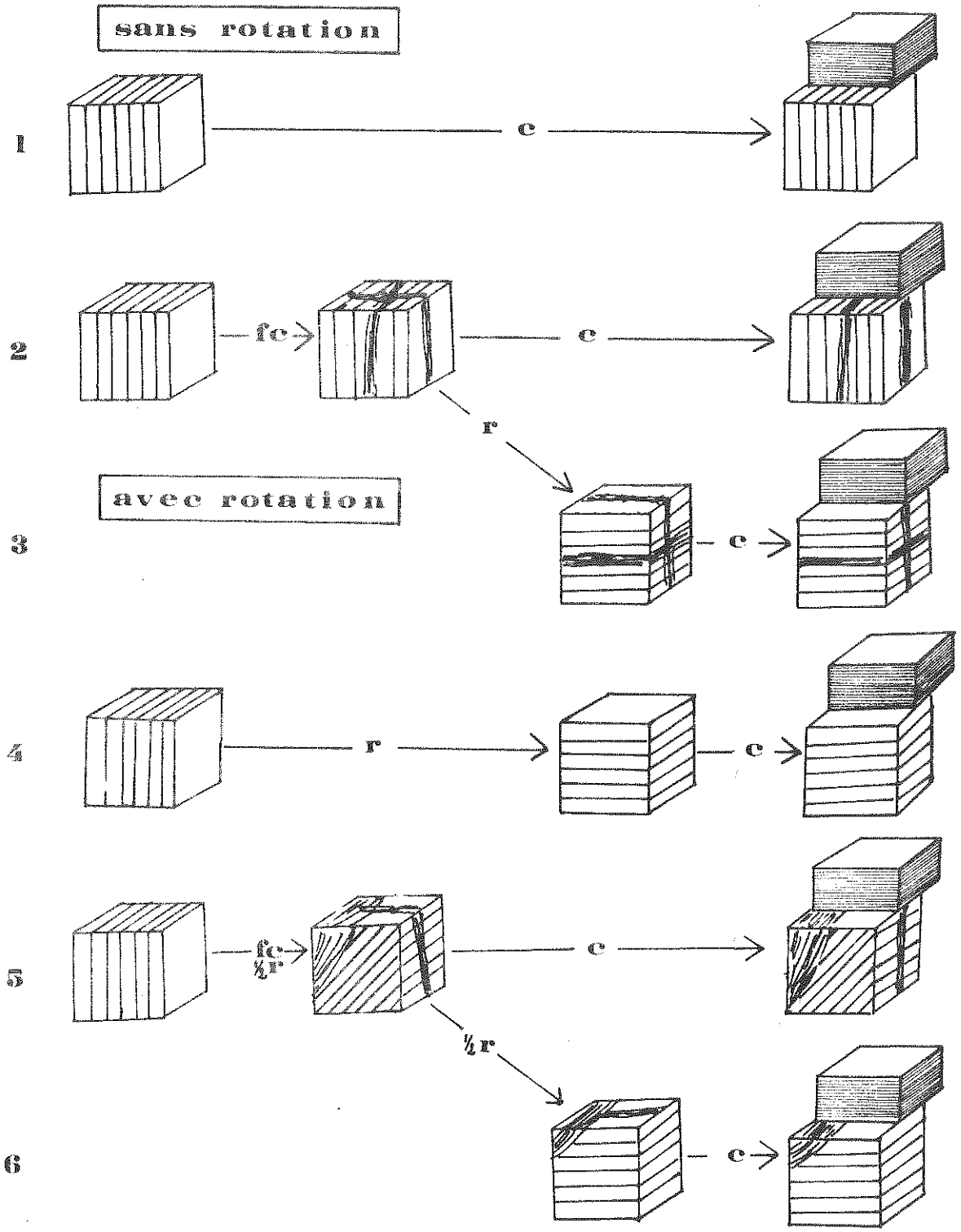


Fig. IV.7 - Relations angulaires entre le rubanement des cumulats et la foliation des péridotites en fonction du moment de la cristallisation des cumulats par rapport à une éventuelle rotation du diapir mantellaire. Le tracé dans les péridotites et les cumulats correspond respectivement à la trace du plan de fluage et aux plans de variation lithologique dans les cumulats. c, correspond à la formation de cumulats dans des chambres magmatiques superficielles. fc est mis pour la formation de cumulat par fluage magmatique. r, indique la rotation du diapir.

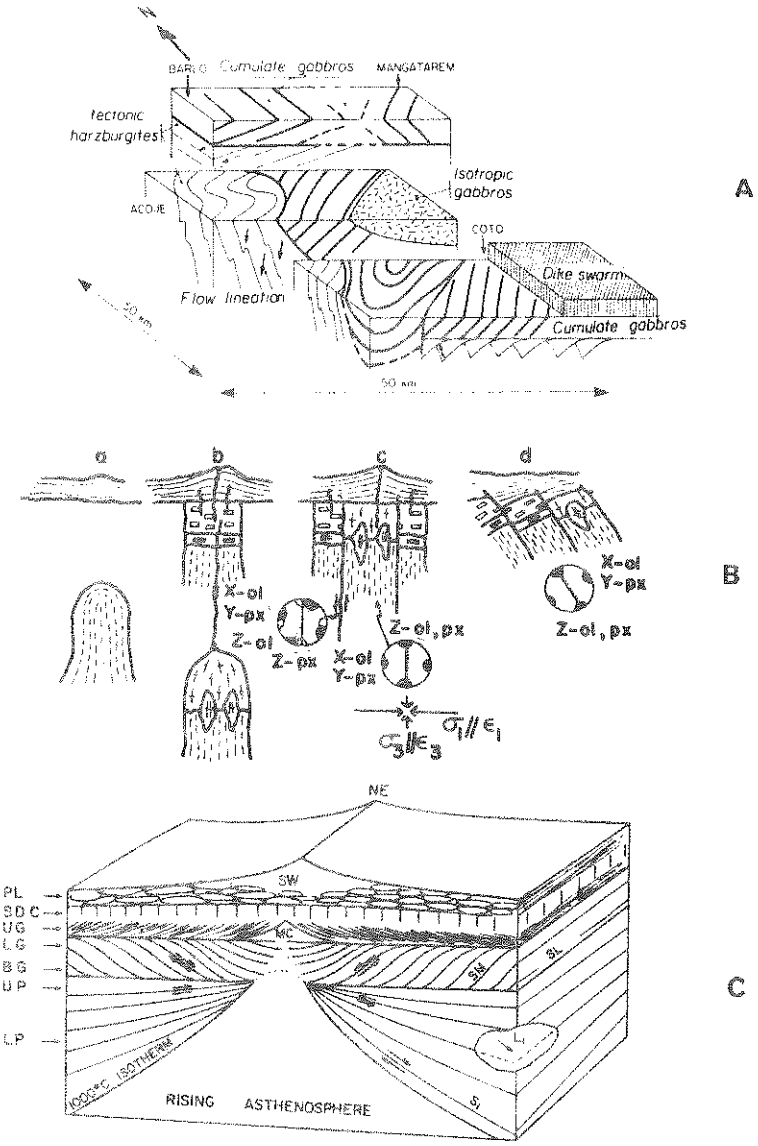


Fig. IV.8 - Relation du rubanement des cumulats avec la foliation des péridotites. A, tiré de NICOLAS et VIOLETTE (1982). Noter la linéation sub-verticale dans les péridotites et le rubanement des cumulats à un angle élevé par rapport à la foliation, principalement dans le secteur d'Acoje. B, tiré de AVE LALLEMANT (1976). Noter la foliation commune dans les péridotites diapiriques et les premiers cumulats formés (c) ; l'angle est au contraire accentué entre les derniers cumulats sub-horizontaux et la foliation précédente (b et c). L'ensemble est ensuite basculé lors de la mise en place tectonique (d). C, tiré de GIRARDEAU et NICOLAS (1981). La disposition des lignes de flux dans le manteau et du plan d'accumulation magmatique SM forme une arête de poisson par suite de l'écoulement asthénosphérique et de la subsidence des cumulats. PL, pillow lavas ; SDC, complexe filonien ; UG, gabbro supérieur ; LG, gabbro inférieur ; BG, gabbro de base ; UP, péridotite supérieure ; LP, péridotite inférieure ; MG, chambre magmatique, SL, contact lithologique ; S_1 foliation ; L_1 linéation.

III - SIGNIFICATION DES FILONS DANS LES PERIDOTITES

Les chapitres précédents ont discuté de la déformation progressive des péridotites lors de leur emplacement dans des zones de plus en plus superficielles tout en tenant compte de leur relation géométrique vis à vis des chambres magmatiques. Au-dessous de ces chambres, les péridotites sont recoupées par des filons de roches grenues interprétées soit comme des produits de fusion partielle synchrone de la déformation mantellaire soit comme des cumulats liés à la formation des chambres superficielles. Etant donné leur texture, ces filons doivent être considérés comme des extraits formés à partir d'un liquide circulant vers la surface. L'ouverture de ces filons, selon les hypothèses pétrogénétiques envisagées peut être reliée au système de contraintes régnant en profondeur lors de la fusion, ou plus près de la surface lors du remplissage de la chambre magmatique. Quel que soit le mécanisme envisagé, les résultats de l'analyse tectonique seront voisins et compatibles avec la géométrie d'une zone d'accrétion.

1 - Les mécanismes d'ouverture

Avec les filons, on aborde aussi, du point de vue mécanique, les problèmes de rupture fragile à grande profondeur, en particulier lorsque la péridotite se déforme plastiquement. Dans les massifs de type alpins on replace généralement les plans de migration des liquides par rapport aux marqueurs de la déformation qui permettent de déduire approximativement la position des contraintes dans le manteau. On peut aussi, uniquement à partir de la position des fractures, envisager le système de contrainte favorable à leur création. Cette dernière approche s'apparente à celle employée lors de l'étude des zones volcaniques superficielles. Lorsqu'il y a éruption volcanique, on peut concevoir que la pression de magma crée les ouvertures ou que le magma s'injecte dans des fractures liées au champ de contraintes dans les plaques. Dans ce dernier cas, la montée des magmas dans des zones distensives nécessite que σ_3 soit horizontal (FYFE et al., 1978) ou que les contraintes maxima σ_1 et σ_2 soient en position verticale (SHAW, 1980). σ_3 vertical permet d'expliquer l'alignement des chaînes volcaniques parallèlement à σ_1. De la même façon, en modélisant la mise en place primaire des ophiolites (JOHNENSTETTER M., 1980, I.2), on a conclu que σ_1 ou σ_2 devait être fortement incliné pour favoriser l'accrétion ophiolitique.

Dans le manteau soumis à une déformation plastique, la seule possibilité pour faire entrer les roches dans le domaine élastique est de diminuer la pression lithostatique. Deux possibilités peuvent être alors envisagées pour y arri-

ver. En faisant remonter la péridotite dans des zones superficielles, sans qu'il y ait rotation importante des contraintes, la pression lithostatique décroît. L'ouverture des filons et, partant de là, leur cristallisation sont tardives par rapport à la déformation plastique et à la fusion partielle qui ont affecté l'ensemble de la péridotite.

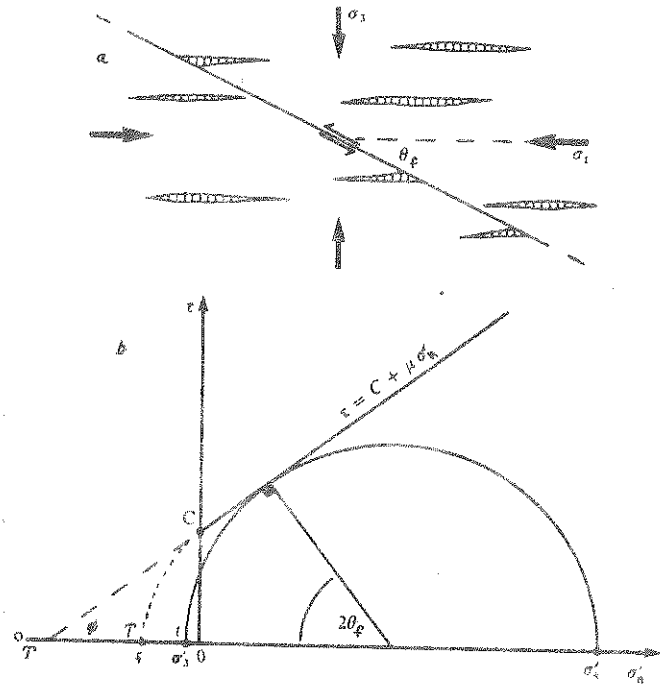


Fig. IV.9 - Cercle de Mohr et position des fractures correspondantes ; a, relations sur le terrain des fractures en extension et des fractures cisaillantes simultanées ; b, déplacement du cercle de Mohr lors de la fracturation hydraulique ; tiré de SIBSON (1981).

Cette possibilité doit être plus fréquente qu'on ne l'imagine étant donné l'abondance des filons à proximité des chambres magmatiques relictuelles. Pour réduire artificiellement les contraintes, il suffit d'augmenter la pression de fluide que l'on peut assimiler à la pression de magma (YODER, 1976). La contrainte effective qui en résulte (ANDERSON, 1936, HUBERT et RUBEY, 1959) : contrainte normale -pression de fluide, provoque un déplacement (fig. IV.9) du cercle de Mohr vers la gauche et les contraintes normales négatives. Si la pression de fluide est forte les fractures s'ouvrent en domaine de traction : elles sont alors perpendiculaires à σ_3 qui est plus ou moins confondu avec la direction d'allongement des minéraux de la péridotite. Si les contraintes normales

sont fortes (dans le cas où la pression de fluide est faible), les fractures s'ouvrent en compression et sont inclinées d'un angle inférieur à 45º par rapport à la contrainte maximum (fig. IV.9). Entre ces deux types d'ouverture, il existe des fractures cisaillantes mais avec une composante extensive non négligeable. Elles sont disposées à 20 ou 30º de la contrainte maximum et sont souvent en échelon. Elles sont symétriques par rapport à σ_1 lors d'une déformation irrotationnelle. Ainsi des variations dans la pression de fluide peuvent entraîner des ouvertures quasiment simultanées d'orientation variable (SHAW, 1980). La fracturation hydraulique ne serait pas facilitée dans des terrains en compression alors qu'elle peut exister à de grandes profondeurs dans des zones en extension (SIBSON, 1981). Le rôle de la pression fluide sur l'ouverture des fractures dans le manteau a été souligné par YODER (1976) KOIDE et BHATTACHARJI (1977) et SHAW (1980). Elle a été appliquée aux complexes alpins et ophiolitiques par JACKSON (1979), REUBER (1982) et REUBER et al. (1982). Dans le cas des complexes de type alpin (JACKSON, 1979), les filons peuvent apparaître à proximité du plan d'extension puis tourner pour se paralléliser avec le plan de fluage. Des variations dans la pression de fluide rendraient compte des différentes inclinaisons par rap-

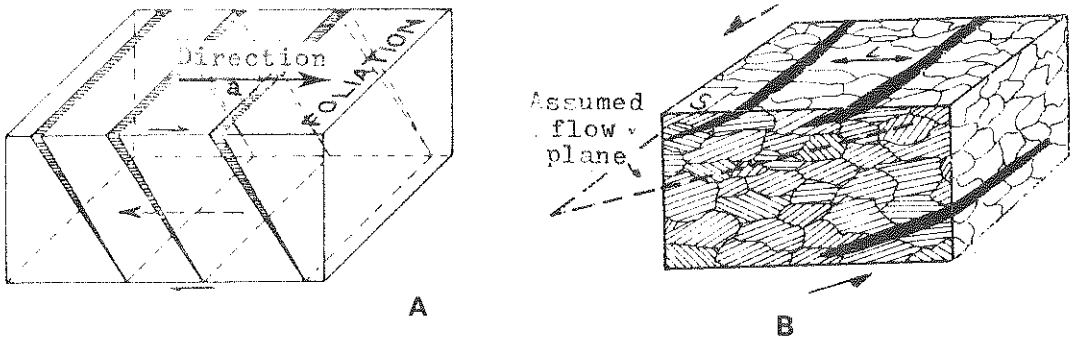


Fig. IV.10 - Position schématique des lentilles feldspathiques. A, dans des plans de tension résultant du cisaillement dans la foliation S_1 suivant la direction a, d'après BOUDIER (1972). B, coalescence des lentilles de feldspath dans le plan de fluage tiré de l'orientation des plans de glissement dans l'olivine et l'enstatite ; d'après BOUDIER et NICOLAS (1977).

port à la contrainte maximum pour expliquer la position de filons qui ont d'abord été placés (fig. IV.10) dans des fentes de tension à 45º du plan de foliation lors du cisaillement de la péridotite (BOUDIER, 1972, 1976) puis à proximité du plan de fluage (BOUDIER et NICOLAS, 1977). Un même mécanisme dans les pérido-

tites ophiolitiques expliquerait les variations du gisement des filons par rapport à l'axe de la dorsale (REUBER et al., 1982).

L'ouverture des fractures inclinées sur σ_1 et/ou symétriques par rapport à σ_1 comme dans le cas des failles cisaillantes avec une composante en extension, peut être expliquée par analogie avec les bandes de Lüder dans les métaux (BURG et HARRIS,1982) : des fentes de tension en position conjuguée sub-perpendiculaire à σ_3 peuvent être obtenues dans des zones préalablement soumises à une déformation plastique (fig. IV.11).

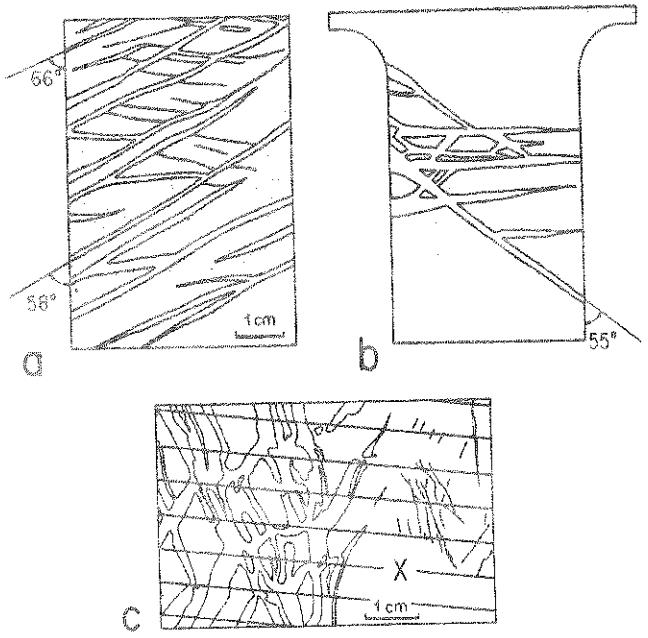


Fig. IV.11 - Position des fentes en extension ; a, fracture oblique à sub-perpendiculaire à la direction d'extension dans des ardoises ; b, bandes de Lüder dans une plaque d'acier ; la direction d'extension est verticale ; c, fractures conjuguée en extension de même position que les bandes de Lüder dans des ardoises aux Deux Alpes ; les fibres à l'intérieur sont parallèles à la direction d'extension. D'après BURG et HARRIS (1982).

2 - L'analyse géométrique des zones de fusion et la migration des magmas

Toutes ces hypothèses permettent de comprendre l'ouverture des filons dont la morphologie témoigne du passage du magma déjà rassemblé. Qu'en est-il de la position des liquides en faible proportion par rapport à la péridotite, par exemple lors de la fusion et qui doivent enrober les minéraux ? La plupart des études expérimentales montre que les liquides même en faible concentration ($\simeq$ 1 %)

se placent préférentiellement à la frontière des grains sub-parallèle à la contrainte maximum appliquée (fig. IV.12A) (AVE LALLEMANT et CARTER, 1970, BOUDIER et NICOLAS, 1977). Les concentrations de magma peuvent définir un nouveau plan de variation lithologique (fig. IV.12B) obtenu lors de processus de dissolution-recristallisation (GEORGE, 1977).

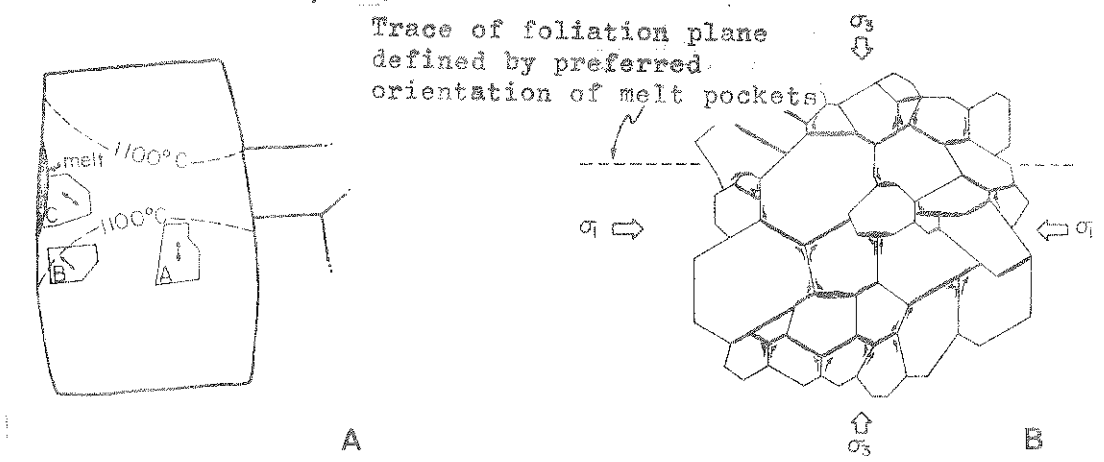


Fig. IV.12 - Concentration de magma au cours de fusion partielle. A, lors d'études expérimentales à 1200° ; la contrainte maximum est verticale. D'après AVE LALLEMANT et CARTER, 1970. B, position des poches de magma lorsque les cristaux sont soumis à des contraintes, d'après GEORGE (1977).

Si on fait intervenir en plus la pression de fluide créée par le magma issu de la fusion (YODER, 1976), on peut imaginer que dans le manteau, il existe un réseau de fentes en extension sub-verticales relayées par des plans plus ou moins inclinés (fig. IV.13A) permettant la propagation des magmas (KOIDE et BHATTACHARJI, 1977). Les fractures cisaillantes (fig. IV.13B) qui ont une composante extensionnelle non négligeable servent de connection entre les dikes et poches de magma (SHAW, 1980). D'après des études expérimentales, la circulation de magma autour des minéraux peut avoir lieu en même temps que la déformation plastique de la péridotite dont la fabrique ne change pas (AVE LALLEMANT et CARTER, 1970, BOUDIER et NICOLAS, 1977). Soulignons que ce mécanisme ne peut juger du processus pétrologique associé à la migration des magmas : fusion ou fractionnement, de même qu'il ne permet pas d'apprécier les écarts de temps entre la déformation plastique et la mise en place des filons. D'autres considérations doivent être prises en compte pour préciser l'évolution pétrologique et structurale du manteau (OHNENSTETTER M., manuscrit, IV.4).

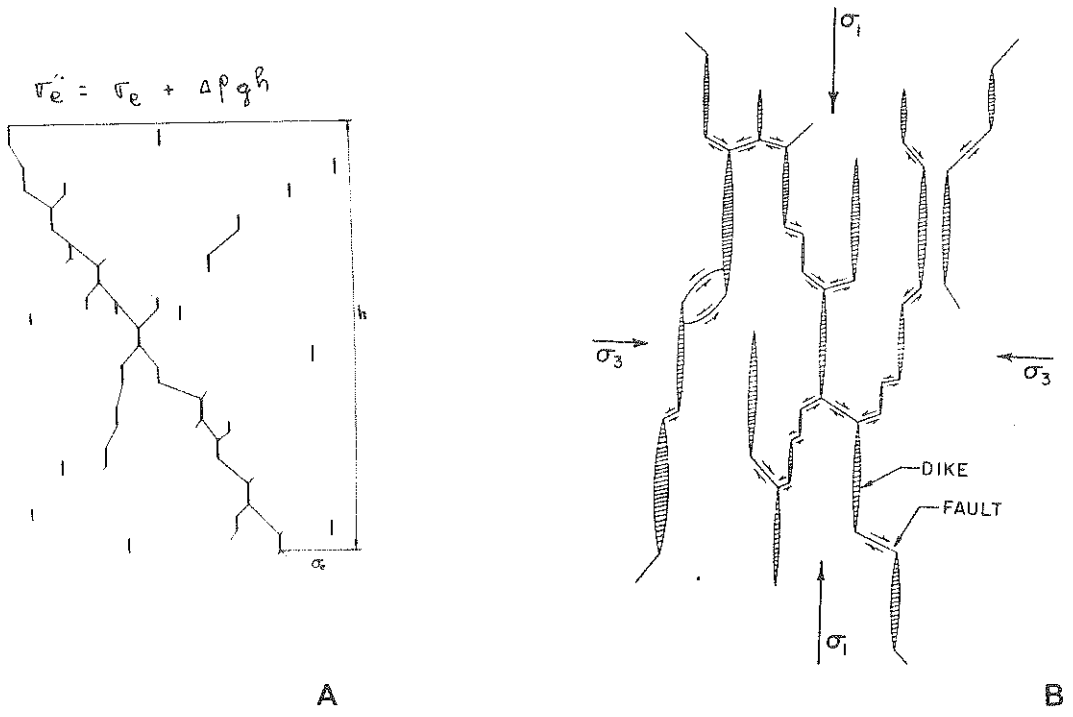


Fig. IV.13 - Migration des magmas dans le manteau. A, propagation des fractures remplies de fluides ; d'après KOIDE et BHATTACHARJI (1977). σ_e' et σ_e sont les contraintes effectives en haut et en bas de la colonne considérée. B, position des dikes et des connections le long de failles cisaillantes inclinées en fonction des contraintes effectives ; d'après SHAW (1980).

3 - La migration des magmas lors de l'ascension des diapirs

Les chapitres précédents ont permis de dégager que les textures protogranulaires et poeciloblastiques étaient compatibles avec des contraintes différentielles faibles. On peut donc supposer qu'au sein du diapir, si la pression de fluide varie, des ouvertures dans de nombreuses directions sont possibles au joint des grains. Ces fractures permettent au cours de la fusion et de la migration des magmas, des rééquilibrations entre toutes les phases, ce qui explique l'homogénéité minéralogique des péridotites résiduelles. Par ailleurs, pour des contraintes différentielles faibles, le magma créé lors de la fusion n'aura pas tendance à s'échapper favorisant le départ de magma en un large volume (batch melting). YODER (1976) a souligné qu'en augmentant les contraintes différentielles, la collection et le départ des liquides se trouvaient facilités, favorisant la fusion fractionnée (fractional melting). Le magma primaire peut occuper jusqu'à 15 % du volume de la roche (MAALOE, 1981) avant que le manteau ne soit perméable (Fig. IV.4) et que les magmas ne s'échappent. Les produits issus de la fusion partielle disparaissent complètement. Il ne reste alors que les traces des plans de migration des magmas.

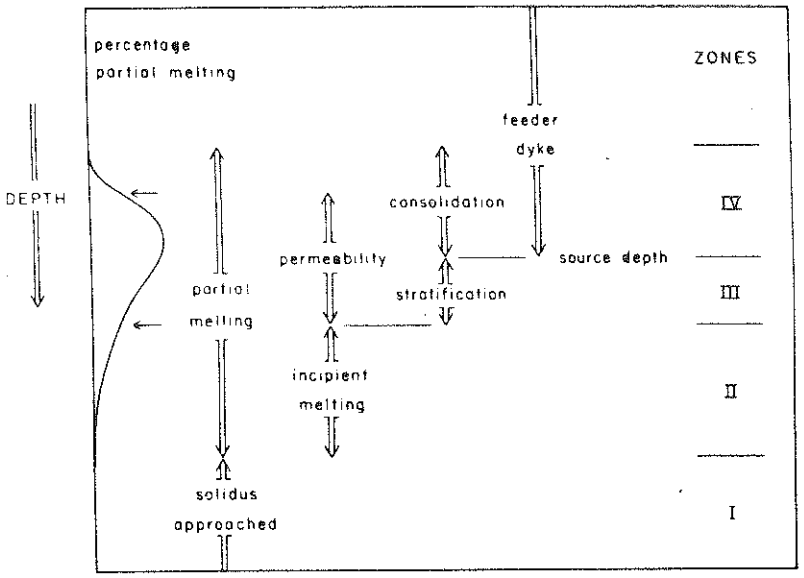


Fig. IV. 14 - Succession des différentes zones dans le manteau en fonction du degré de fusion partielle et de la perméabilité du manteau; d'après MAALOE (1981).

Les chapitres précédents ont également montré qu'au cours du temps les contraintes différentielles augmentaient vers la surface. Ceci conduit à l'obtention des filons mieux réglés en direction (voisin de σ_1) au contact plus ou moins flous. A grande profondeur, si la température de la péridotite est proche du solidus, il y aura réaction entre liquide et solide conduisant à des rééquilibrations suivies du dépôt de la phase principale au liquidus (Fig. IV. 15). A des pressions lithostatiques inférieures, la température de la péridotite sera également abaissée et la cristallisation des magmas conduira à des assemblages polycristallins pendant que la déformation de la peridotite se poursuivra avec l'acquisition des textures porphyroclastiques. On voit que tout au long de l'évolution la géométrie de l'ensemble est $\pm$ constante même si les processus pétrologiques varient. Cette évolution peut correspondre à la mise en place des diapirs de péridotites quel que soit l'environnement géotectonique. Cependant en fonction du site, la composition des magmas et résidus seront différents.

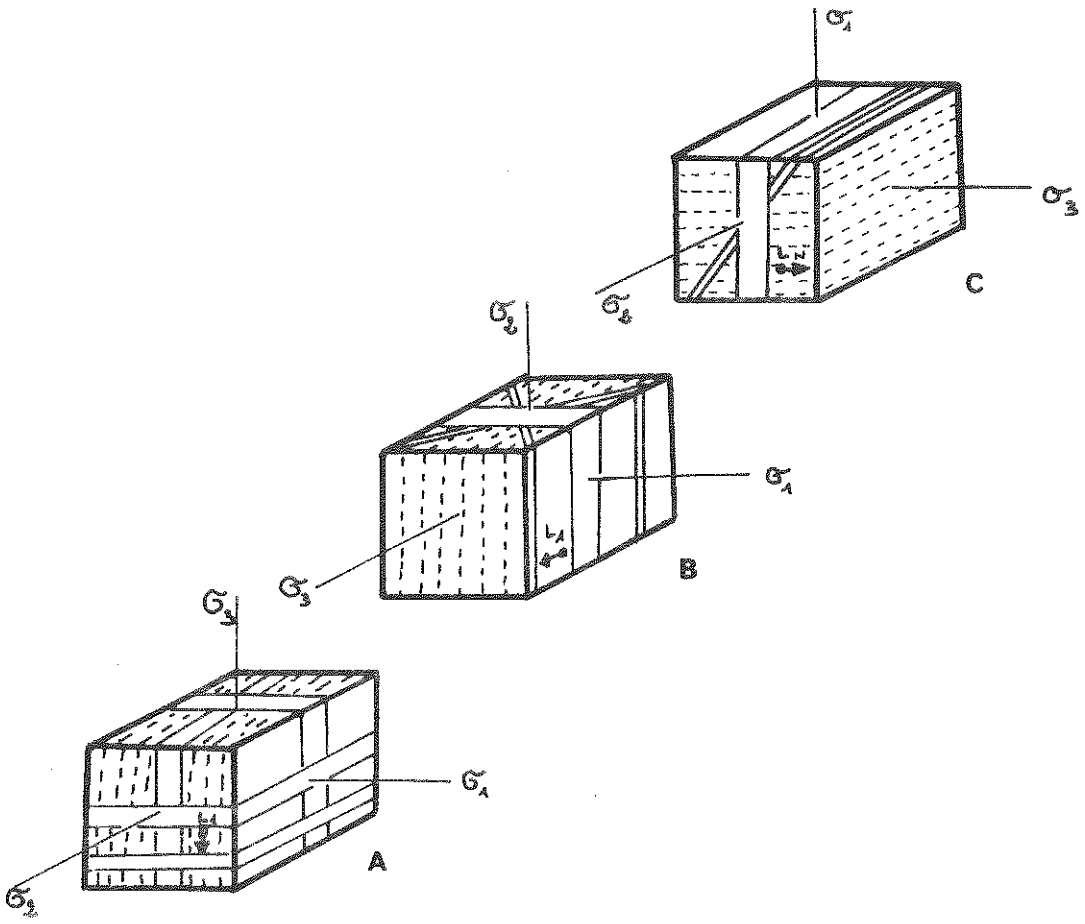


Fig. IV.15 - Position des liquides par rapport aux plans marqueurs de la déformation et aux contraintes depuis le diapir jusqu'aux zones superficielles. A, lors de la fusion et/ou de fractionnements de hautes pressions sous de faibles contraintes différentielles. B, lors de cristallisations de hautes pressions et températures élevées lorsque les contraintes différentielles augmentent. Des réactions aux épontes existent. C, après la rotation du diapir dans un système en extension. Le stade B pourrait exister encore à proximité de la surface mais sans éponte réactionnelle.

IV - LES CARACTERISTIQUES PETROGRAPHIQUES DES PERIDOTITES OPHIOLITIQUES

1 - L'environnement géotectonique

On ne connaît pas de péridotite à grenat antérieure au métamorphisme régional dans les complexes ophiolitiques ou dans les environnements océaniques actuels à proximité des dorsales. Les résidus mantellaires sont donc équilibrés

à des pressions relativement basses, inférieures à 15 Kb, dans le faciès à spinel et celui à plagioclase (O'HARA, 1967). Un diagramme P-T de stabilité des faciès permet de concevoir que si les diapirs viennent d'une zone profonde, aux environs de 100 km, les péridotites ont été équilibrées soit dans le faciès à grenat, soit dans celui des pyroxènes alumineux (fig. IV.16). Des gradients géothermiques élevés supérieurs à 20º/km comme dans les rides médio-océaniques (O'HARA et al., 1971) favoriseraient l'existence du faciès à pyroxène alumineux aux dépens de celui du faciès à grenat.

Il est maintenant reconnu deux sites géotectoniques majeurs pour la création des ophiolites. Ils sont représentés sur la figure IV.17 : les rides médio-océaniques issues du développement de rifts intra-continentaux et les zones d'arc et de bassins marginaux en arrière des zones de subduction. Dans ce dernier cas, on peut constater que des diapirs plus ou moins appauvris géochimiquement et minéralogiquement (harzburgites plus ou moins riches en Cpx) et qui sont moins denses qu'une lherzolite fertile (OXBURG et PARMENTIER, 1978) peuvent coexister. La montée de diapir n'est pas forcément liée à une zone de subduction : si la convection mantellaire ne peut exister par suite d'une expansion limitée le long de fractures profondes, le manteau qui n'est pas renouvelé s'appauvrit très rapidement avec le temps. Cette hypothèse rend compte de la diminution plus ou moins régulière des éléments incompatibles dans les liquides (troisième partie).

Dans les dorsales médio-océaniques, l'expansion, de vitesse variable, conduit à imaginer une succession d'intrusions diapiritiques qui délivrent en surface des magmas de composition constante avec le temps.

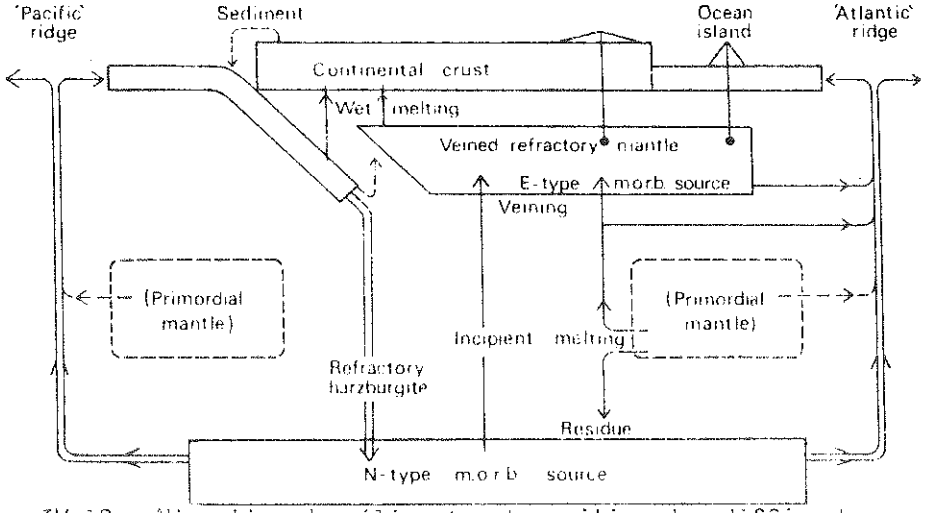


Fig. IV.18 - Migration des éléments et position des différentes sources mantelliques, d'après TARNEY et al. (1980).

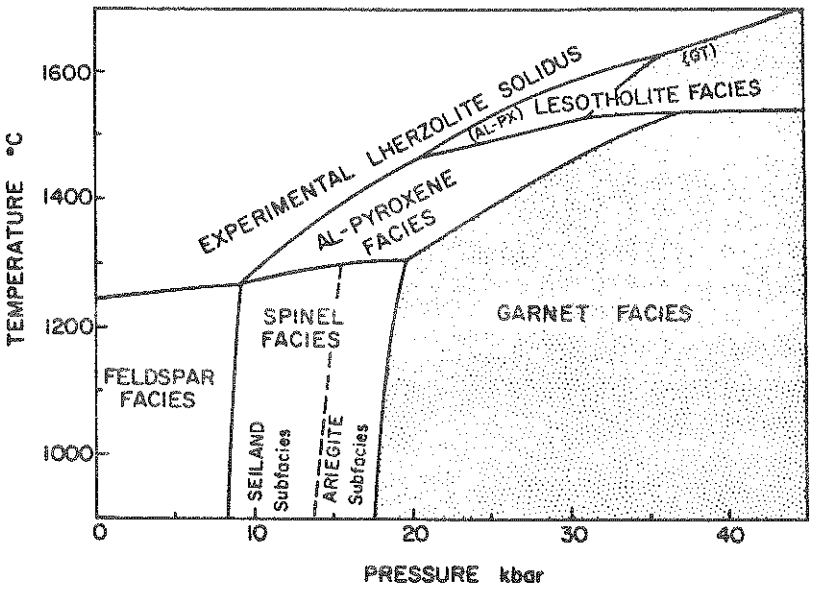


Fig. IV.16 - Les faciès minéralogiques des péridotites naturelles, d'après O'HARA (1971 et 1975), tiré de MERCIER (1977).

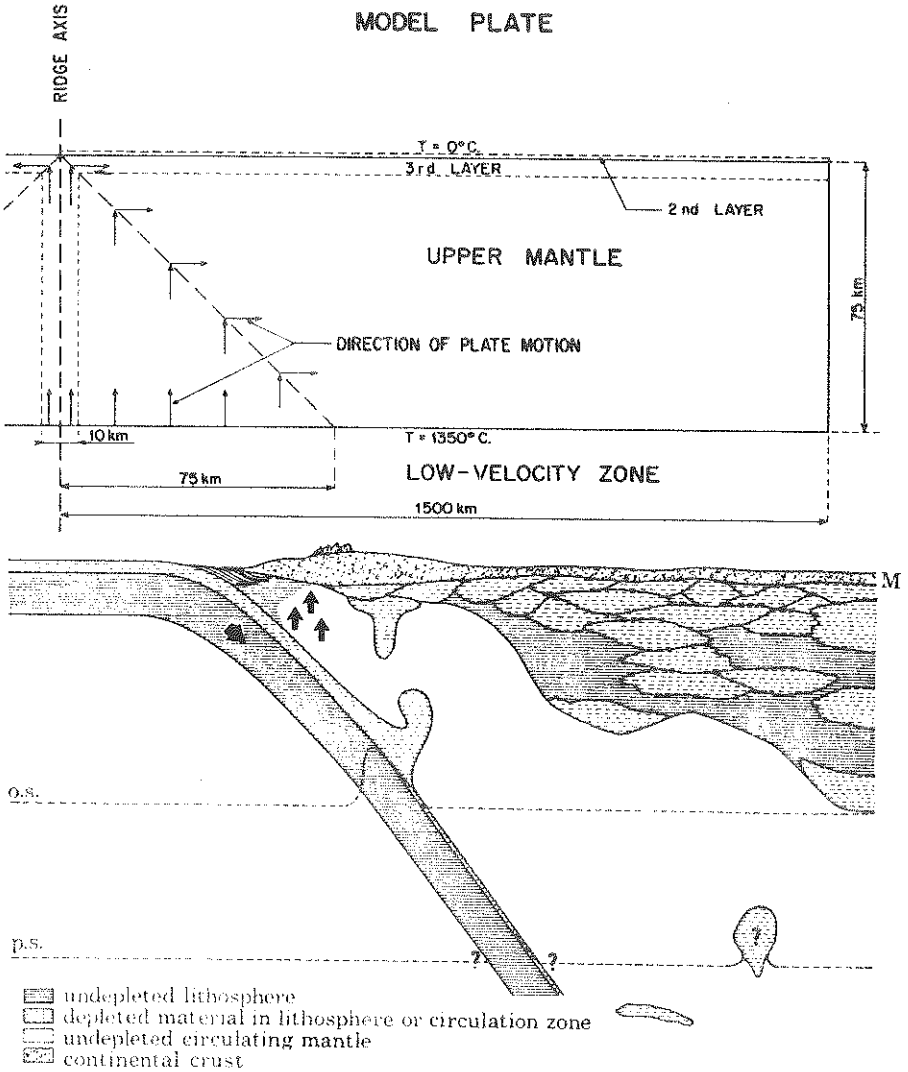


Fig. IV.17 - Structure des plaques ; A, au voisinage de la ride, d'après BOTTINGA (1974) ; B, en arrière d'une zone de subduction ; tiré d'OXBURG et PARMENTIER (1977).

Des phases essentielles en proportion plus ou moins variable et des fusions plus ou moins hydratées permettent d'expliquer des divergences entre le magmatisme des dorsales et celui des tholéïtes d'arc immature (voir la compilation de WYLLIE et al. 1981). Cependant dans chacun des sites géotectoniques mentionnés, on envisage souvent l'existence d'hétérogénéités mantellaires d'après la répartition des éléments incompatibles et des isotopes dans les liquides observés. En particulier, dans les rides médio-océaniques, le manteau ne serait pas homogène parallèllement à l'axe de la ride (SCHILLING et al. 1980). Des fluides et/ou des liquides contamineraient localement les péridotites résiduelles avant qu'elles ne subissent d'autres fusions partielles : le manteau veiné réfractaire (Fig. IV. 18) serait proche de la surface (TARNEY et al. 1980). Près des dorsales, il constitue la couche 4a de BUTTINGA (1974) qui est un lieu de cristallisation situé sous la couche 3. La mise en évidence d'hétérogénéités minéralogiques ou géochimiques directement à partir des échantillons issus du manteau est plus délicate. Celles-ci n'ont été signalées que dans les péridotites de certains nodules des basaltes alcalins et kimberlites, encore que dans ces cas, il soit difficile de préciser à quelles profondeurs elles interviennent (FRANCIS 1976 , LLOYD et BAILEY, 1975, BOETTCHER et al., 1977, WOOD, 1979, BOETTCHER et O'NEIL, 1980, WASS et al. 1980).

2 - Les caractéristiques pétrographiques des péridotites

Dans les complexes de type alpin des hétérogénéités liées à des circulations de liquides enrichis en incompatibles ont été rarement mises en évidence. Elles sont signalées à propos des péridotites de St-Paul (MELSON et al. 1972). Dans les péridotites du Monte Maggiore, Corse, la présence de phases accessoires comme apatite et kaersutite et la modification de la composition des phases majeures résulteraient de la migration le long de certaines fractures de liquides riches en alcalins et incompatibles. (OHNENSTETTER M., manuscrit ; IV. 4). Ce métasomatisme conduit à la juxtaposition de lherzolites géochimiquement appauvries et de harzburgites enrichies, un bilan qui s'oppose à celui classique de "parent fertile et résidu appauvri" comme l'a aussi souligné DAWSON et al. (1980) à propos des péridotites des kimberlites.

D'une manière générale, la comparaison des péridotites des complexes ophiolitiques porte principalement sur la proportion de Cpx et la composition des phases, en particulier sur la valeur du rapport Mg/Mg+Fe qui doit refléter

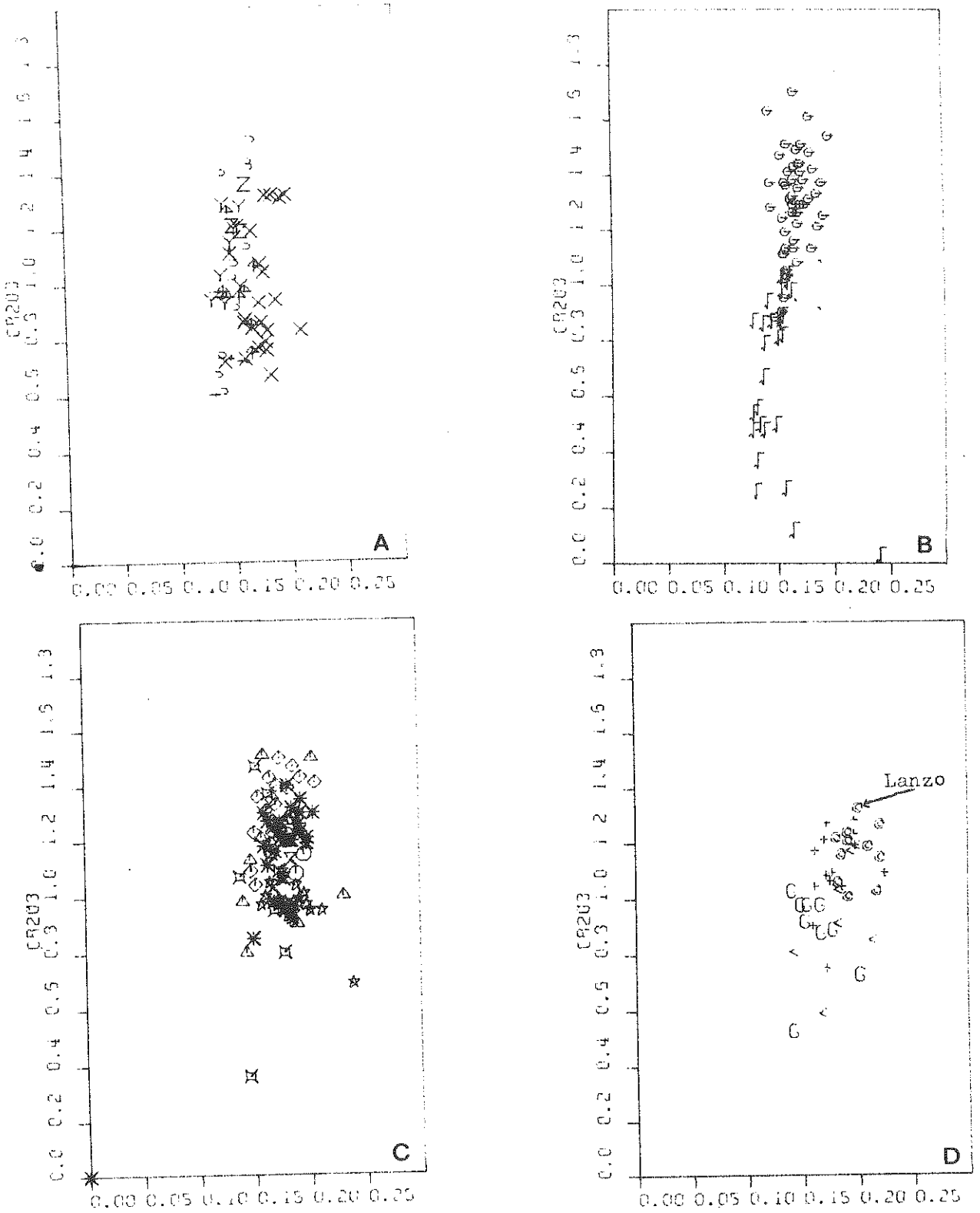


Fig. IV.19 - Variation de Cr_2O_3 en fonction de FeO/FeO+MgO. A, péridotites océaniques à gros grain ; B, péridotites océaniques à texture porphyroclastique. C, péridotites du Monte Maggiore. D, autres péridotites de Corse et de Lanzo (OHNENSTETTER, en préparation).

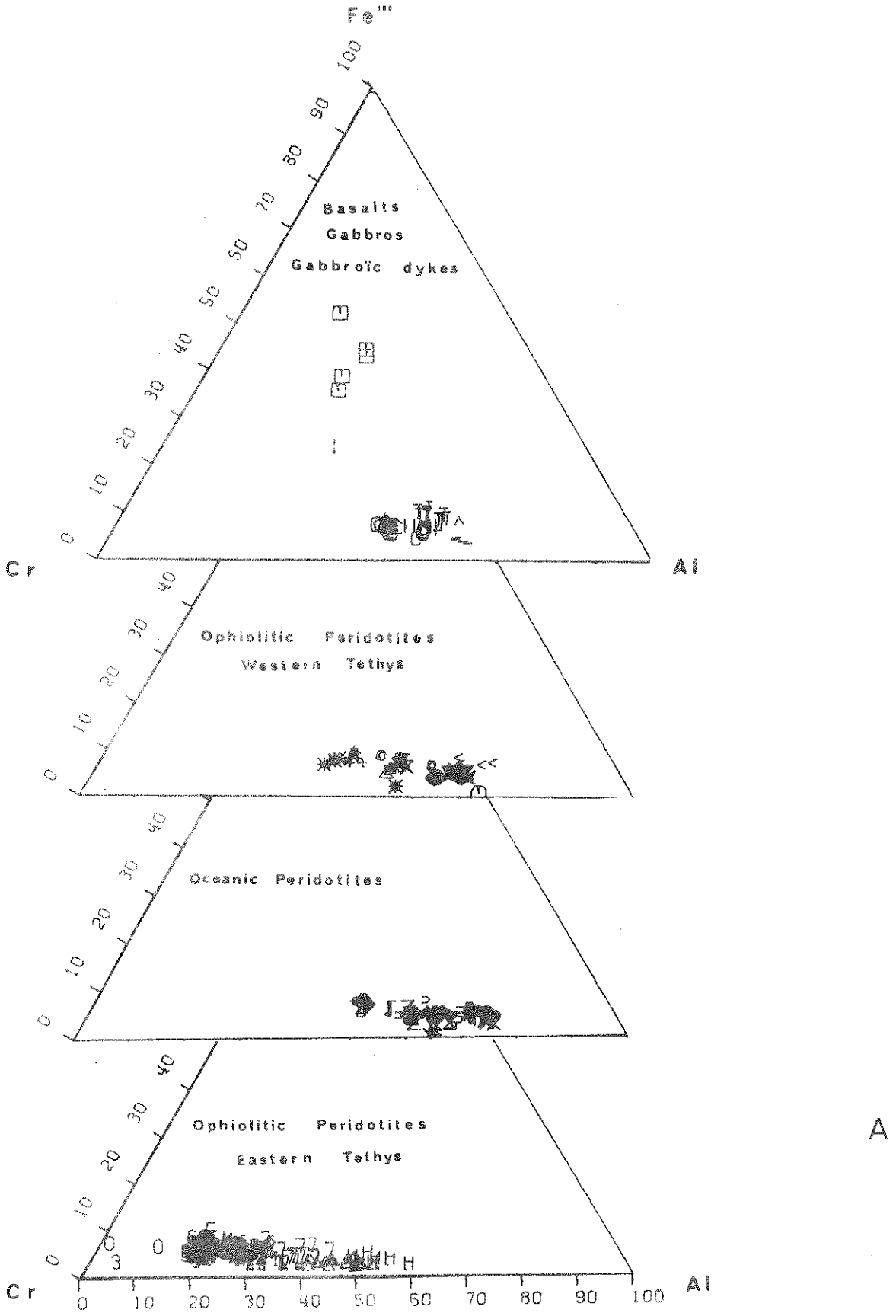
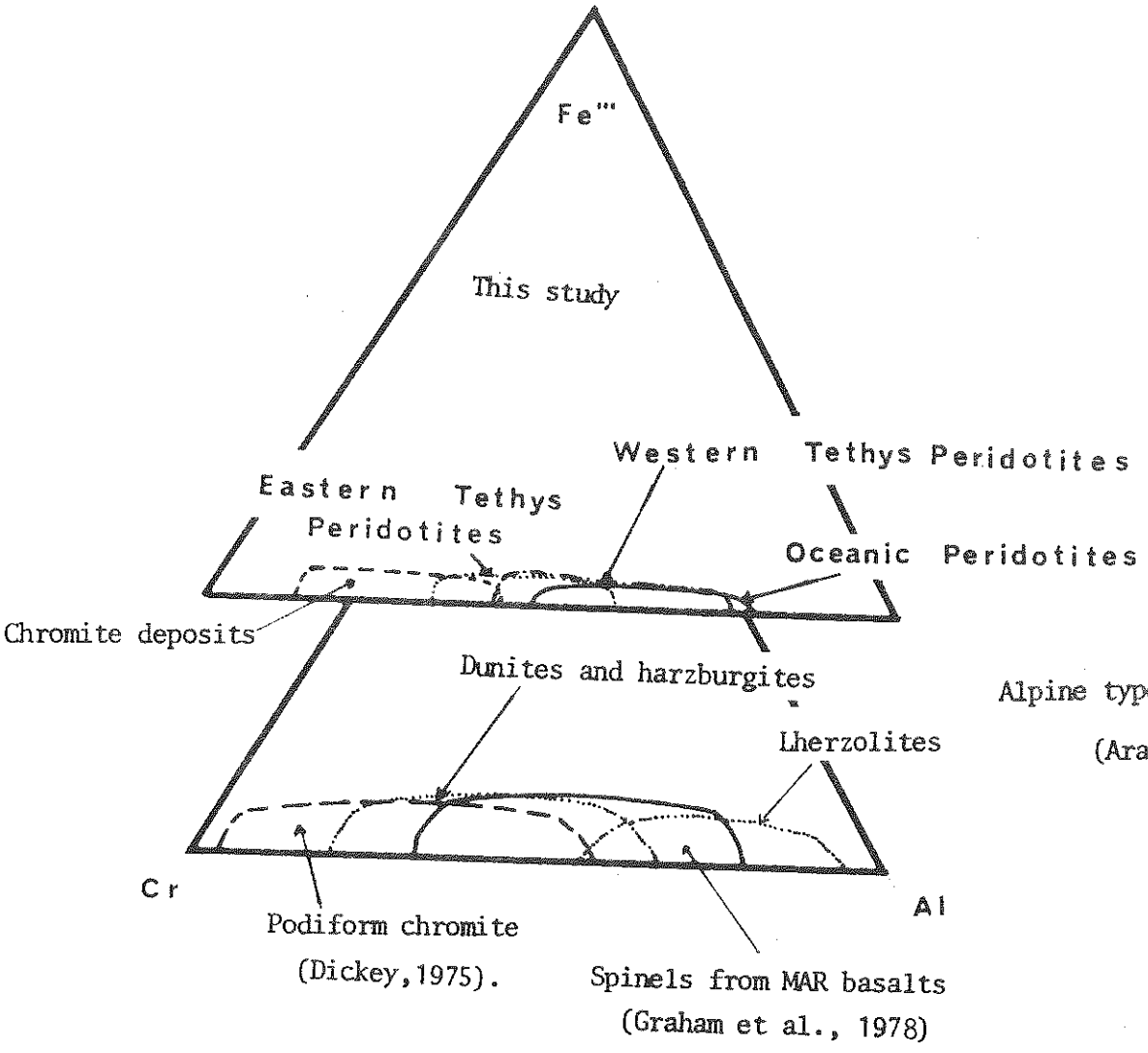


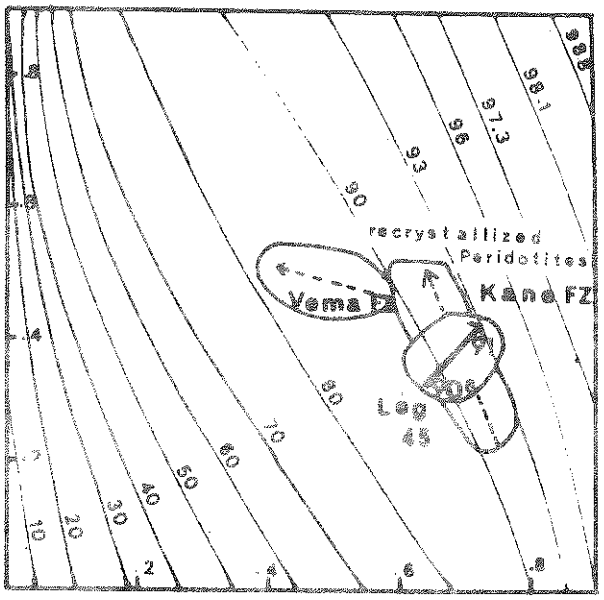
Fig. IV.20 - Diagramme triangulaire Cr, Al, Fe^{3+} dans les spinelles. A, la répartition des points ; B, la position des principales enveloppes comparée à quelques autres tirées de la littérature (DHNENSTETTER, en préparation).



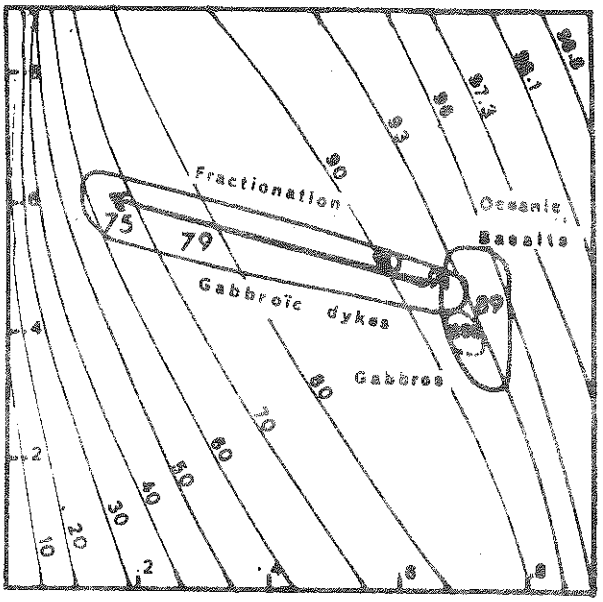
Alpine type peridotites

(Arai and Fujii, 1978)

B

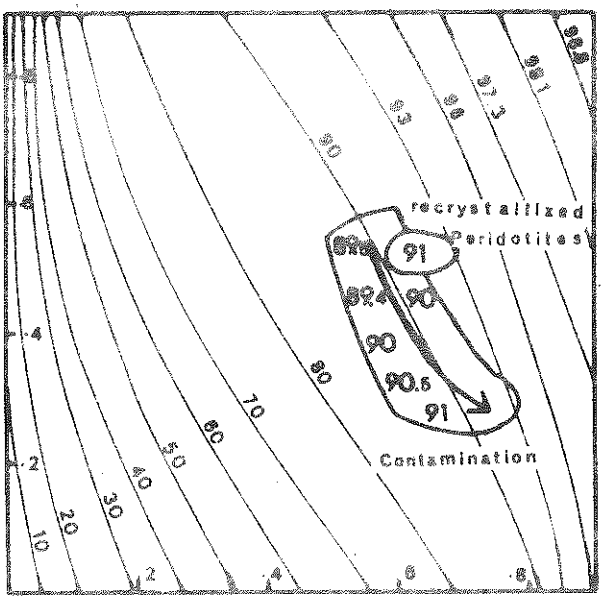


Oceanic Peridotites

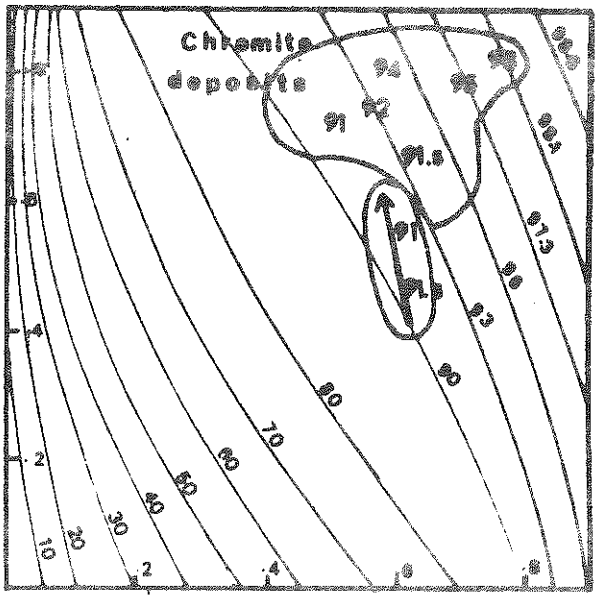


Corsican Gabbros and Oceanic Basalts

Cr/Al+Cr



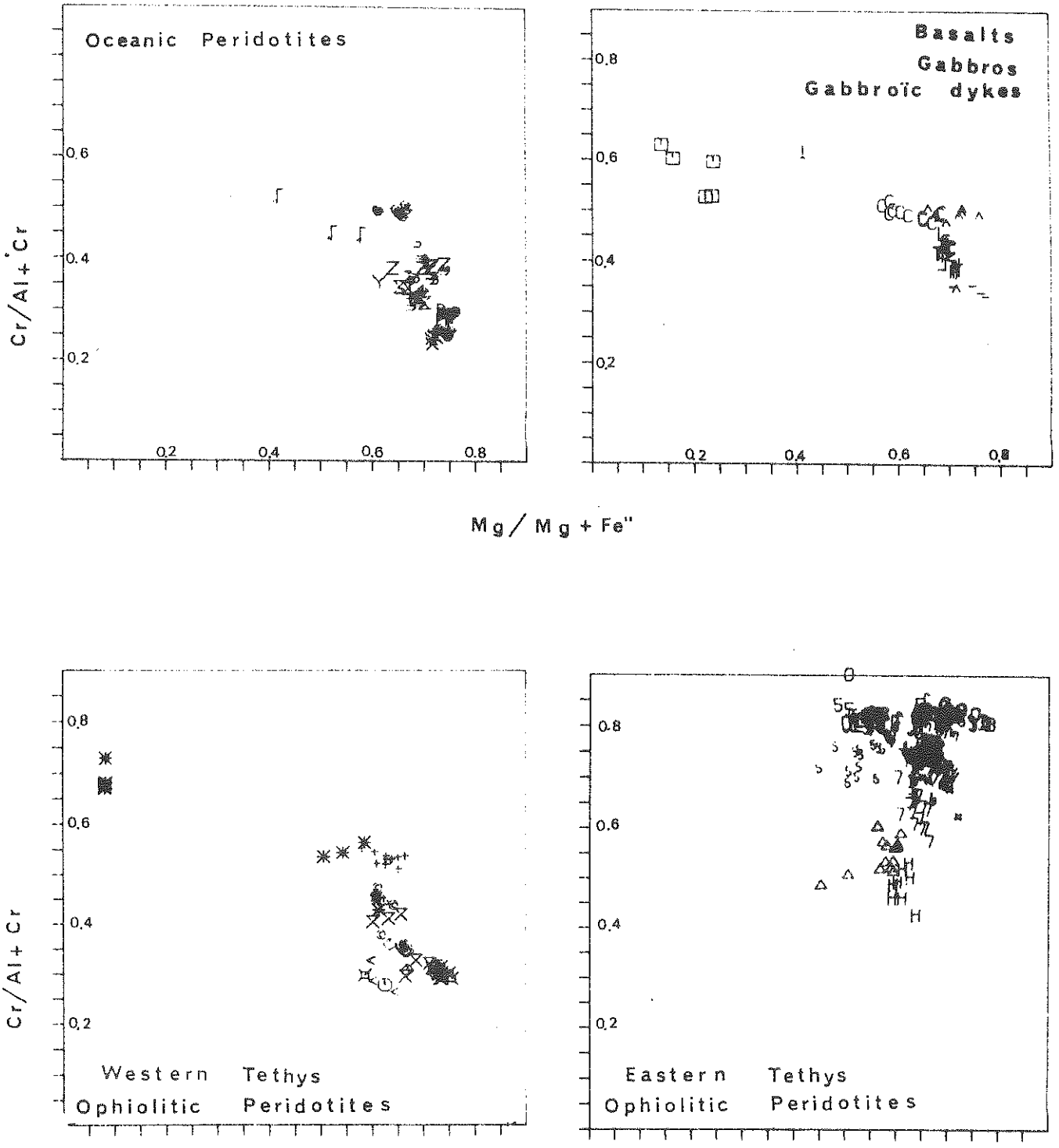
Western Tethys Peridotites

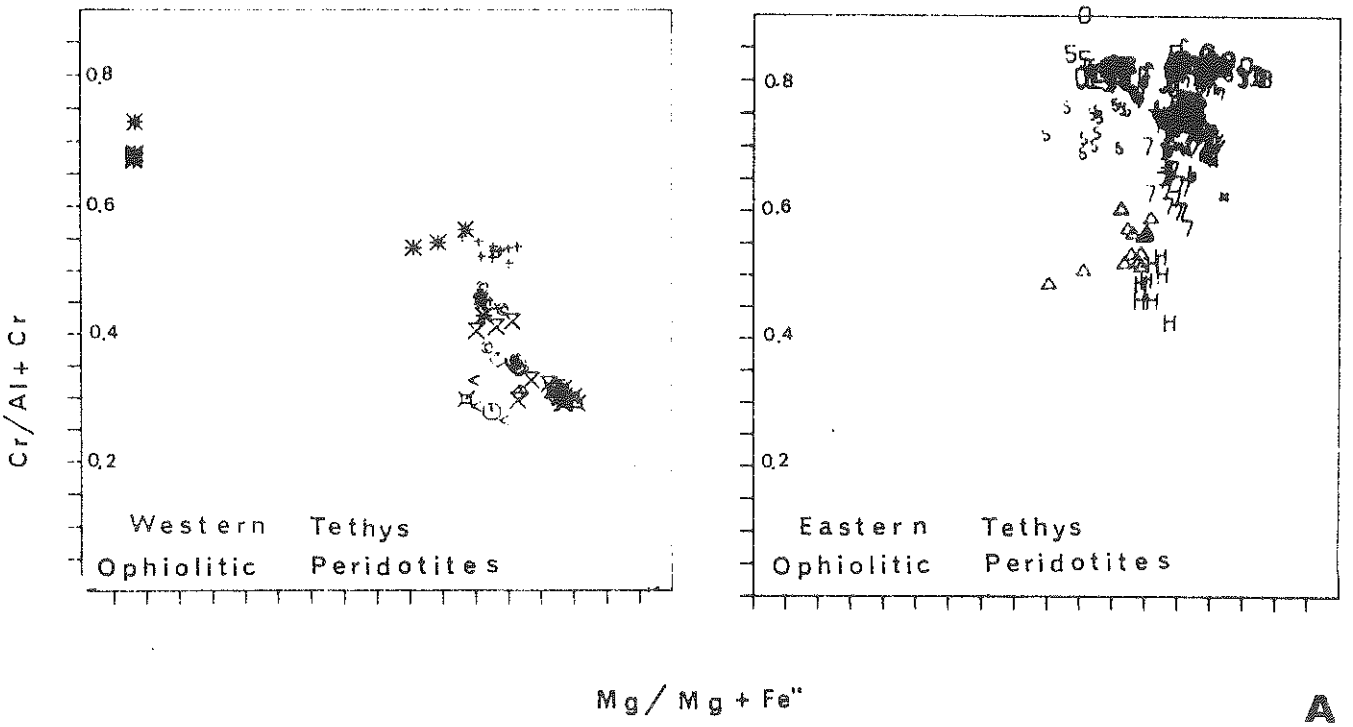


Eastern Tethys Peridotites

B

Mg / Mg + Fe"





A

Fig. IV.21 - Evolution des rapports Cr/Cr + Al en fonction de Mg/Mg+Fe
A, disposition des points (OHNENSTETTER, en préparation)
B, position des enveloppes des points projetés par rapport aux lignes équipotentielles des olivines tirées d'IRVINE (1965). Les flèches indiquent les principaux processus pétrogénétiques reconnus à partir de l'évolution de la composition des spinelles.

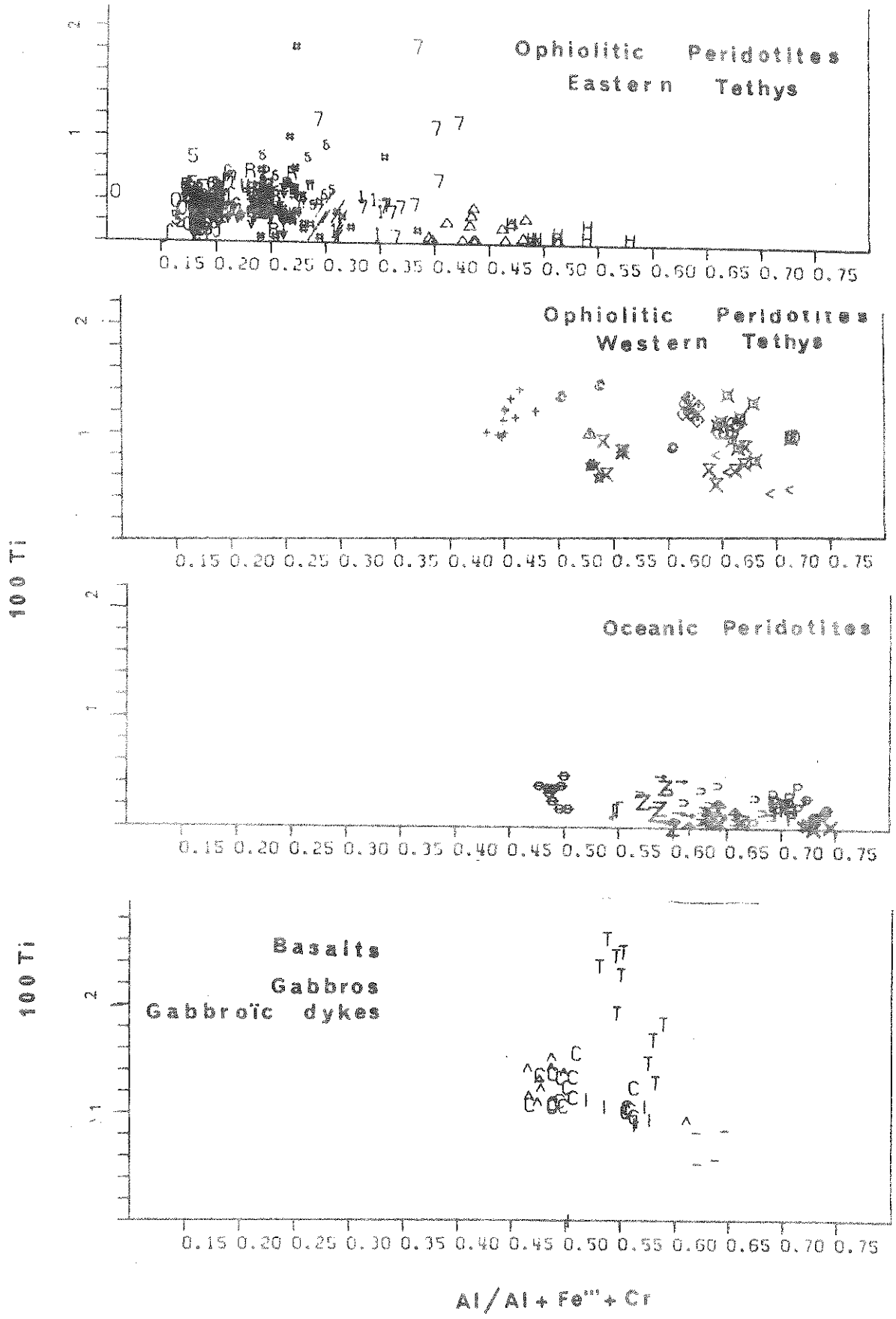


Fig. IV.22 - Variation de Ti en fonction de Al/AL + Fe^{3+} + Cr dans les spinelles. Les spinelles des basaltes appartiennent aux laves océaniques. Ils ne sont distingués ni des gabbros en filons sécants sur les péridotites du Monte Maggiore, ni des gabbros corses issus des chambres magmatiques sous-crustales (DINNEWSTETTER, en préparation).

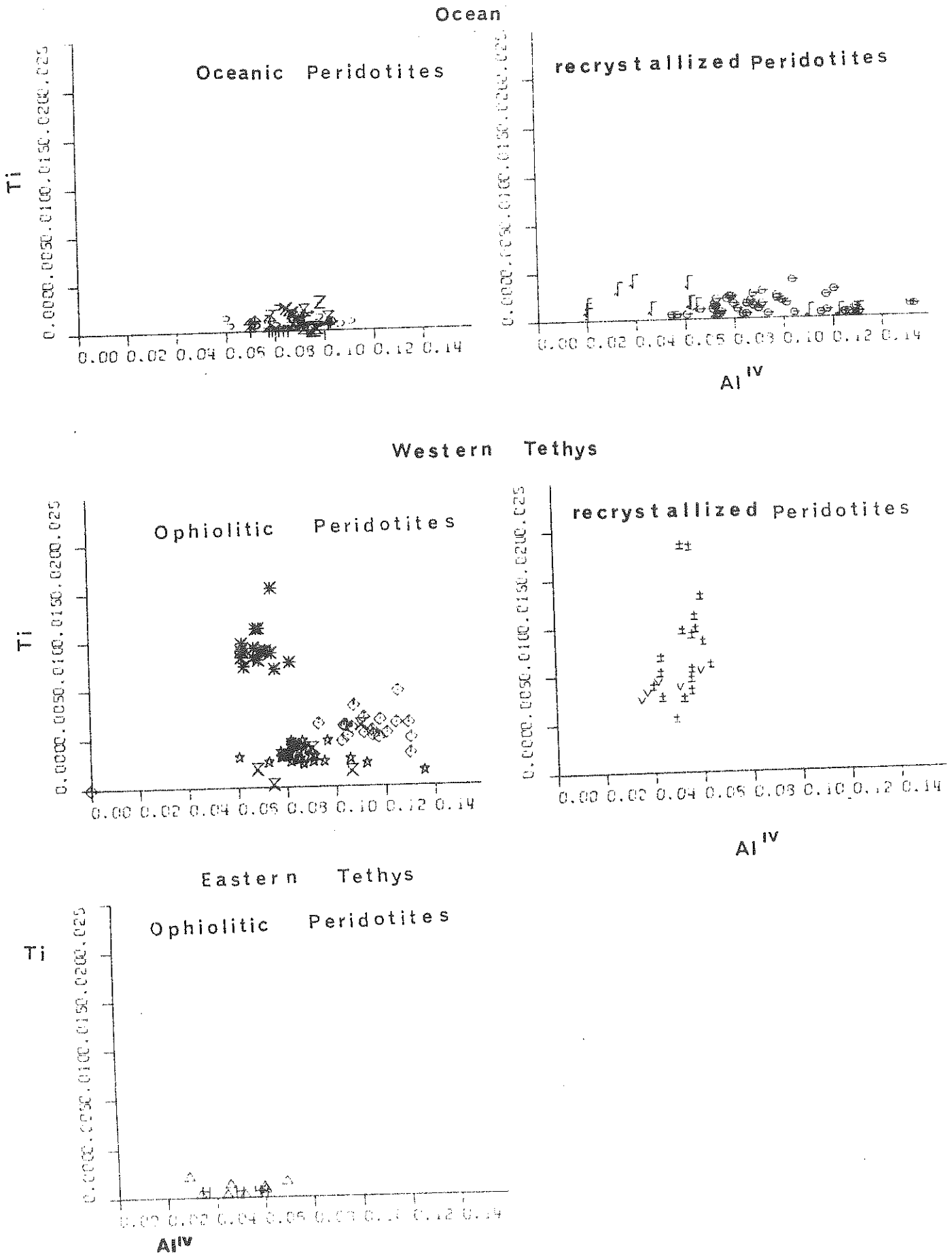


Fig. IV.23 - Distribution de Ti et Al^{IV} dans les Opx des péridotites (OHNENSTETTER, en préparation).

un appauvrissement généralement lié aux fusions partielles. Sur la figure IV.19, on peut considérer les variations de Cr_2O_3 en fonction de FeO/FeO+MgO dans les Cpx des péridotites ophiolitiques et océaniques. Cr_2O_3 et MgO doivent augmenter lors de la fusion partielle. On peut constater que leur variation quoique faible montre une augmentation en Mg tandis que Cr varie peu ou diminue. Dans les minéraux, la répartition des éléments permet de constater que les péridotites de type pétrographique II sont en général plus réfractaires que celles de Méditerranée occidentale qui ont de nombreuses affinités avec les péridotites océaniques. Ol, Opx, Cpx et Sp sont plus magnésiens dans les péridotites de Méditérranée orientale que dans celles de Méditérranée occidentale (Fig. IV.19, 20 et 21). De la même façon leurs spinelles sont plus chromifères. Bien que les spinelles de Méditérranée occidentale et des océans occupent des domaines voisins dans la plupart des diagrammes, les premiers sont cependant plus riches en Ti que les seconds et s'apparentent à ceux des basaltes océaniques (Fig. IV.22). L'enrichissement en Ti est aussi très net dans les orthopyroxènes (Fig. IV.23). Cette distinction basée sur Ti peut être liée au degré d'ouverture et donc à l'âge du bassin (voir aussi la deuxième partie).

Le diagramme Na-Cr dans les Cpx des péridotites (KORNPROBST et al., 1981, IV.1) souligne aussi une plus grande richesse en alcalins dont Na dans les péridotites de Méditérranée occidentale (Fig. IV.24). La plupart des diagrammes mettant en évidence la répartition des éléments dans les minéraux essentiels conduisent à distinguer les péridotites ophiolitiques entre elles. On peut également tenter de dégager les principaux processus pétrologiques responsables de leur composition : changement de faciès, fusion partielle, fractionnement et contamination. Dans la plupart des diagrammes, les variations minéralogiques liées aux fusions partielles sont peu nettes par suite de l'équilibration finale entre liquide primaire et minéraux (Fig. IV.25). Le caractère résiduel des péridotites est néanmoins évident. Comme la plupart des derniers processus pétrologiques se produisent alors que P et T décroissent régulièrement, les migrations entraînent généralement des diminutions conjointes de Cr et Al dans les pyroxènes tandis que Mg/Mg+Fe varie peu. En séparant les faciès selon leur degré de recristallisation, on peut constater que celle-ci entraîne une diminution de Cr et Al dans les pyroxènes qui sont respectivement redistribués dans les spinelles

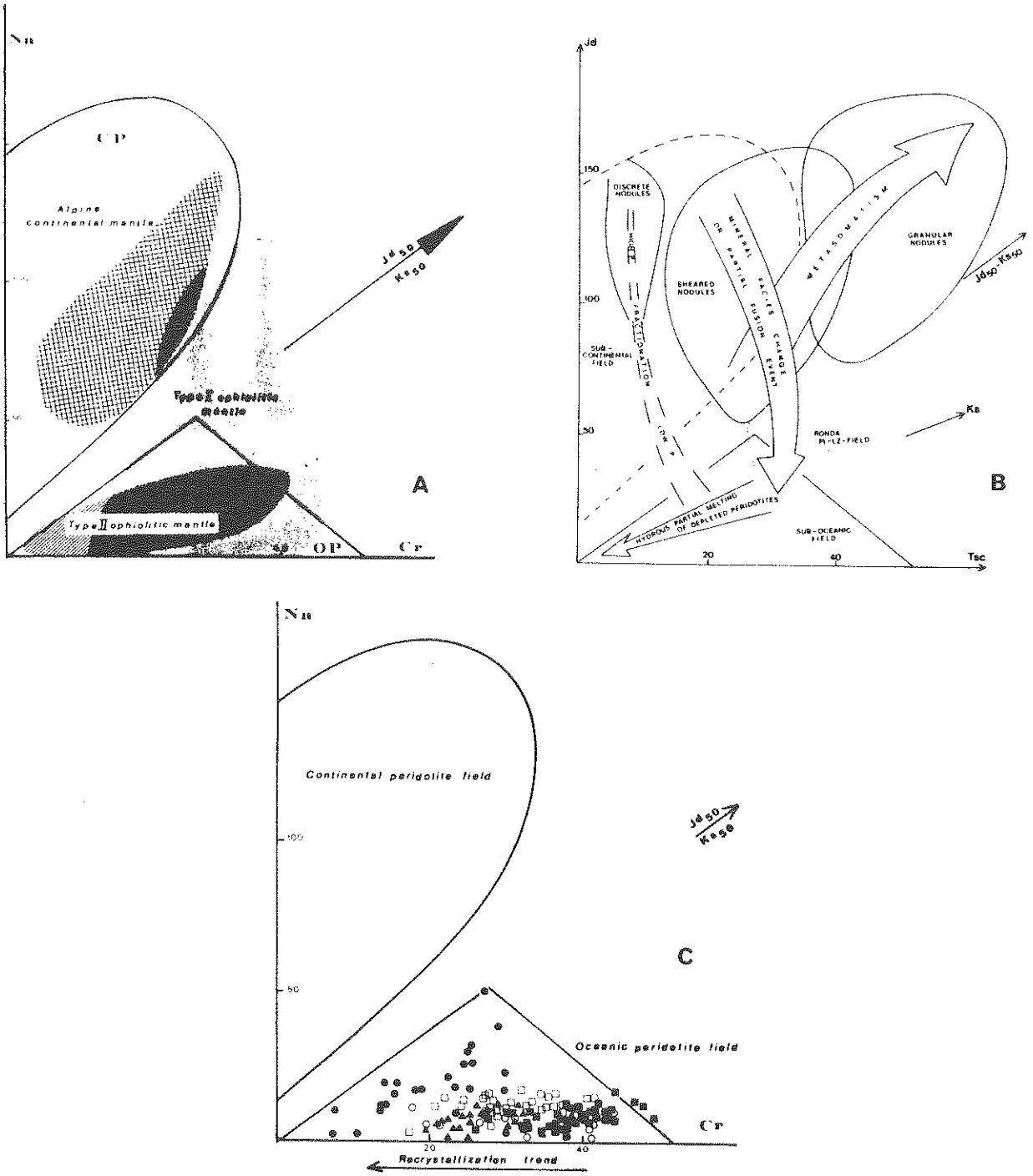
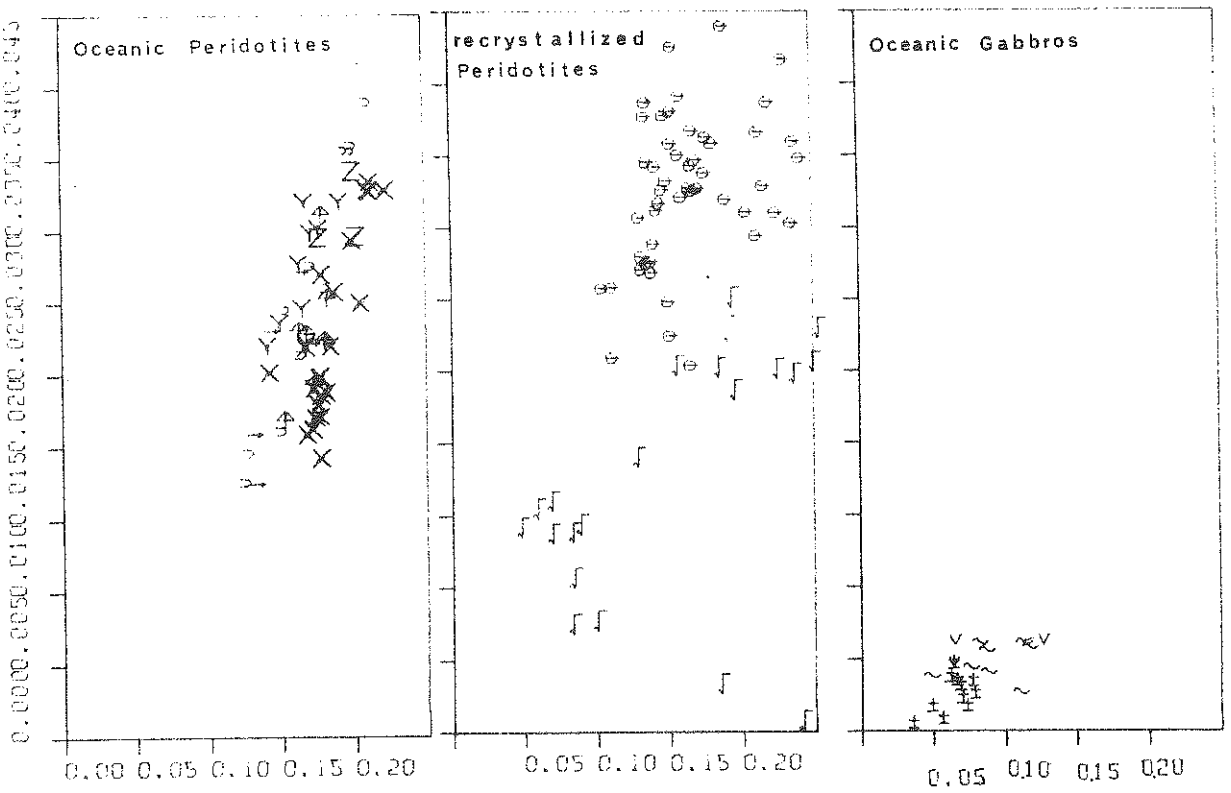


Fig. IV.24 - Diagramme Na-Cr pour les Cpx des péridotites, d'après KORNPROBST et al. 1981. A, principaux domaines ; CP, péridotites sub-continentales ; OP, péridotites sub-océaniques ; B, position des principales évolutions pétrogénétiques. C, variation de Cr lors de recristallisation, tiré de KORNPROBST et al., 1982.

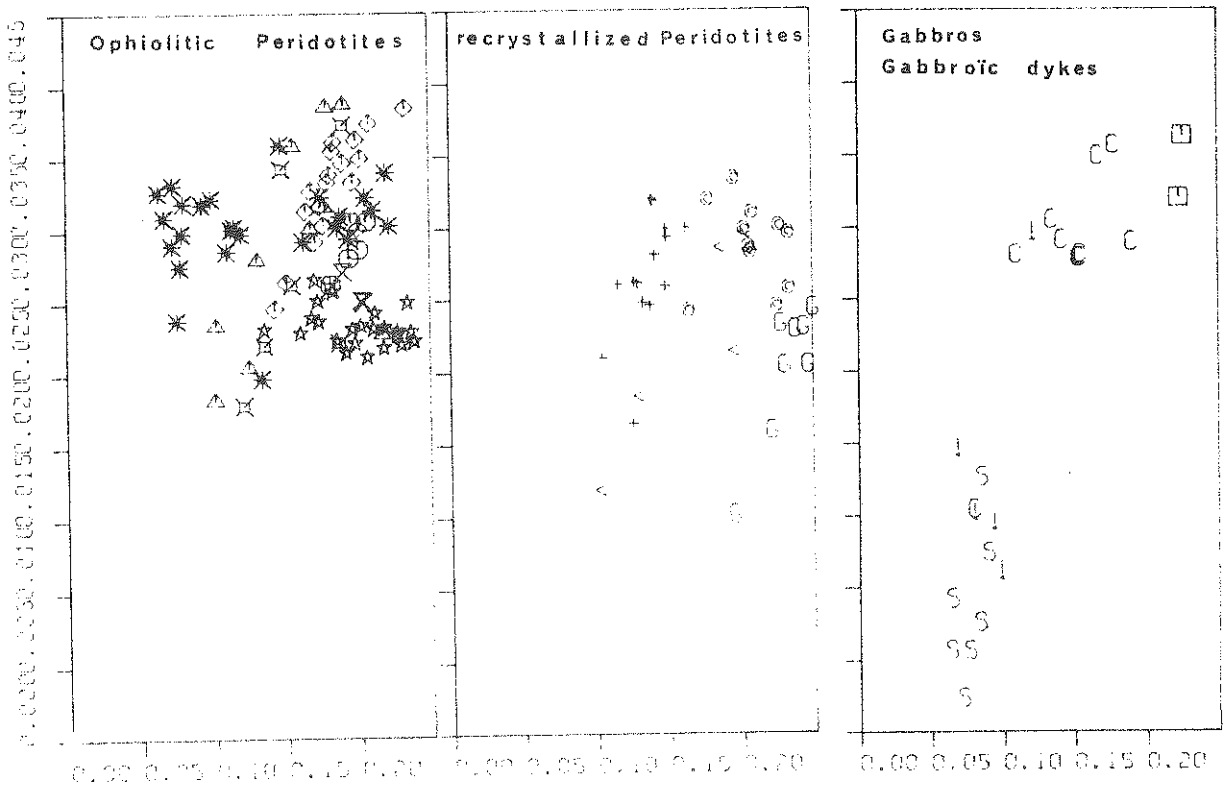
et les plag[illegible]es (fig. IV 21 et 25). Les fusions partielles seraient au moins antérieures au faciès à plagioclase.

Dans certaines péridotites et roches associées, les variations minéralogiques résultent de fractionnement et de contamination qui témoignent du passage

Ocean



Western Tethys



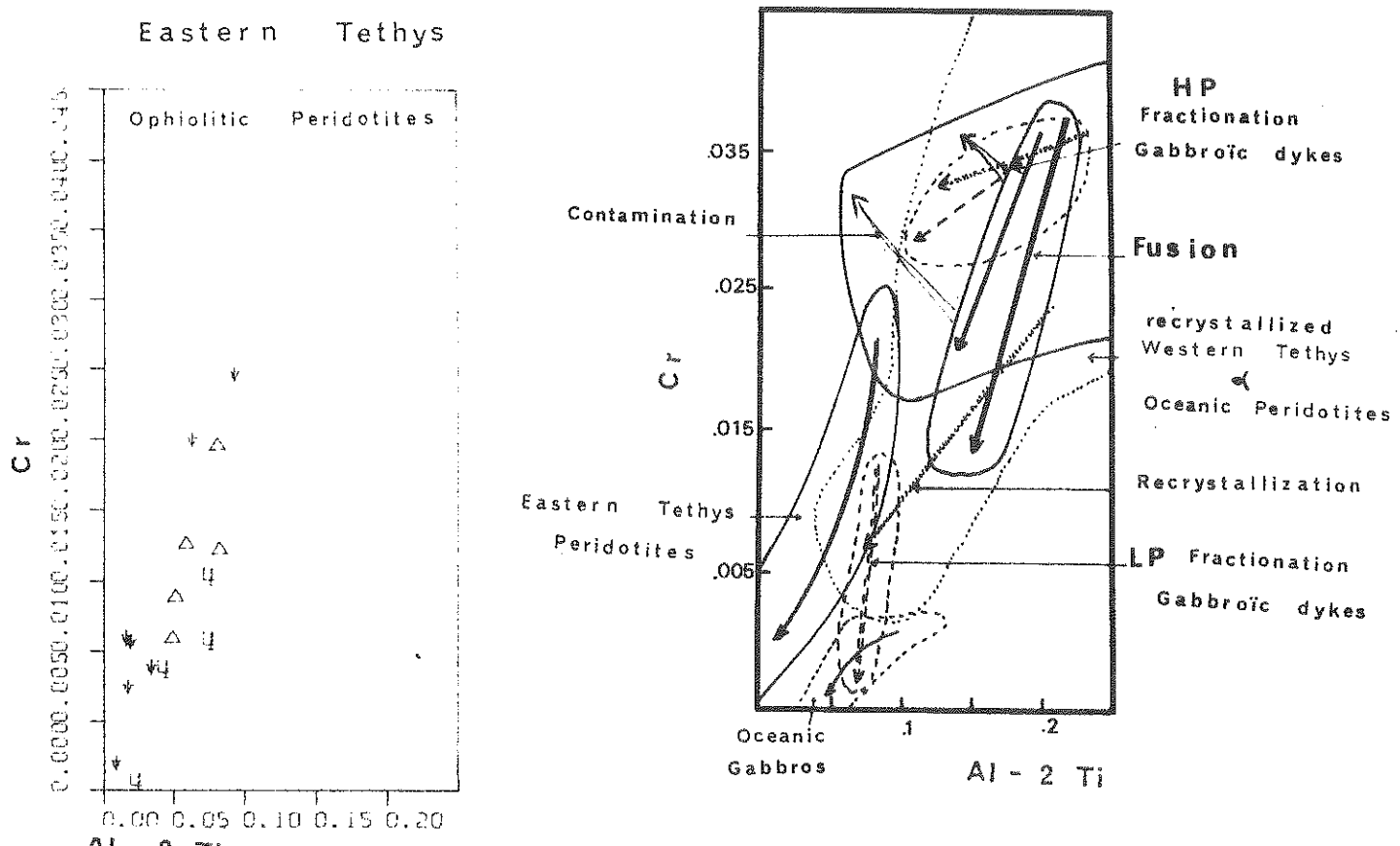


Fig. IV.25 - Variation de Cr en fonction de Al-2Ti dans les Cpx des péridotites et gabbros. Les principales évolutions pétrogénétiques sont indiquées dans un diagramme supplémentaire (OHNENSTETTER, en préparation).

d'un liquide. Ceci est particulièrement net dans le cas des spinelles. En Corse, des fractionnements de hautes pressions conduisent à la cristallisation de spinelles sub-automorphes riches en Al et Mg par rapport aux spinelles sous forme de vermicules (fig. IV.26). Dans les filons à bord diffus, le fractionnement à partir d'un magma picritique relativement riche en Al conduit à une diminution progressive de Al dans le Sp alors que Fe^{3+} augmente. Il y aurait toujours équilibration entre Sp, Ol et les liquides. Les fractionnements de basses pressions conduisent à des diminutions rapides de Cr dans les Cpx et Sp comparés à ceux de hautes pressions dans le faciès des péridotites à plagioclase. Les contaminations provoquent l'enrichissement en Na, Ti et Cr des pyroxènes et Sp (Fig. IV.22, 23, 24) postérieurement au stade de fusion partielle principale. L'étude du Monte Maggiore montre en particulier la coexistence de liquides plus ou moins magnésiens dont certains sont riches en alcalins et éléments incompatibles. Ils permettent d'expliquer certaines des hétérogénéités observées dans les bassins corses ou océaniques. Enfin, la percolation de liquides, responsables des déséquilibres minéralogiques et chimiques a conduit à des écarts de tempé-

rature importants lors de l'estimation des conditions thermométriques. Les températures décroissent depuis le stade des faciès à pyroxène alumineux jusqu'aux péridotites contaminées dans le faciès à plagioclase.

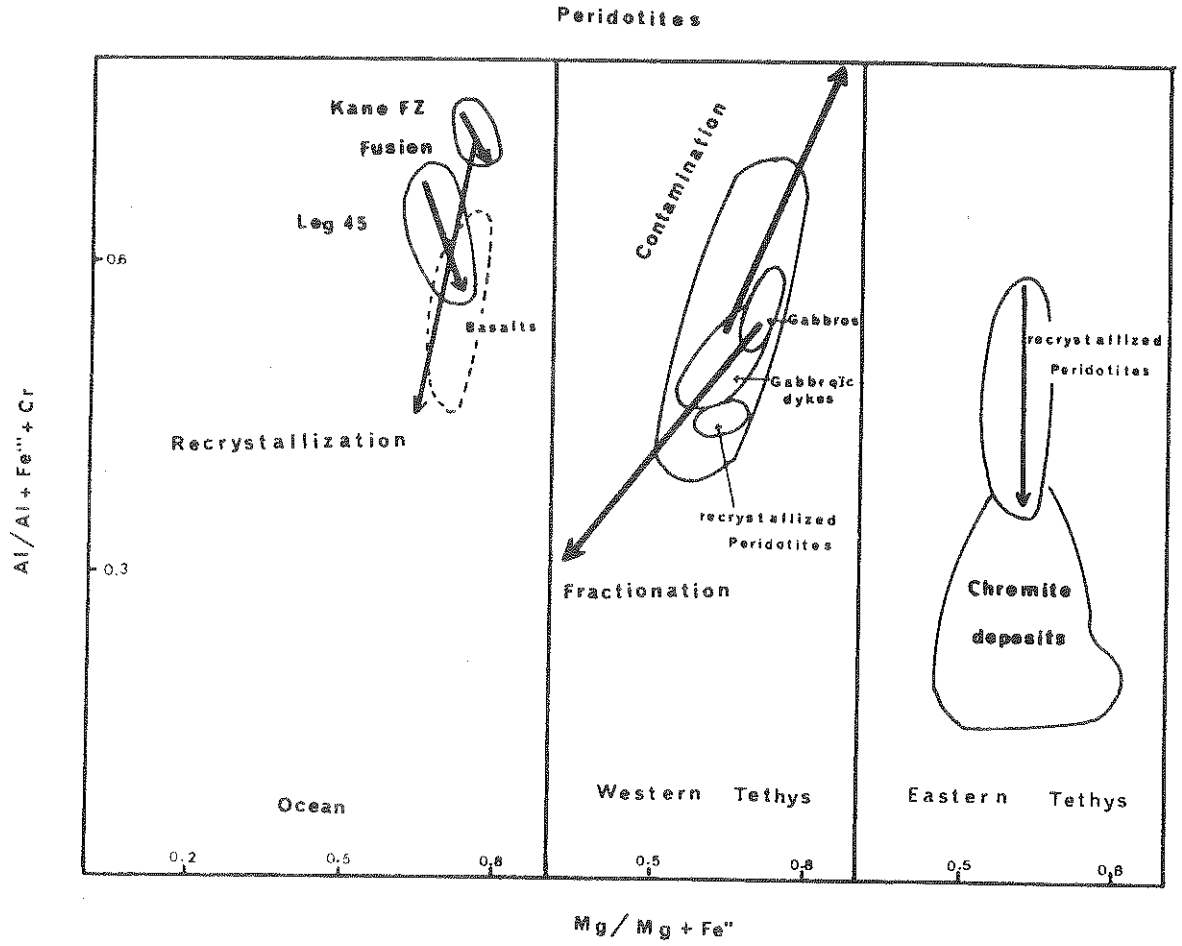


Fig. IV.26 - Variation du rapport Al/Al + Fe^{3+} + Cr en fonction de Mg/Mg+ Fe^{2+} dans les spinelles lors de quelques processus pétrologiques. (OHNENSTETTER , en Prep.).

V - CONCLUSION

Les études pétrologiques entreprises sur les manteaux ophiolitiques ont permis de distinguer les résidus de Méditerranée orientale et occidentale par rapport aux péridotites océaniques. Dans les péridotites de Méditerranée orientale, le caractère résiduel plus accusé se traduit par de plus faibles porportions de Cpx, un rapport Mg/Mg+Fe plus élevé de même que les concentrations en Cr des Sp et des concentrations d'éléments incompatibles et alcalins plus faibles. Ces divergences peuvent correspondre à des sites géotectoniques différents. La formation des ophiolites de Méditerranée occidentale peut concorder avec un début d'océanisation au cours duquel l'asthénosphère est renouvelée. En Méditerranée orientale l'ouverture des bassins ne s'accompagne pas dans un premier temps d'une régénération du manteau, ce qui amène à des fusions de plus en plus appauvries avec le temps.

Toutes les péridotites se sont d'abord élevées dans des diapirs caractérisés par des textures protogranulaires et poecilobl**a**stiques. Dans les diapirs des fusions et fractionnements de hautes pressions peuvent avoir lieu. Avec le temps, les contraintes déviatoriques qui augmentent provoquent la déformation du diapir et la circulation des magmas le long de plans bien réglés. Les roches grenues des filons résultent d'extractions réalisées à différentes pressions à partir d'un magma ascendant. Selon les températures du magma et de l'encaissant les effets de mur seront plus ou moins nets.

Les cumulats ont cristallisé alors que l'inclinaison des lignes de flux était généralement assez forte. Ce dispositif est compatible avec le modèle d'intrusion en dike dans une ride qui atteint son état d'équilibre. Il peut aussi traduire une océanisation peu avancée où le refroidissement de la croûte et du manteau sont relativement rapides. En Méditerranée orientale, l'importance de l'extension et du magmatisme augmentent des ophiolites à structures verticalisées aux ophiolites à structures horizontales où il y aurait rotation des lignes de flux.

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Na AND Cr CONTENTS IN CLINOPYROXENES FROM PERIDOTITES: A POSSIBLE DISCRIMINANT BETWEEN "SUB-CONTINENTAL" AND "SUB-OCEANIC" MANTLE

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The compilation of data available in the literature and new analyses show that clinopyroxenes are significantly richer in Na and poorer in Cr in peridotites associated with high-grade metamorphic rocks than in ultramafites from oceanic environments, considered as "sub-continental" or "sub-oceanic" mantle, respectively. Two distinctive fields can be drawn in the Na-Cr plot. This fact is related to the large amount of basic magma provided by the oceanic mantle along the mid-oceanic ridges.

Application of this Na-Cr diagram to clinopyroxenes from peridotites in orogenic belts and appearing as xenoliths in volcanic rocks and kimberlites ("nodules") allows us to specify their origin, taking into consideration that the clinopyroxene composition is controlled by several factors each of which gives rise to a particular trend: *P-T* conditions, mineral facies, partial melting and crystal fractionation, metasomatism. It appears that oceanic-type mantle may be found under continents in extensional areas having evolved towards rift systems, and in ophiolites. The latter exhibit different degree of depletion related to their formation in two main geotectonic situations: mid-oceanic ridges and active margin systems.

1. Introduction

From geophysical investigations as well as from petrological observations, peridotites are thought to make up the upper mantle beneath both oceanic and continental crust. Fragments of such peridotitic material of deep origin are well known as inclusions in volcanic rocks [1] and may also have been emplaced as tectonic slices [2] at shallow depth, in some specific situations: (1) along continental margins during collision [3] or opening [4] processes; (2) along transform faults or ridges [5] or as serpentinite protrusions [6,7] in oceanic environments.

Because continental and oceanic areas may be expected to have suffered different evolutions with time, it would be of interest to know whether the mantle is of different composition beneath continental or oceanic crust; if so such differences might be used to characterize the geotectonic origin of the various ultramafic masses observed in orogenic belts.

During recent years numerous ultramafic samples have been collected from oceanic environments and comparisons are now possible with on-land peridotites for which sub-continental or sub-oceanic origins have already been proposed [8] on the basis of the mineral proportions. Nevertheless, bulk-rock geochemistry as well as modal compositions cannot give reliable information if, as is usual, the ultramafic rocks suffered low-temperature metasomatic alteration (serpentinization, argillization). The present work is an

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attempt to find a way to distinguish between sub-oceanic and sub-continental peridotites by using the Na and Cr contents of their clinopyroxenes. It enables one to point out several petrological processes expected to play a part at upper mantle level.

2. Composition of the clinopyroxene phase in the ultramafic tectonites: Na and Cr contents

2.1. Na and Cr in clinopyroxene

Na enters the M2 site of diopside ($CaMgSi_2O_6$) and may be coupled with Al, Fe^{3+} and Cr in the jadeite ($NaAlSi_2O_6$), acmite ($NaFe^{3+}Si_2O_6$) and kosmochlor (= ureyite $NaCrSi_2O_6$) end-members, respectively. At 800°C complete solid-solution between jadeite and kosmochlor exists above 18 kbar [9]. Jadeite and acmite are able to form a large range of solid-solution with diopside, whereas the diopside structure can only accept up to 24% of kosmochlor at 1 atmosphere [10]. The solubility of kosmochlor decreases strongly with increasing pressures so that kosmochlor is not present in diopside above 45 kbar [11]. Cr may be incorporated in kosmochlor (M1 site) and/or in the Tschermak-type molecule ($CaCr_2SiO_6$; M1 and probably the tetrahedral site in the Tschermak molecule). The highest solubility of $CaCr_2SiO_6$ in diopside at one atmosphere is 14% at 1200°C [12] and 6.7% at 940°C [13]. The solubility of the Tschermak-type molecule decreases with falling temperatures [14].

2.2. Mineral facies and clinopyroxene composition

The mineral parageneses of ultramafic rocks involve mainly the following associations [15]: olivine + orthopyroxene + clinopyroxene + an aluminous phase, the nature of the latter being related to the pressure of crystallization. In the spinel- and garnet-lherzolite fields, above 8 or 10 kbar [16], clinopyroxene is the only Na-bearing phase (the minor amount of Na in garnet and orthopyroxene being taken out of consideration); its composition thus directly reflects the Na content of the whole rock. In the plagioclase-lherzolite field, below 8 kbar, clinopyroxene is equilibrated with plagioclase and Na is distributed between these two phases. On the other hand spinel, and to a lesser extent garnet, but not plagioclase, are able to accept Cr in their structures; as a consequence clinopyroxene should be poorer in Na and richer in Cr in the plagioclase-lherzolite field than in the spinel- and garnet-lherzolite fields.

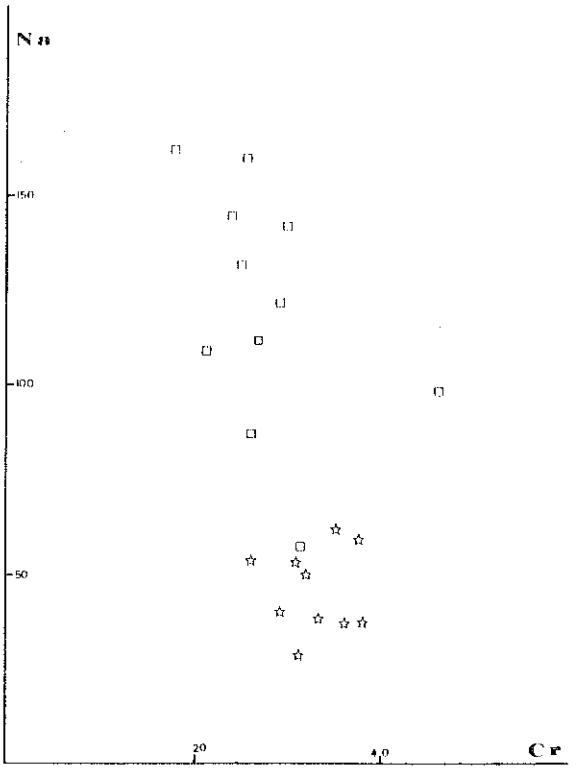

Fig. 1. Na versus Cr diagram for clinopyroxenes from the Ronda peridotite massif. Na and Cr are expressed in parts per mille from the structural formula calculated on the basis of six oxygen atoms. Squares = spinel-lherzolites and spinel-harzburgites; stars = plagioclase-bearing lherzolites and harzburgites. A correlation coefficient ($r = 0.722$) has been calculated by using the 20 points excluding the Cr-richest composition; thus the variables appear to be interdependent. Data from references 19, 24, 72, 73.

These assumptions seem to be fairly well supported by a plot (Fig. 1) of Na and Cr contents of clinopyroxenes from the Ronda massif (southern Spain). This massif clearly shows [17–19] the transition in the field from the spinel- to the plagioclase-lherzolite facies through the Seiland subfacies [15]. In terms of Na and Cr contents, the clinopyroxene compositions are strictly related to the mineral facies (Fig. 1).

3. Discrimination between "sub-continental and "sub-oceanic" peridotites

Sheared high-pressure ultramafic bodies are frequently related to high-grade metamorphic rocks (high- and low-pressure granulite facies) either in close structural association with them (Beni Bousera, Morocco; Ronda), or outcropping along the same fault zones (North Pyrenean zone, etc.). It is believed [20, 21] that such associations could have been derived from a deep continental crust/upper mantle situation ("paleo-Moho") prior to the solid state tectonic emplacement of the rocks. The peridotites (mainly belonging to the spinel-lherzolite facies with either subordinate amount of garnet- or plagioclase-bearing rocks) could be considered to have a "sub-continental" origin. Na and Cr contents of clinopyroxenes from peridotites at Beni Bousera, Lherz and Ronda have been plotted in Fig. 2, as representative of such a "sub-continental" mantle.

Ultramafic rocks have been dredged and cored in oceanic environments such as in transform fault zones and in the vicinity of mid-oceanic ridges. They are mainly spinel-peridotites (lherzolites and harzburgites) with more or less secondary plagioclase. No garnet-peridotite has yet been described. Na and Cr contents of clinopyroxenes from peridotites collected in the Atlantic and Indian Oceans have also been plotted in Fig. 2.

The Na-Cr diagram for clinopyroxenes (Fig. 2) clearly shows separate fields for "sub-continental" and "sub-oceanic" peridotites: (1) the "sub-continental" field is characterized by Na-rich clinopyroxenes in the spinel- (and garnet-) lherzolite facies; the Na content decreases and Cr content increases as the following reaction takes place: cpx + opx + sp → ol + pl (or, more precisely [22]: cpk + opx + Al-sp → ol + pl + Cr-sp); (2) the "sub-oceanic" field would be mainly characterized by clinopyroxenes with very low Na contents, whatever the mineral facies; nevertheless, a few samples are rich both in Na and Cr (IN19X [22]).

From the above observations peridotites appear to be more depleted in Na (and probably in all other lithophile components) in oceanic areas than beneath the continents. This is probably the result of the large volume of melt which is supposed to have been extracted to provide the abyssal tholeiites of the oceanic crust, whereas the observed "sub-continental" peridotitic bodies would only have suffered small-scale partial melting [23,24] or none at all [25] (or have suffered melting at pressures where Na is retained in clinopyroxene rather than eliminated into the melt).

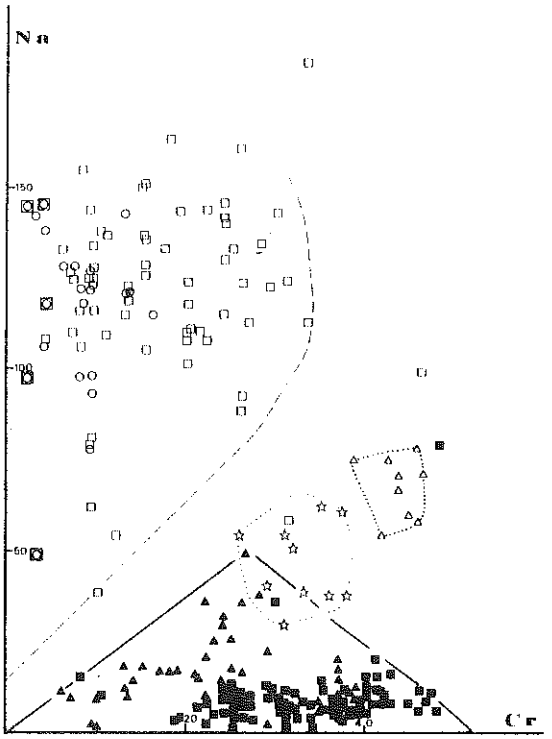

Fig. 2. Na-Cr diagram for clinopyroxenes from "sub-oceanic" (filled symbols) and "sub-continental" peridotites (open symbols). Squares = spinel-peridotites; triangles and stars = plagioclase-bearing peridotites; circles = garnet-peridotites; open circles in squares = spinel plus garnet peridotites. Solid line: "sub-oceanic" field; dashed line: "sub-continental" field; dotted line around stars: plagioclase-bearing peridotite field at Ronda; dotted line around open triangles: plagioclase-bearing peridotites on the Atlantic continental margin, west of the Galicia Bank [4]. "Sub-continental" peridotites: data from French Pyrenees [25,35,74–76], Beni Bousera [20,23] and Ronda [19,24,72,73]. "Sub-oceanic" peridotites: data from references 22, 77, 78, 79, 80, 81, 82, 83, 84.

4. Application to some alpine-type peridotites

4.1. Peridotites from the Alpine belt

Various mantle derived peridotite bodies outcrop in the Alps, either related to high-grade metamorphic rocks, as in the Ivrea zone, or associated with the ophiolitic fragments of the Liguro-Piedmont Basin (e.g. Zermatt–Saas Fee). The peculiar situation of the Lanzo massif, at the intersection of two structurral zones [26] may raise some questions about its origin. The Na-Cr diagram may be used to help to assign geotectonically these ultramafic fragments.

The Balmuccia and Baldissero bodies (Ivrea-Verbano zone) may be representative of sub-continental mantle [27,8,28] and they contain Na-rich clinopyroxenes (Figs. 3 and 4). Finero has clinopyroxenes

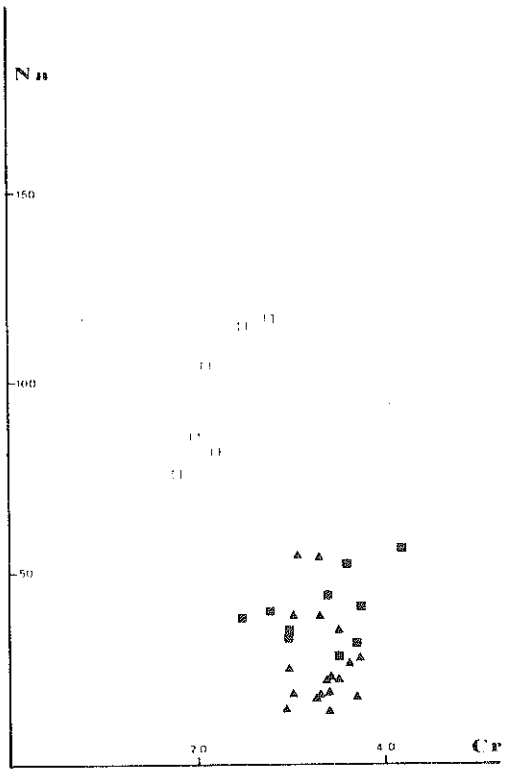

Fig. 3. Na-Cr diagram for clinopyroxenes from peridotites from the Alps. Open symbols: Baldissero; filled symbols: Lanzo. Squares = spinel-lherzolites; triangles = plagioclase-bearing lherzolites. Data from references 26, 27, 85, 86, 87, 88.

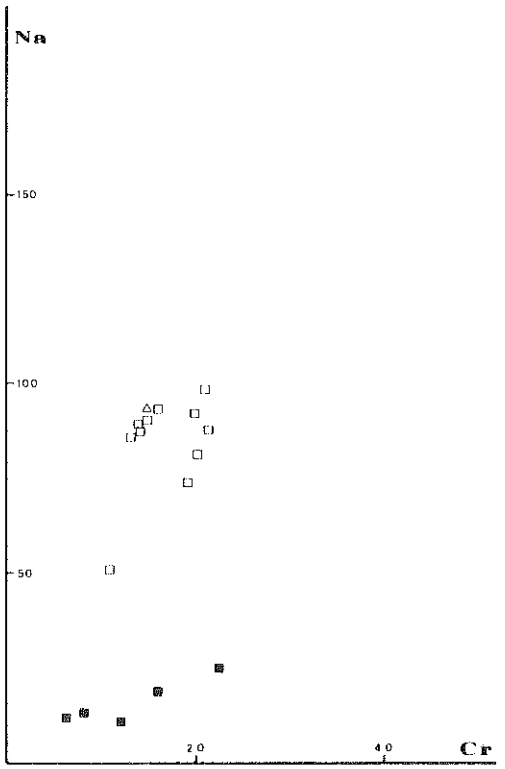

Fig. 4. Na-Cr diagram for clinopyroxenes from peridotites from the Alps. Open symbols: Balmuccia; filled symbols: Finero. Squares = spinel-lherzolites; triangles = harzburgite. Data from references 27, 28, 29, 87.

strongly depleted in jadeitic and Tschermak-type molecule and falls in the "sub-oceanic" field (Fig. 4); but its peridotite and associated gabbros are considered to be the products of fractional crystallization [29] and then this body should be considered separately. The clinopyroxenes from the Lanzo massif spread near the plagioclase-bearing peridotite field of Ronda, slightly more Na-rich than the clinopyroxenes in "sub-oceanic" peridotites.

The northern Appenines involve the western Ligurides (with the Voltri Group) and the eastern Ligurides divided into the Internal and External Ligurides. Peridotites in the Internal Ligurides should be more depleted than in the External Ligurides, the latter being suspected to have suffered an earlier rifting stage [30]; the Na-Cr plot for clinopyroxenes (Fig. 5) effectively defines two different areas, compositions in the Internal zone being close to the "sub-oceanic"

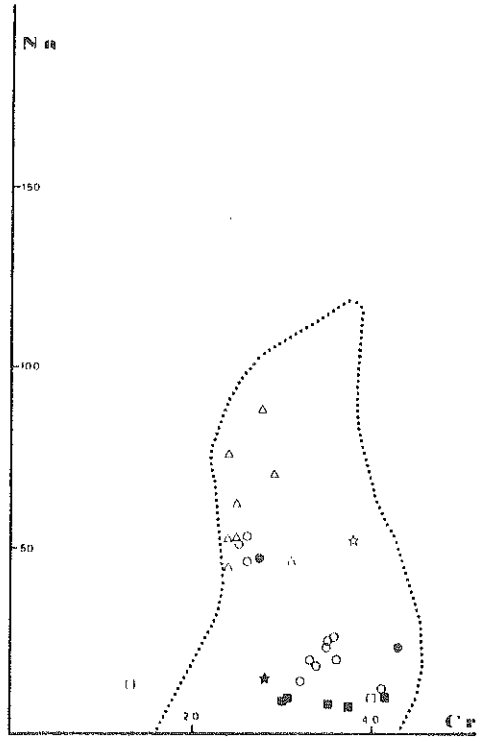

Fig. 5. Na-Cr diagram for clinopyroxenes from peridotites of northern Apennine ophiolites. Voltri Group: open circles = spinel-lherzolites; filled circles = pyroxenites. Internal Ligurides: open squares = plagioclase-bearing lherzolites; filled squares = websterites; filled star = average of clinopyroxene compositions [30]. External Ligurides : triangles = plagioclase-bearing lherzolites; open star = average of clinopyroxene compositions [30]. Other data from references 89, 90, 91, 92. Dotted line: Monte Maggiore clinopyroxene composition field (see Fig. 6).

field. In the Voltri Group, the clinopyroxenes from the Erro-Tobio massif plot in each of the two Ligurian fields.

The lherzolitic body at Monte Maggiore (Corsica) is described as the sole of a reconstructed ophiolitic pile which would represent old oceanic crust [31, 32]. It underwent syntectonic recrystallizations and a partial melting event during upwelling at an accreting plate boundary possibly in the vicinity of a heterogeneous ridge segment [33]. The field defined by the Na-Cr clinopyroxene plot covers the "sub-oceanic" domain and reachs the "sub-continental" spinel-lherzolite field. By comparison with the Alpine bodies considered above, the Monte Maggiore plot covers the whole area corresponding to Lanzo and both External and Internal Ligurides fields. It is not possible to point out any simple correlation in that massif between clinopyroxene compositions and modal proportions of the different samples: a clinopyroxene from a dunite presents the highest Na content, whereas a lherzolite and a pyroxenite are characterized by clinopyroxenes having the lowest Na contents, although these samples represent petrographically the best candidates for further melt extraction. On the other hand important differences may be observed for Na contents of clinopyroxene crystals in a single type of rock, either dunite or lherzolite; moreover, a positive correlation appears between Na and Cr in clinopyroxenes from a dunite (Fig. 6). The lherzolite sample exhibiting various

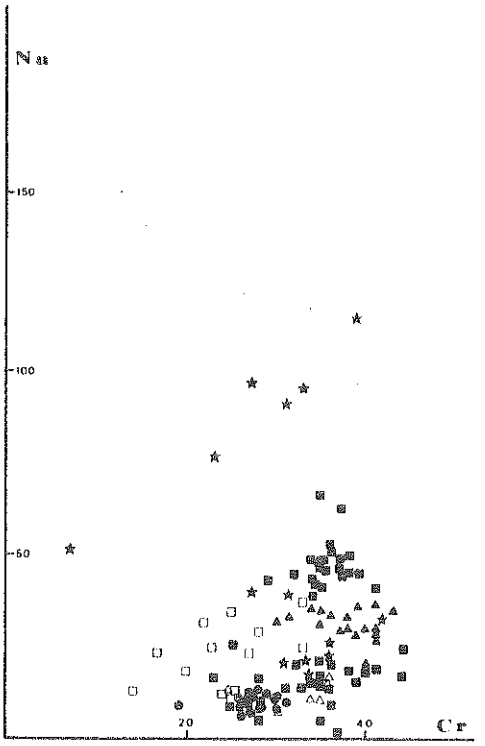

Fig. 6. Na-Cr diagram for clinopyroxenes from plagioclase-bearing peridotites at Monte Maggiore, Corsica (filled symbols). Open symbols: clinopyroxenes from other Corsican Massifs. Squares = lherzolites; circles = pyroxenites; triangles = harzburgites; stars = dunites. Data from reference 34.

high Na contents in clinopyroxene contains accessory phases like Ti-kaersutite, Ti-rich spinel and apatite which indicate the relative abundance of incompatible elements in the rocks. All of these petrographic and geochemical features are best interpreted by a metasomatic hypothesis [34]: fluids and liquids enriched in incompatible elements, migrating under tectonic control and trapped in some specific situations, may locally modify the mineralogy of the more or less refractory surrounding peridotites. The amphibole-bearing lherzolites at Caussou (French Pyrenees) could be another example of such a process [35]. Thus the compositions of the clinopyroxenes from the Monte Maggiore lherzolite may reflect both partial melting and metasomatism, perhaps during uprising of a mantle diapir in an oceanic accretion zone.

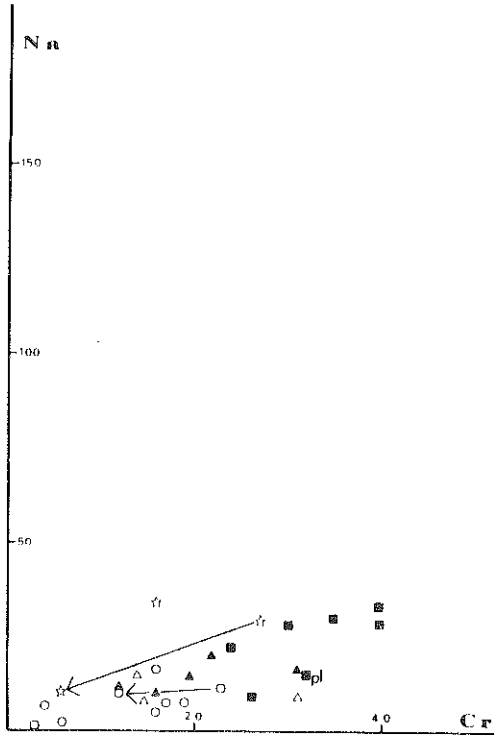

Fig. 7. Na-Cr diagram for clinopyroxenes from peridotites in the peri-Pacific ophiolite belt. Open symbols: New Zealand; filled symbols: New Caledonia. Squares = spinel-lherzolites except one point (*Pl*) representing a plagioclase-bearing lherzolite; triangles = harzburgites; circles = pyroxenites; stars = dunites. The arrows represent the core to rim zoning of clinopyroxene. Data from references 93, 94, 95.

4.2. Peridotites from the peri-Pacific ophiolitic belt

All analyzed clinopyroxenes from New Zealand and New Caledonian sheared ophiolitic peridotites plot in the "sub-oceanic" field (Fig. 7). The refractory samples, harzburgites and dunites, have clinopyroxenes which are Na- and especially Cr-poorer than those in the lherzolites from New Caledonia, the clinopyroxenes of which plot in the fields for the Internal Ligurides and some of the Corsican peridotites. As discussed below, this correlation, also exhibited by compositional changes between cores end rims, may be linked to a particular partial fusion event.

5. Behaviour of clinopyroxenes from inclusions in basaltic rocks

Ultramafic "nodules" are mostly present in alkali basalts and related rocks erupting in continental, circum-oceanic and oceanic regions [36]. They are generally classified in two groups more or less closely associated in the field: the wehrlite series generated by crystal accumulation from the host basalts, and the lherzolite series mainly representative of mantle rocks [37–39]. Only the latter group, essentially made of spinel-lherzolites and harzburgites with minor garnet-peridotites, will be considered here (Fig. 8).

5.1. Ultramafic "nodules" from oceanic areas

The clinopyroxenes from garnet-peridotite inclusions at Hawaii fall in the "sub-continental" field determined above, with high Na contents and no evidence of depletion features. Those of the xenoliths of spinel-harzburgites and lherzolites dredged near St. Paul Rock (Atlantic Ocean), supposed to be fragments of residual mantle unrelated to the host basanitoid [40], also fall in the "sub-continental" field or in the Na-rich part of the Monte Maggiore area, as do the clinopyroxenes from the ultramafic rocks of the Canary Islands [41] and of the ultramafic suite from Tahiti [42,43]. The existence of relatively Na-rich clinopyroxenes from such sub-oceanic peridotites may indicate that sub-oceanic mantle equilibrated in the spinel- or garnet-facies does not present every-

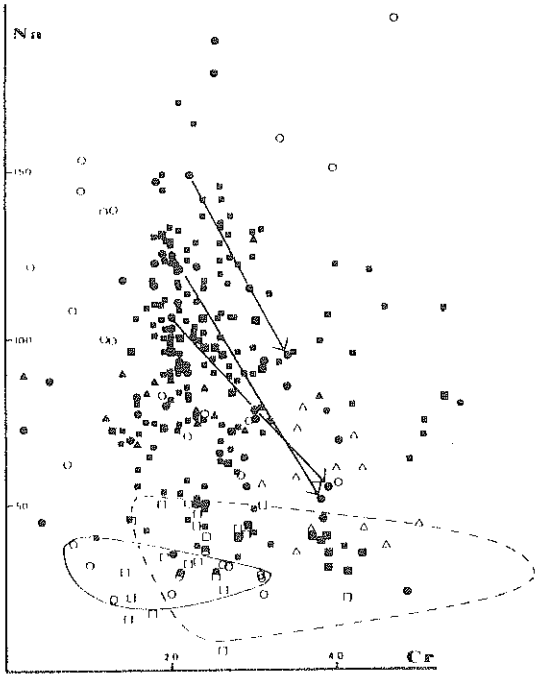

Fig. 8. Na-Cr diagram for clinopyroxenes from "nodules" in basaltic lavas. Open squares = island arcs; open circles = oceanic Islands; filled triangles = Kenya Rift system; open triangles = Assal Rift system; filled squares = French Massif Central; dots = other occurrences in continental alkali basalts. Solid line: envelope of compositions from Reunion Island [48,49,73]; dashes line: envelope of compositions from the Goss Pile area [47]. Data from references 38, 39, 41, 44, 45, 46, 47, 48, 49, 59, 64, 73, 96, 97, 98, 99, 100, 101, 102, 103, 104, 105, 106, 107, 108.

where a strongly depleted character. On the other hand xenoliths from Itinome Gata (Japan) a circum-oceanic area [44], exhibit Na-poor clinopyroxene sometimes plotting in the "sub-oceanic" area, although they may contain garnet.

5.2. Ultramafic "nodules" from continental areas

On the continents ultramafic xenoliths are known from extensional zones of rift systems and from alkaline basalt districts. Spinel-lherzolites from the Assab Rift have clinopyroxenes plotting around the "Lanzo area"; peridotites from the Tanzanian Rift, considered as cumulates [45], have clinopyroxenes richer in Na and poorer in Cr. Concerning the within-continental plates alkali basalts, their ultramafic inclusions have clinopyroxenes spreading widely in the Na-Cr diagram, plotting in the "sub-continental" as well as in the "sub-oceanic" fields and exceeding strongly these area limits towards low Na and low Cr contents. Some zoned clinopyroxenes, for instance from Alaska [46], present a Na-depleted and Cr-enriched rim (Fig. 8); this feature could be linked with partial melting events, even if that trend appears to be roughly parallel to the compositional evolution at Ronda for the mineralogical change from spinel to plagioclase facies; the scarcity of plagioclase-bearing ultramafic xenoliths does not allow one to establish this latter effect. Conversely, the large dispersion of the compositions suggests that partial melting is not the only effect to be taken into consideration; some of the "nodules" whose clinopyroxenes are both Na- and Cr-poor, could be related to early crystal segregation in basaltic melts, by comparison with some of the Goss Pile and Reunion xenoliths which are known to be cumulates in alkali or transitional basalts, respectively [47–49].

6. Behaviour of clinopyroxene from garnet-bearing inclusions in kimberlites

Amongst the peridotite "nodules" brought up to the surface in kimberlite diatremes, two main groups are generally recognised, according to their textures and geochemistry. The sheared nodules, representative of a slightly depleted mantle, show higher temperatures and pressures of equilibration than the more depleted granular peridotites [50]. They may have been affected by metasomatic events before their incorporation in the kimberlitic magma [51–54]. Beside these two groups of peridotites, the formation of "discrete nodules" resembling to the sheared ones is related to crystallization in early magmas [55, 56]. A cumulate origin is also proposed for some granular lherzolites [57].

The distribution of clinopyroxene compositions in the Na-Cr diagram is such that it may be classified into three main groups (Fig. 9). The discrete nodules have the lowest Cr contents in clinopyroxene, the granular the highest. Considering the Na contents of the rock-forming clinopyroxenes, depleted or undepleted mantle has broadly similar compositions.

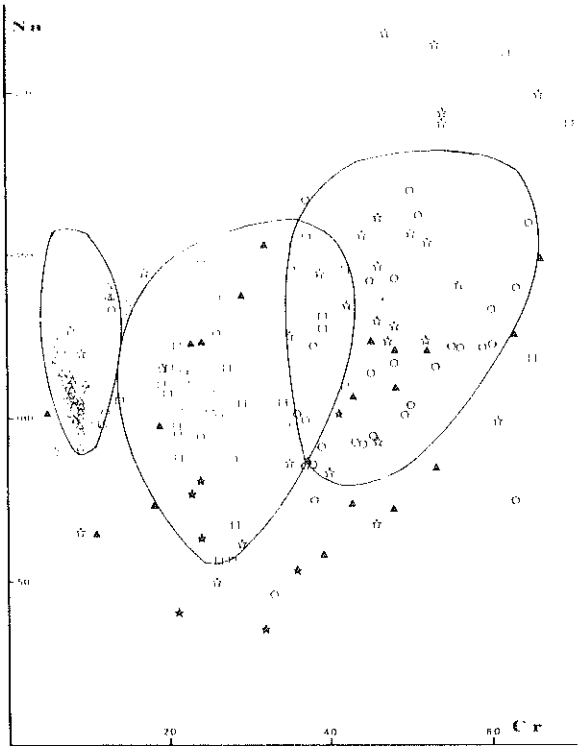

Fig. 9. Na-Cr diagram for clinopyroxenes from "nodules" in kimberlites. Open squares = sheared garnet-peridotites; open circles = granular garnet-peridotites; open stars = unclassified garnet-peridotites; filled stars = spinel-peridotites; open triangles = discrete nodules; filled triangles = other samples. Solid lines from left to right: envelopes of discrete nodules, sheared peridotites and granular peridotites, respectively. Data from references 45, 50, 53, 55, 58, 109, 110, 111, 112, 113, 114, 115, 116, 117, 118, 119, 120, 121, 122, 123.

Some of the granular rocks having Na- and Cr-rich clinopyroxenes may testify to a metasomatic event as does the Na-Cr correlation in clinopyroxenes from the inclusions in the Matsoku pipe [58].

The scarce spinel-lherzolite samples associated with the garnet-bearing rocks seem to have clinopyroxenes significantly poorer in Na.

7. Discussion

The above observations suggest that the variations in the Na and Cr contents of clinopyroxenes from four-phase peridotites depend on P and T conditions, mineral facies, partial melting, fractional crystallization and metasomatism, all playing their part in different geotectonic situations. The following discussion is an attempt to evaluate these different factors (Fig. 10).

P and T conditions. The axis along which clinopyroxene is enriched in jadeite-type (especially jadeite, kosmochlor and Jd_{50}-Ks_{50}) and Tschermak-type molecules are represented in the Na-Cr diagram (Fig. 10). As the stability of kosmochlor decreases with increasing pressures [11], this end-member may play a reduced part at high pressures, when Tschermak-type molecules may record chromium variations. This variation nearly horizontal in Fig. 10 is well displayed by the sheared and granular kimberlite nodules as well as by the nodules in alkali basalts and in the different massifs taken all together. Nevertheless, falling temperatures may also account for the decreasing solubility of Tschermak-type molecules [14].

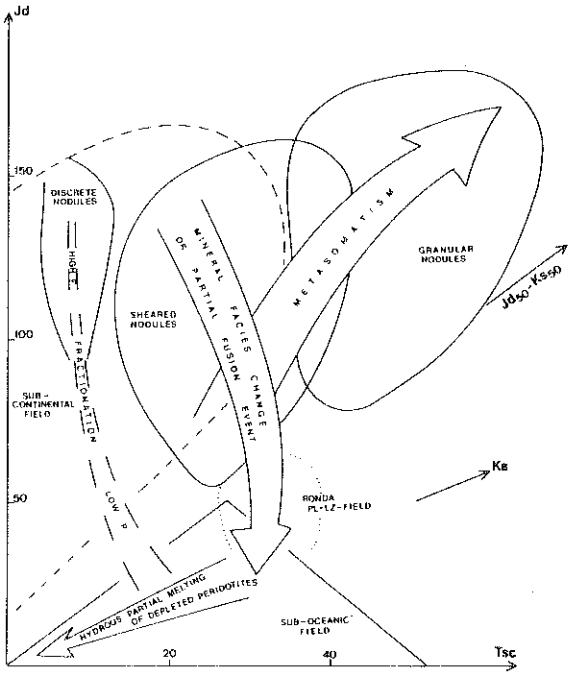

Fig. 10. Evolutionary trends of clinopyroxene composition in the Na-Cr diagram. Tsc: $CaCr_2SiO_6$; Ks: $NaCrSi_2O_6$; Jd: $NaAlSi_2O_6$; Jd_{50}-Ks_{50}: $Na_2AlCrSi_4O_{12}$. For discussion see text.

Mineral facies. As plagioclase is the accessory Na-rich and Cr-poor aluminous phase in the peridotites, the main variations of Na and Cr contents in associated clinopyroxenes appear to be at the transition between spinel- and plagioclase-lherzolite facies. When spinel- and garnet-peridotites are associated in some bodies or amongst ultramafic nodules, the Na and Cr contents of their clinopyroxenes do not present noticeable variations [59]. On the contrary a strong Na-depletion and clear Cr-enrichment are evident with the change from spinel- to plagioclase-bearing peridotites as in the Ronda massif.

Partial melting events. In almost every type of solid/liquid equilibration, Na enters preferentially into the liquid and Cr into the solid. Thus partial fusion in peridotites leads to an Na content decrease and a Cr content increase in the residual clinopyroxenes. In the Na-Cr diagram this evolution appears to be parallel to those of the spinel- to plagioclase-peridotite facies change (Fig. 10). From this observation, restricted partial melting events seem to have affected the "sub-continental" mantle. On the contrary extensional processes in rift zones as well as in oceanic areas seem to be the reason for the high depletion of the upper mantle, as in the Assab, Lanzo and ophiolitic peridotites. In oceanic areas the high geothermal gradient prevents garnet-bearing assemblages [60] and partial fusion may occur in garnet- and spinel-free rocks (Lesotho facies [15,61]) or in spinel-peridotites. The depleted mantle would be spinel- or plagioclase-lherzolites, more or less impoverished in clinopyroxene, rather than a harzburgitic body [22].

Another type of partial fusion event seems to be characterized by decrease of both the Na and Cr contents of the residual clinopyroxenes, along the kosmochlor and possibly the Ca-Cr-Tschermak molecule lines. This appears by plotting the core to ring evolution in clinopyroxenes (Fig. 7). The entry of Cr into liquids (future source for the chromite deposits) during remelting of a depleted lherzolitic mantle [62] is favoured by a hydrous partial melting of an oceanic mantle above a subduction zone [63].

Crystal fractionation. The four-phase peridotitic associations are sometimes considered as representative of high-pressure crystal fractionation products in picritic melts rather than depleted mantle after partial fusion events [25,64]. Xenoliths at Goss Pile [47] and on the Reunion Island [48,49], rocks in the Finero massif [29], have been suggested to have such an origin, as have the discrete nodules in kimberlites [55]. The Cr content of the clinopyroxene is most generally low in high-pressure assemblages, but can be found much higher by low-pressure precipitation together with chromite (Goss Pile, Reunion). The Na content must be controlled by the pressure of crystallization (incorporation of jadeite) and/or by the oxygen fugacity (incorporation of acmite). The field of Na-Cr compositions for clinopyroxenes having such an origin must thus be at once roughly parallel to the Na-axis of the diagram, spreading from the discrete nodules to the Reunion areas (Fig. 9); it bends and becomes after roughly parallel to the Cr-axis (secondary strong subsolidus Cr enrichment of clinopyroxene side by side with chromite crystal have been observed in Reunion xenoliths). Some of the ultramafic inclusions from alkali basalts in the French Massif Central (and elsewhere?) have clinopyroxenes having compositions related to that field rather than to the depletion trend as represented in Fig. 10.

Metasomatism. All the petrological processes discussed above lead to Na content decreases in the residual products, with increasing or decreasing amounts of Cr, as the case may be. The positive correlation observed for Na and Cr contents in clinopyroxenes from some Corsican samples and kimberlite nodules needs another explanation taking into account *enrichments* in jadeite and/or kosmochlor molecules, in so far as the products believed to be the more depleted have also the most Na- and Cr-rich clinopyroxenes (granular nodules in kimberlites, some of the Corsican dunites). As Na and Al show affinities for the liquids, the observed jadeite enrichment may be due to local metasomatism of various kinds of peridotites at different pressures. Secondary enrichment in kosmochlor as in Corsica may occur at lower pressures than jadeite enrichment in kimberlite nodules. Alkali and incompatible element-rich liquids may locally invade more or less depleted ultramafic upper mantle along well-defined planes [34]. The veining liquids which petrographically and geochemically modify the surrounding rocks would be responsible

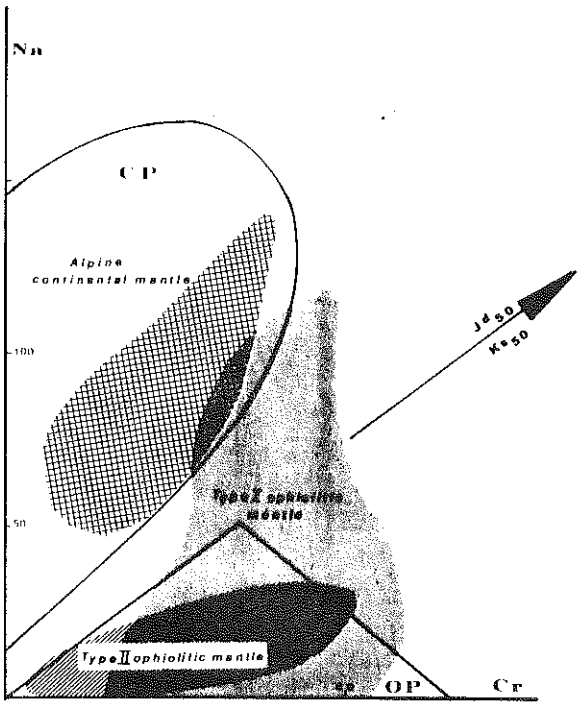

Fig. 11. Respective fields of ophiolitic (dotted and hachured areas) and alpine mantle (cross ruled area). Envelopes of "sub-oceanic" (*OP*) and "sub-continental" (*CP*) peridotites as in Fig. 2. For discussion see text.

for some heterogeneity. This process may easily occur in mantle fragments ascending as nodules in alkali-rich basalt [46] and kimberlites [65]. In oceanic rocks it is to be expected rather in anomalous ridge segments and in transform fault zones, where alkali basalts are known [66,33]. It is unexpected or scarce in the large massifs from continental areas, where partial fusion is limited, and in peridotites from some "type II ophiolites" [67] associated with liquids strongly impoverished in incompatible elements.

8. Conclusion

The observations and discussion above lead to the conclusion that Cr and Na contents in clinopyroxenes from four-phase peridotites appear to be controlled by various petrological mechanisms and, therefore, it is generally difficult to draw any unique interpretation from the observed values. Nevertheless, some points can be emphasized.

(1) Being much poorer in Na, clinopyroxenes from "sub-oceanic" peridotites plot in an area almost completely separated from the "sub-continental" clinopyroxene field. Because both type of peridotites are mainly equilibrated in the spinel-lherzolite facies, this clear discrepancy is better explained by different degrees of depletion related to the geotectonic setting: thus the Na-Cr discriminant diagram allows one to distinguish between peridotites having suffered extensive partial melting in an oceanic environment and poorly depleted rocks from the bottom of the continental crust. This is not actually in good agreement with information drawn from the Rb-Sr contents in peridotites and their clinopyroxenes [68], according to which sub-continental and sub-oceanic mantle may have suffered the same kind of evolution. Even if the "sub-continental" peridotites do suffer single [23,24] or manifold [69] partial melting events, the amounts of liquids produced seem to have been much smaller than the volume extracted from peridotites considered here as belonging to the "sub-oceanic" mantle.

(2) Sheared peridotites having clinopyroxenes with "sub-oceanic" characters are also found in rift systems and in ophiolitic piles. In this latter case clinopyroxenes in peridotites from type I and type II ophiolites [67] occupy two different overlapping fields characterized by negative and positive Na-Cr correlations, respectively (Fig. 11). Type I ophiolites could be explained by a more or less advanced stage of rifting; some relatively Na-rich clinopyroxenes plot in the plagio-clase-bearing field of the "sub-continental" mantle which could be regarded here as the edge of the continental margin at the time of the extension. Clinopyroxenes in peridotites from type II ophiolites exhibit a stronger overall Na depletion, characteristic of two main partial fusion events at least, the last one showing a Cr-depletion trend [70]. This behaviour is compatible with the generation of type II ophiolites in active continental margin systems, where the depleted sub-oceanic mantle may rise as diapirs [71].

(3) On the basis of the Na and Cr contents of their clinopyroxenes, some of the ultramafic nodules in basalts and kimberlites could be interpreted as early fractionation products in basic melts. They nevertheless appear to be commonly more or less depleted

fragments of peridotitic mantle, having sometimes suffered metasomatism. The striking fact is that the clinopyroxenes from nodules in basaltic rocks completely overlap whatever their origin, from oceanic or continental environments. It suggests sampling at rather different (deeper?) levels than those from which the ultramafic bodies came from, either in continental or in oceanic areas.

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"Na and Cr contents in clinopyroxenes from peridotites: a possible discriminant between 'sub-continental' and 'sub-oceanic' mantle"—a reply to G. Sen

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In a previous paper [1] a discrimination diagram based on the Na and Cr contents of clinopyroxenes was proposed to distinguish "sub-continental" mantle peridotites from dredged and cored oceanic peridotites. The positions of clinopyroxenes from some Alpine massifs, from nodules in basalts and kimberlites were also considered with respect to this diagram. The reliability of the diagram as well as some of the conclusions of the paper were disputed by Sen [2] mainly on the basis of oceanic sampling and clinopyroxene compositions in xenoliths from Hawaii, Tahiti and Kilbourne Hole, New Mexico. Nevertheless, Sen emphasized the main results of our paper, that is: (a) mantle peridotites (bodies) suffered different histories linked to their assumed geotectonic environment, (b) clinopyroxenes from xenoliths in alkali basalts do not show significant differences in the diagram, whatever their origin (oceanic islands or continental districts).

1. Comments on the establishment of the Na-Cr diagram

Sen [2] raised several problems mainly concerning the establishment of the discriminant diagram. In the present case, we have plotted Na-Cr contents of clinopyroxenes from mantle peridotites associated with high-grade metamorphic rocks and peridotites from oceanic areas. In both cases a choice was made only on the basis of the geotectonic setting. Consequently, petrogenetic considerations (such as degree of partial fusion, fractionation and recrystallization processes) were not taken in account at this stage of the work. The only hypothesis was that all the peridotites were assumed to be mantle rocks. As most of the data were taken from the literature, the origin of some samples can be debated.

2. Discussion of the origin of oceanic mantle peridotites

Sen [2] criticized the samples chosen to define the oceanic field, mentioning the cumulate of residual mantle origin of some peridotites dredged or cored from the ocean, especially for Leg 37 data. Oceanic peridotites are generally considered to be mantle fragments [3–10] having suffered more or less partial fusion. Nevertheless, a cumulate origin was not discarded [7,9,11–15]. Sometimes both hypotheses have been proposed for the same samples. Similarly a magmatic origin has been suggested [16] instead of a tectonic one for the banded ultramafites in continental massifs. It was not the aim of this paper to determine the cumulate or residual origin of peridotites, either in oceanic or in continental environments. The structural criteria advocated by Sen: "textural and mineral chemical evidence of deformation, recrys-

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tallization and partial fusion" to decide that ultramafic rocks are residual mantle samples rather than cumulates is insufficient and not acceptable, especially for samples whose field relationships are unknown. Moreover, the study of ophiolitic cumulates and layer 2 from the ocean has revealed that shear planes and strong deformation occur at low pressure in samples crystallized in a magma chamber [17]. On the contrary, numerous mantle ultramafites from the ocean are coarse grained and only slightly deformed in agreement with their protogranular textures [9,18], as for example samples from DSDP Leg 45 [8,10]. It may be noted that Leg 37 samples [14,19] are also only slightly deformed. Because of the difficulty of determining the origin of oceanic peridotites, all the clinopyroxenes from dredged and cored ultramafics were used, whatever their supposed origin. As suggested by Sen [2] we may discuss the origin and importance of Leg 37 samples to establish the Na-Cr diagram. The intrusives from site 334 belong to a "mélange" [20] made up of breccias with gabbro and peridotite clasts in a nannofossil foraminiferal ooze matrix interlayered with plutonic rocks. Eight clinopyroxenes from peridotites analyzed by Hodges and Papike [12,13], Clarke and Loubat [14], Symes et al. [15] were taken from three sections of the drill core: 22-2 (61–63, 70–75, 80–82 cm); 23-1 (46–54 cm); and 26-1 (20–22 cm). On the basis of textural and mineralogical

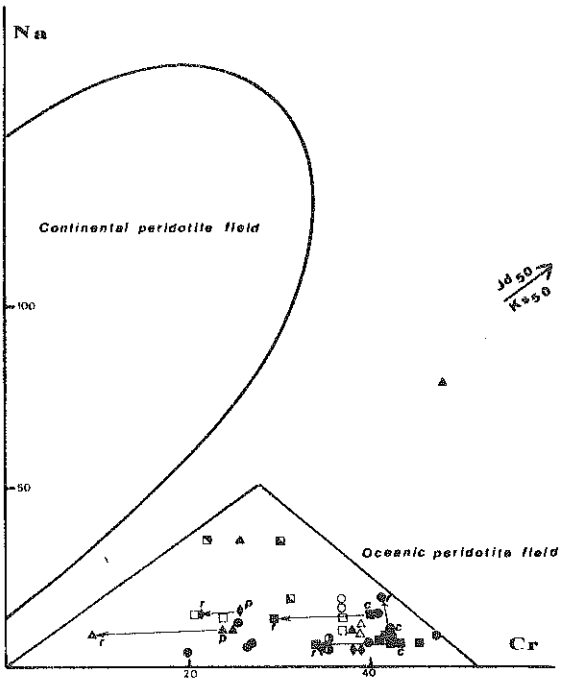

Fig. 1. Literature data on clinopyroxenes from oceanic peridotites. Open squares [14], open circles [15] and open triangles [13] represent spinel-plagioclase peridotites from DSDP Leg 37 assigned as cumulates; filled symbols represent residual mantle samples from Leg 45 (squares [10] and dots [8]), as well as Owen fracture zone [6] spinel lherzolites (triangles) and spinel + plagioclase (lozenges). Peridotites from the Vema fracture zone [7] (half filled dots) and spinel + plagioclase lherzolite (upper half filled square) and spinel lherzolites (lower half filled squares) from the Romanche fracture zone [7] are also plotted.

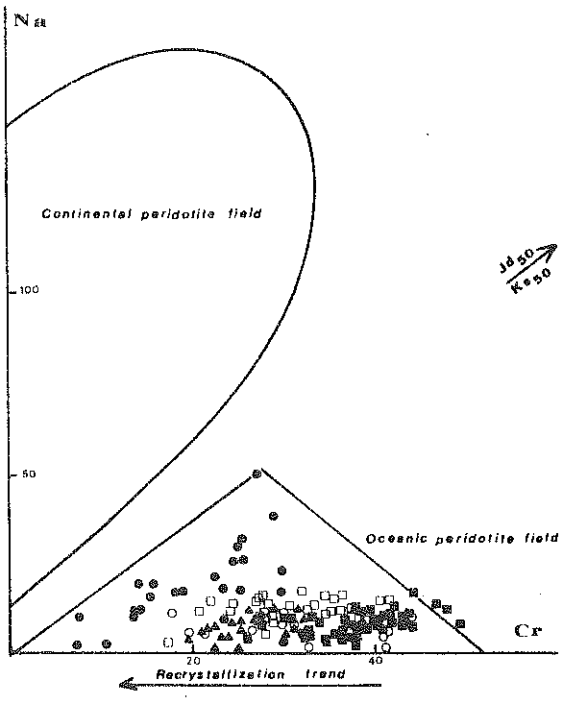

Fig. 2. Own data [36] for oceanic peridotites. Spinel lherzolites (open circle) and spinel harzburgites (open square) from DSDP Leg 45; spinel + plagioclase lherzolites from the Vema fracture zone (dots), spinel harzburgite (triangles) and spinel + plagioclase lherzolite (squares) from the Kane fracture zone. Note that the recrystallization trends (porphyroclast (p) to recrystallized grains (r) in Fig. 1 as well as core (c) to rim (r) variations indicated by arrows) have nearly constant Na and decreasing Cr and not Na depletion and Cr enrichment as expected by Sen [2].

data, they have been considered to be cumulates [12–15]. However, Strong and Jamieson [21] denied that the harzburgites might be related to mafic rocks by low-pressure crystal fractionation involving mainly olivine, plagioclase and clinopyroxene [22] and are better explained by a residual mantle hypothesis [21]. Consequently, it may be expected that residual mantle exists among analyzed ultramafites from Leg 37. In the Na-Cr diagram, clinopyroxenes from these rocks constitute less than 5% of the oceanic samples and fall in the same field (Figs. 1 and 2).

3. Overlap of Na-Cr clinopyroxenes from peridotites

Besides well admitted mantle processes such as partial fusion and reequilibration, it was shown that other petrogenetic effects may modify the Na and Cr repartition and account for some overlap. These last mechanisms were criticized by Sen [2]. To avoid discussion of some particular petrogenetic processes, Sen suggested comparison of only residual mantle peridotites from both continent and ocean equilibrated in the same facies and having "undergone similar degree of partial fusion". It appears unrealistic to assume that peridotites similar in bulk composition have suffered the same geochemical evolution in different areas. Nevertheless, some incriminated examples [2] may be further considered. At Ronda where geotectonic position is clear and recrystallization in the plagioclase field later than partial fusion [23,24], it is satisfactory to obtain an acceptable correlation for Na and Cr in clinopyroxenes, in agreement with the multivariant reactions involved [1]. Compared to other massifs, the reactions observed at Ronda [24,25] were more or less complete and equilibrium nearly reached. For the Lanzo peridotites [26] the interpretation of the Na-Cr plots is more debatable. The peridotites may result at least from partial fusion process, tectonic recrystallization and reequilibration in the plagioclase facies, all playing locally more or less simultaneously. It may be added that as for oceanic samples, clinopyroxenes from spinel- and plagioclase-lherzolites overlap at low Na and Cr content mostly in the oceanic field. In any case the discrimination observed for Lanzo may permit an oceanic origin not to be discarded.

4. Behaviour of Na and Cr in clinopyroxenes from nodules

Strictly speaking, the usefulness of the Na and Cr diagram to assign either an oceanic or a continental origin should be restricted to ultramafic bodies tectonically included in orogenic belts and not used for xenoliths. Consequently the main discussion raised by Sen [2] may concern only the second part of our paper [1] about the *behaviour* of clinopyroxenes from inclusions. The main criticism about this point made by Sen bears on the fact that clinopyroxenes from xenoliths in alkali basalts from ocean islands do not plot in the "sub-oceanic" field [2, fig. 1]. We strongly agree with this remark which was already noticed [1, fig. 8] and discussed in our own paper (p. 246). We concluded about xenoliths found in oceanic islands that "the existence of relatively Na-rich clinopyroxenes from such "sub-oceanic" peridotites may indicate that "sub-oceanic" mantle equilibrated in the spinel or garnet facies does not represent everyhwere a strongly depleted character". Na and Cr from xenoliths from both continents and oceanic islands fall in a more or less similar field covering the defined continental and oceanic areas of ultramafic bodies [1, p. 247]. Moreover, we agree with Sen emphasizing the role of spinel to plagioclase reequilibration and partial fusion processes to modify the Na and Cr content, a fact pointed out when we established the diagram (p. 242) for ultramafic bodies. A few detailed studies of ultramafic xenoliths from oceanic islands [27] prevent us from drawing a more consistent conclusion. Nevertheless, it can be noted that as for the continental alkaline districts, the oceanic islands belong to the so-called "within-plate volcanism" [28]; on the other hand, as frequently discussed [29], ultramafic xenoliths originated in the mantle are incorporated in alkaline volcanics and never in the associated tholeiitic or transitional basalts (which only contain low-pressure ultramafic cumulates [30]). Alkali basalts have very similar characteristics in con-

tinental and oceanic areas [31], their ultramafic xenoliths also present close similarities. In our paper, we omitted to point out more clearly this evidence which could in part be interpreted as the result of mantle contamination by percolating basaltic melts [1,32–35].

This discussion has reinforced our conviction that the Na-Cr plot of clinopyroxenes from *ultramafic bodies* has an important meaning for bulk-rock composition and geotectonic assignment of otherwise well studied Alpine ultramafic bodies. Sen [2] does not bring in his comments any useful information to disprove the interpretation given in our paper. We knew and produced relevant data to show that xenoliths in alkali basalts and kimberlites give additional problems. In this last case some of the suggested mantle processes may be discussed from each particular xenolith occurrence like those from Oahu and Kilbourne Hole [2]. We hope to be able to give more information later, after having worked on and thought carefully about some particular occurrences.

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PERIDOTITE AND GABBROIC STRUCTURES IN THE MONTE MAGGIORE MASSIF, ALPINE CORSICA[1]

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ABSTRACT

Penetratively deformed peridotites from the Monte Maggiore spinel/plagioclase lherzolite massif in Alpine Corsica are intruded by gabbroic dikes ranging in mineralogy from troctolites to Fe and Ti gabbros. The dikes cross-cut diffuse plagioclase segregations and veinlets which may represent trapped mafic melt within the residual peridotite. Elliptical shaped dunite pods with refractory compositions are often bounded by lace-like networks of plagioclase (An_{88}) which invade neighboring undepleted lherzolite. Lherzolite mineral compositions range from olivine Fo_{90-91}, orthopyroxene $Ca_{1.8}$ $Mg_{91.3}$ $Fe_{4.8}$ to $Ca_{46.7}$ $Mg_{90.1}$ $Fe_{9.5}$, and clinopyroxene $Ca_{45.8}$ $Mg_{91.3}$ $Fe_{4.6}$ to $Ca_{46.7}$ $Mg_{90.1}$ $Fe_{5.2}$. Peridotite and gabbroic structures and textures suggest that the emplacement of the dikes was contemporaneous with the high temperature deformation of the peridotite, occurring under upper mantle or deep crustal conditions. Locally the intrusion of the dike magma followed preferential shear fracture planes related to flow structures in the enclosing peridotite. The similarity of differentiation patterns in the gabbroic dikes with gabbroic cumulates in the overlying Corsican ophiolitic section suggests that the Mt. Maggiore dikes may have fed these magma chambers.

INTRODUCTION

The study of ultramafic and mafic dikes in peridotites from alpine-type bodies and ophiolite complexes has provided a great deal of information about the processes of differentiation and emplacement of these rocks, now widely considered to be of mantle origin (Hess 1965; Coleman 1971). In particular, ophiolite complexes containing a downward stratigraphic succession of pillow lavas—sheeted dike complex—gabbroic and ultramafic cumulates resting on a tectonized peridotite floor (Moores and Vine 1971) often provide easy access to the observation of partial melt products and processes.

Dike injection into the oceanic crust of a spreading plate is fed by underlying mafic magma chambers continually replenished by partial melting processes within the upper mantle. The foliated or metamorphic peridotites forming the base of an ophiolitic section represent mantle rocks from which most of this melt fraction has been extracted, leaving behind depleted lherzolites (ol + cpx + opx + sp/plag) and harzburgites (ol + opx + sp/plag) (Boudier and Nicolas 1972; Himmelberg and Loney 1973; Menzies and Allen 1974). They may enclose partial melting products such as foliated and non-foliated patches of dunite, pyroxenite veins and banding, gabbroic dikes and plagioclase segregations. In Mediterranean ophiolite, plagioclase tectonized peridotite facies have been described at Lanzo, Italy (Boudier and Nicolas 1972), Makirrakhi, Greece (Menzies 1973), Othris, northern Greece (Menzies and Allen 1974), and Troodos, Cyprus (Menzies and Allen 1974; George 1978).

In the Mt. Maggiore peridotite massif (2.5 × 1.5 km), northeastern Alpine Corsica, weakly tectonized spinel plagioclase lherzolite containing diffuse olivine websterite banding is cross-cut by gabbroic dikes of variable composition. Through a study of dike structures and orientation, variable rock compositions, textural relations, petrofabric, and strain analysis, this paper investigates the deformation of the Mt. Maggiore peridotite body to determine the chronology and injec-

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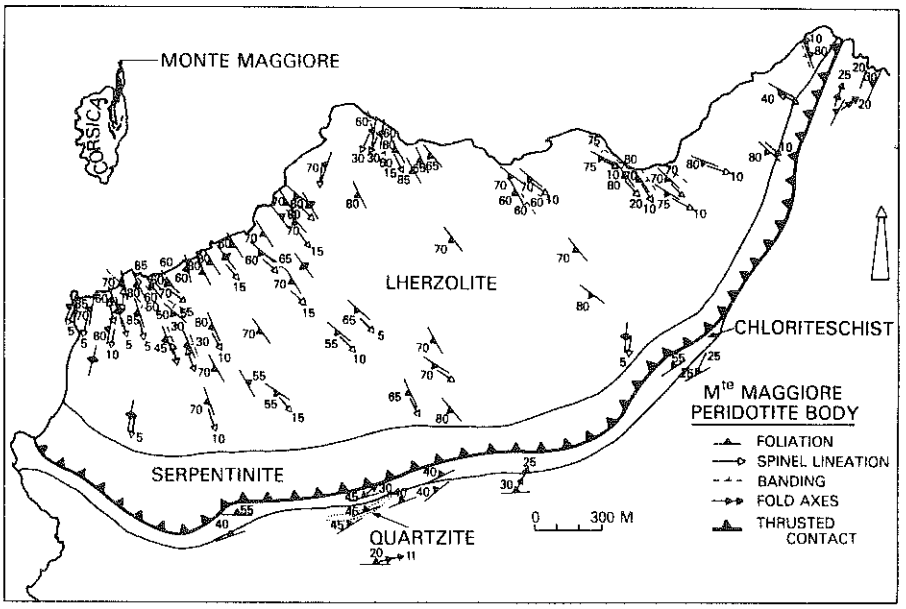

FIG. 1.—Monte Maggiore massif. Tectonic foliation plane and pyroxenite banding.

tion mechanism of the cross-cutting dikes. We also present new geochemical data on the lherzolite, associated dunites, and ultramafic banding as well as the gabbroic dikes.

We find that the peridotite deformed by left-lateral simple shear at high temperature (>1100°C) and that the gabbroic dikes injected along planes oblique to the foliation, suggesting that their emplacement was contemporaneous with the solid state deformation. Hydrostatic pressures of less than 9 Kb must have existed in the peridotite during this time since spinel grains are rimmed with plagioclase and in-situ collections of plagioclase melt (lherzolite dry solidus = 1200°C) are cut by the intrusive gabbroic dikes. Bulk rock as well as mineral compositions are similar to other residual oceanic peridotites. Tectonic fabrics in the peridotite, nearly identical to weakly tectonized spinel lherzolites drilled from the western margin of the mid-Atlantic ridge (Boudier 1978*b*), suggest a deformation by high temperature plastic flow near a spreading ridge (Nicolas et al. 1980).

REGIONAL GEOLOGY

Paleogeographic reconstructions suggest that the Corso-Sardinian block, now trending NS in the western Mediterranean area, may have originally been attached to the Liguro-Appennine zone before rotation into its present position during the Eocene (Alvarez 1972). In this case, the Corsican, as well as the western Alpine and Appennine ophiolites, may be considered as fragments of oceanic crust, created during Jurassic time in the Liguro-Piedmont oceanic basin (Dal Piaz 1974), which separated the paleo-European and South-Alpine-Appennine continental blocks. Geochemical features and phase assemblages of the Mt. Maggiore peridotite and its associated gabbroic dikes are similar to both those of oceanic cored and dredged samples and to other lherzolite massifs of the western Alpine system (Boudier 1978*a*; Ernst and Piccardo 1979). These lherzolite bodies, often associated with ophiolitic gabbros and lavas, contrast sharply with the more depleted harzburgite massifs usually found at

the base of the eastern Mediterranean ophiolites (Nicolas and Jackson 1972; Menzies and Allen 1974; Rocci et al. 1975).

The tectonized, mainly lherzolitic peridotites of the Mt. Maggiore massif represent the deepest part of a reconstructed ophiolitic sequence outcropping in eastern Alpine Corsica (Ohnenstetter et al. 1975). Within the ophiolite, troctolites associated with dunites and anorthosites, olivine-gabbros, gabbros, noritic gabbros, ferrogabbros, ferrodiorites, and albitites form disrupted members of low pressure magma chambers. Associated extrusive liquids are represented by massive dolerites and pillow lavas. Petrological and geochemical studies (Ohnenstetter et al. 1975; Ohnenstetter and Ohnenstetter 1975) have revealed close similarities between the Corsican ophiolitic rocks and samples dredged or cored from mid-oceanic ridges and transform fault zones (Ohnenstetter and Ohnenstetter 1975; Beccaluva et al. 1977; Sinton 1978).

The ophiolite and its late Jurassic and early Cretaceous *schistes lustrés* cover have been deformed and metamorphosed during several tectonic phases. First, westward thrusting of the ophiolitic block onto crystalline basement and Mesozoic sediments occurred during the upper Cretaceous (Mattauer and Proust 1976). Mattauer et al. (1977) have related this thrusting to subduction and subsequent blockage of the sialic block at depth, with accompanying obduction of the *schistes lustrés*-ophiolite nappe. A zone of intense plastic deformation marks the actual thrust contact with the basement rocks and cuts NNW through the island of Corsica. Later tectonism, particularly during Eocene time, has also caused numerous displacements, sometimes with an eastern vergence associated with folding.

Three successive metamorphic phases have been recognized in the Corsican ophiolite. First, greenschist and amphibolite facies metamorphism developed during the formation of an ocean in Jurassic time. Then, high pressure-low temperature metamorphism, most often considered to be upper Cretaceous in age (Ohnenstetter et al. 1976), accompanied the emplacement of the thrusted ophiolite slabs. Later greenschist metamorphism may be associated with alpine tectonism during Eocene and Oligocene time (Ohnenstetter et al. 1976). The inferred simultaneity of tectonic and metamorphic events must be tempered by the fact that some important tectonic movements may have occurred after the latest metamorphic phases (Amaudric du Chaffaut and Saliot 1979). In this regard, the peridotites of the Mt. Maggiore massif from the northwestern part of the Cap Corse peninsula are reasonably fresh and show only minor late stage alteration.

A NNE dipping thrust fault of Ligurian origin (Durand-Delga 1978) underlies the Mt. Maggiore massif and placed the already cooled lherzolite sheet into tectonic contact with meta-sedimentary rocks of the Jurassic and Cretaceous *schistes lustrés* series of Eastern Alpine Corsica. The massif is thrust against the right-way-up limb of a recumbent fold (Guillou 1962; Guillou et al. 1962) composed of chlorite and albite mica schists showing no thermal aureole related to the peridotite. A narrow band of intensely deformed schist (chlorite + quartz + epidote) marks the fault contact and outcrops at the base of a zone of serpentinized lherzolite, surrounding the fresher peridotite of the interior of the massif. Twenty meters below the peridotite contact a quartzite band outcrops within the chlorite schist country rocks adjoining the massif. A petrofabric study of the deformed quartz from this band shows a preferred orientation of C-axes by low temperature basal slip (J. L. Bouchez, pers. comm.) (fig. 2) with a sense of shear from east to west. These transversal microstructures, as well as elongation lineations from the neighboring schists, are similar to those observed by Mattauer et al. (1977) for the whole of Alpine Corsica and delineate a westerly movement of the *schistes lustrés*-ophiolite nappe.

FIELD RELATIONSHIPS AND PETROGRAPHY

The Lherzolite.—Within the northern, more central area of the Mt. Maggiore massif, relatively fresh, penetratively deformed spinel

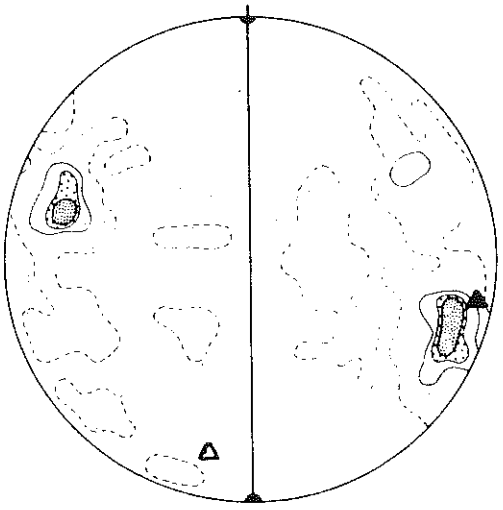

FIG. 2.—Preferred orientations of quartz C-axes from quartzite band adjoining the thrust fault contact of the massif. Lower hemisphere projection. 125 measures. Contours 1%, 2%, 4%, 8%. Solid line: foliation. Dots: lineation. Black triangle: computed point maximum. Open triangle: computed pole to best computed plane.

lherzolite encloses a variety of non-penetrative features. These include diffuse pyroxenite banding, ellipsoidal dunite pods, and cross-cutting gabbroic dikes of variable composition.

In the field, the peridotite displays an S_1 foliation plane, defined as the plane of mineral flattening and a lineation, L_1, which is the mineral elongation direction contained in the foliation plane. These structures are best marked in the massif by elongate spindles of spinel, often rimmed with plagioclase on the slopes of Mt. Maggiore. Foliations and lineations are quite homogeneous in orientation through the massif, with an average S_1 orientation striking 120°, dipping 60° to the west; lineations are sub-horizontal (fig. 1).

The lherzolite (fig. 3) exhibits coarse-grained protogranular texture (Mercier and Nicolas 1975; Harte 1977) with an average olivine and enstatite grain size of 2.0–5.0 mm. Enstatite is characteristically associated with green Cr-diopside and brown chromium spinel. Boundaries between grains are straight or smoothly curved, and recrystallized neoblasts are absent. Within the plagioclase lherzolite facies, interstitial pockets of plagioclase gather at olivine grain boundary junctions similar to plagioclase lherzolites described at Lanzo, Italy (Boudier 1978*a*) and Othris, northern Greece (Menzies and Allen 1974). Olivine grains show some evidence of strain by the presence of relatively sharp (100) olivine subboundaries lying normal to the trace of the foliation plane. Rotations from one subgrain to another are small, from 2° to 5°. Olivine and enstatite kink band boundaries are rare in the massif lherzolite. Olivine shows a weak lattice preferred orientation (fig. 4) with a [010] girdle at a high angle to the foliation plane and [100] grouped around the L_1 lineation direction. This fabric pattern has been interpreted as the result of high temperature (010) [100] and {0kl} [100] olivine intracrystalline glide (Nicolas et al. 1973). A very slight sinistral obliquity can be noted between the lattice and shape fabrics in olivine and suggests that the flow may have been rotational. The shear sense of this rotational flow can be observed in any XZ type thin section of peridotite, cut normal to the foliation and parallel to the lineation, where the approximate flow direction is given by the average extinction direction under crossed polarizers (Boudier 1978*a*). In 13 out of 14 such measures, the Mt. Maggiore lherzolite exhibits a sinistral sense of shear (fig. 5).

Pyroxenite Banding.—A diffuse, coarse-grained banding composed of olivine websterite and pyroxenite, most often rich in orthopyroxene, outcrops throughout the massif with an orientation closely concordant with that of the S_1 tectonic foliation plane (fig. 1). The banding forms 5–10 cm thick segregations of websterite alternating with dunite and pyroxene-poor lherzolites. Typically, these structures are only continuous on the outcrop scale; zones of alternating pyroxenite-dunite bands may reach thicknesses of several meters.

Suggestions of isoclinal folding (fig. 6) and boudinage paralleling the S_1 foliation plane may help explain the diffuse, discontinuous nature of the bands. Fold axes, where ob-

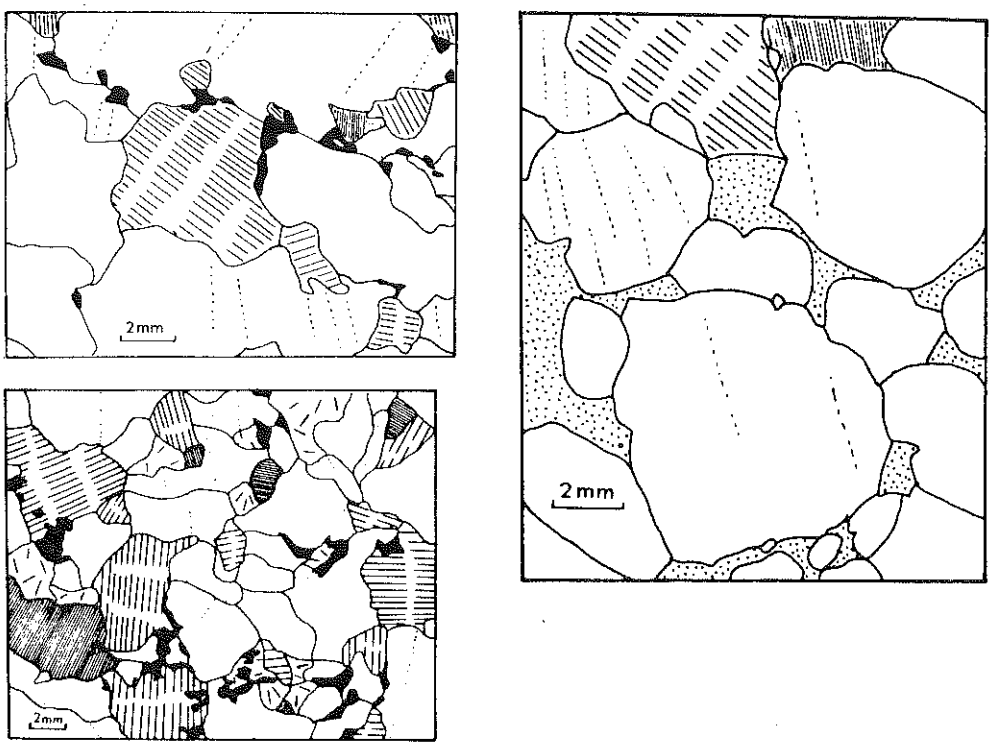

FIG. 3.—Peridotite microstructures. Drawings after thin sections cut perpendicular to the foliation and parallel to the lineation (XZ plane). Foliation trace EW. Olivine: unpatterned (dotted lines are sub-boundaries). Orthopyroxene: unpatterned (when parallel, dashes correspond to trace of (100) plane). Clinopyroxene: thin lined areas. Plagioclase: dots. Spinel: black. *A*. Clinopyroxene-poor spinel lherzolite. Sample 78CC43. Protogranular texture. *B*. Plagioclase lherzolite. Sample 78CC35. Protogranular texture. *C*. Contact of websterite banding with enclosing dunite. Sample 79CC2. Bars are 2 mm.

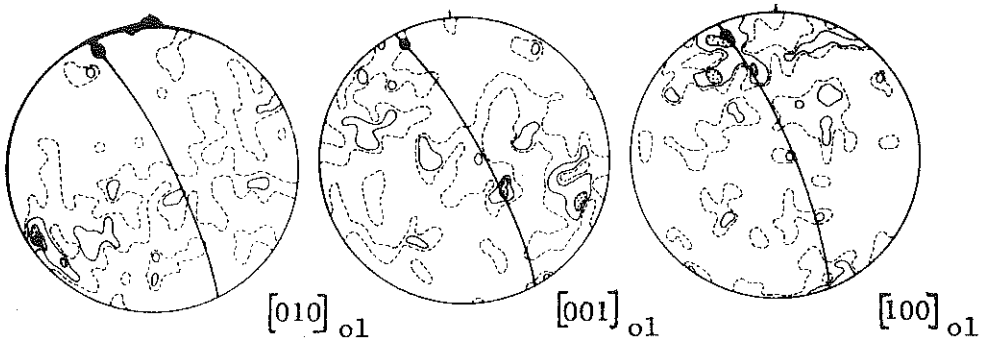

FIG. 4.—Olivine lattice preferred orientations in very weakly deformed protogranular lherzolite. Lower hemisphere projection. Contours 1%, 2%, 4%, 5%. Solid line: foliation. Dots: lineation. Sample 77CC1. 100 measures.

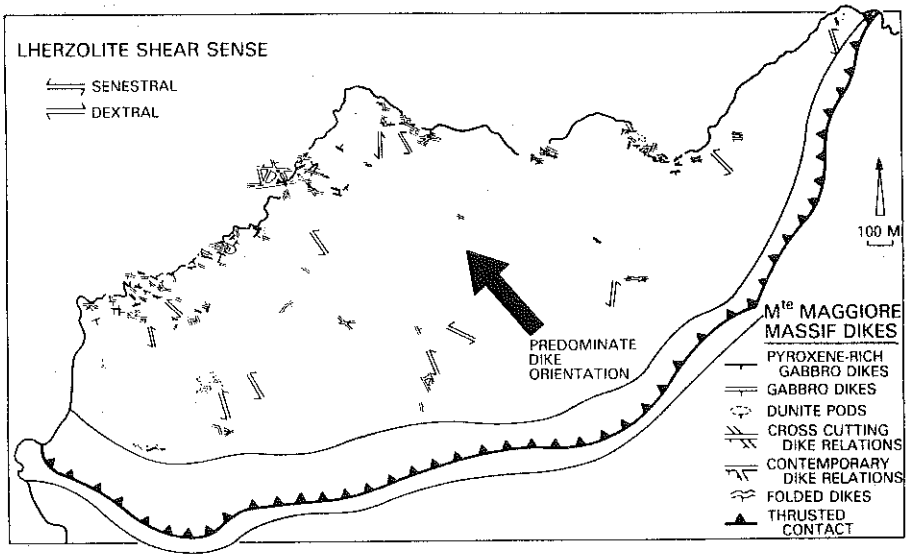

FIG. 5.—Monte Maggiore massif. Gabbroic dikes and lherzolite shear sense.

served, plunge about 20–25° to the south, parallel to the lineation direction in the lherzolite.

Microstructures in the websterite banding are identical to those in the neighboring lherzolite and also display coarse protogranular textures. The contact between the banding and the enclosing dunite (fig. 3) shows only a progressive increase in orthopyroxene content. Within the banding, large (0.5–1.0 cm), generally unstrained orthopyroxene coexists with varying proportions of diopside. Both show (100) exsolution lamellae. Red brown chromium spinel occurs most often at orthopyroxene grain boundaries; plagioclase is occasionally exsolved from pyroxene.

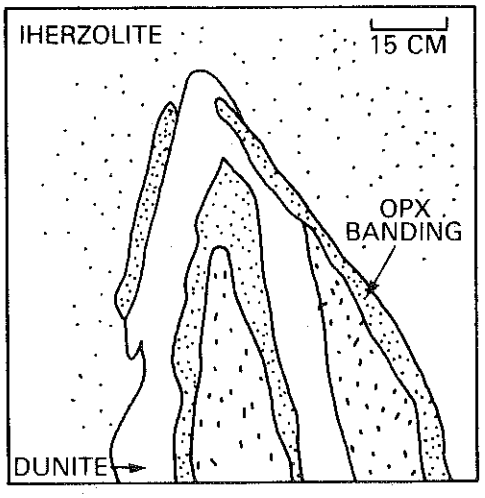

FIG. 6.—Folded olivine websterite banding, axial planar with peridotite foliation. Drawing after photograph.

Dunites.—Dunite in the massif outcrops either as thin bands 5–10 cm thick associated with the websterite banding or as ellipsoidal patches up to 3 m in diameter and sometimes rimmed with plagioclase before being surrounded by normal lherzolite (fig. 7). The foliation in the dunites parallels that of the neighboring lherzolite and is marked by aggregates of spinel, often rimmed with plagioclase which may also form small (0.2–4.0 cm) discontinuous, criss-crossing veinlets, oblique to the foliation within the dunite. Sharp-walled gabbroic dikes cut across the dunite pods and continue along into neighboring undepleted lherzolite. In one specific pod, a gabbro dike appears to be rooted within the dunite.

Dunite microstructures exhibit the same

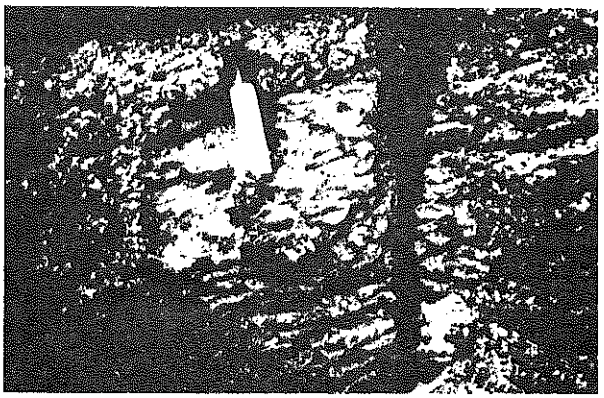

FIG. 7.—Dunite pod surrounded by coarse grained protogranular lherzolite. Plagioclase segregations form a lace-like network around the border of the pod.

weakly developed olivine substructure as that in the lherzolite. Chromium spinel, sometimes associated with plagioclase, occurs in interstitial aggregates between 2–3 cm large olivine grains. When present, plagioclase is unzoned and xenomorphic, showing undulatory extinction.

Plagioclase Segregations.—Plagioclase lherzolite outcropping on the NW slopes of Mt. Maggiore frequently contains millimetric sized segregations of plagioclase, probably generated during partial melting and incompletely extracted from the peridotite. Analogous textures have been described and interpreted at Lanzo, Italy (Boudier 1972), Makirrakhi, Greece (Menzies and Allen 1974) and the Josephine Peridotite, Oregon (Dick 1977*b*). Orientations of the plagioclase blebs within the host lherzolite are random, probably due to the undeformed nature of the peridotite which failed to align the plagioclase accumulations by subsequent incremental strain (Boudier and Nicolas 1977).

Occasionally plagioclase blebs associated with the pyroxenite banding form thin (1–2 cm) lace-like networks parallel to it. A similar plagioclase-bearing ariegite banding has been described by Boudier (1972, 1978*a*) at the Lanzo massif, Italy. More commonly, thin, fuzzy-walled discontinuous bands of plagioclase lie oblique to the foliation with no particular preferred orientation. Larger intrusive-type gabbroic dikes cut sharply through these discontinuous veinlets.

Gabbro dikes.—Gabbroic dikes of varying composition, thickness, orientation, and internal structure compose about 2% of the massif's surface area (fig. 5). The dikes have sharp contacts against undepleted lherzolite, although rarely 3–5 cm thick walls of lherzolite depleted in pyroxene (= dunite) may enclose the dikes (Boudier 1972; Dick 1977*b*).

The dikes range in composition from olivine gabbros to oxide-rich gabbros and diorites, giving troctolites, olivine gabbros, ferro gabbros, Fe and Ti gabbros, and ferrodiorites. Two broad dike compositional groups, varying from olivine gabbros to Fe-Ti gabbros, are defined on the basis of varying proportions of olivine or oxide minerals. Olivine and plagioclase have been altered by both the greenschist facies and the high pressure facies of alpine metamorphism.

Within the olivine gabbro group, variations in the proportion of plagioclase and olivine produce troctolites, plagioclasites, and olivine gabbros, often zoned within a single composite dike in bands parallel to the dike walls. In the Fe and Ti gabbros, apatite bearing leuco-ferrogabbros and ferro diorites sometimes coexist with pyroxene-rich ferrogabbros. Intersections between dikes of the two compositional groups show no consistent pattern of cross-cutting age relations,

suggesting that the dikes are effectively contemporaneous in age (Escher et al. 1976).

Dike thicknesses range from centimetric to metric with grain sizes and internal structure varying accordingly. Thicker dikes tend to be composite type dikes, exhibiting extreme variations in grain size with facies changes usually zoned with bilateral symmetry parallel to the dike walls. Coarse-grained dikes often have a pegmatitic facies with pyroxene (5.0 × 3.0 cm) and plagioclase laths oriented subperpendicularly to the dike wall. Many dikes exhibit tectonic flaser structures with sheared, deformed zones (fig. 8) where pegmatitic clinopyroxenes, initially oriented perpendicular to the dike walls, have been bent and broken parallel to the dike walls. In thin section the relict pyroxene porphyroclasts are surrounded by aggregates of tiny, rounded pyroxene fragments.

GEOCHEMICAL ANALYSES AND INTERPRETATIONS

Whole-rock chemical analyses of the Mt. Maggiore peridotites (table 1; dunite, harzburgite, and lherzolite) are low in Al_2O_3, CaO, and alkali oxides, with Mg/Mg + Fe ratios of .897–.886. These refractory compositions, as well as the presence of five mineral phases (ol + opx + cpx + plag + sp) in the lherzolite agree with Menzies's (1973) criteria for distinguishing primary tectonite lherzolite from highly deformed cumulate ultramafic. We interpret the Mt. Maggiore peridotite as depleted upper mantle, residual to partial melting of aluminous upper mantle peridotite (O'Hara 1970) that has retained part of its generated basaltic melt as plagioclase segregations (Himmelberg and Loney 1973). In addition, tectonized peridotite fabrics support the interpretation of the Mt. Maggiore massif as the residual, metamorphic base of the ophiolitic section in Corsica since none of the overlying cumulate rocks, representing ultramafic magma chambers, are deformed by solid flow.

Peridotite and gabbroic whole-rock analyses from the Mt. Maggiore massif and from the overlying Corsican ophiolitic section (Ohnenstetter and Ohnenstetter 1975) are shown by figure 9. Peridotites containing varying proportions of clinopyroxene fall on a line joining lherzolites to pyrolite (Green and Ringwood 1967) and to dunite; plagioclase peridotites from the massif plot above it. The extension of this line passes near the mean

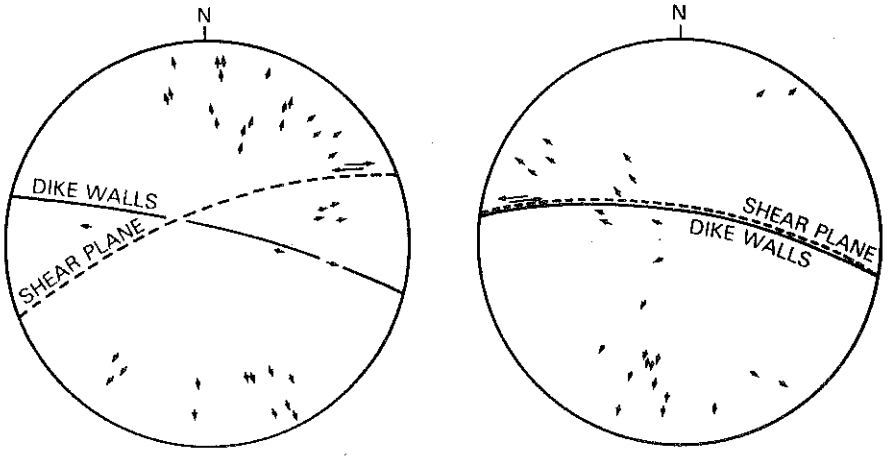

FIG. 8.—Stereograms of gabbroic flaser structures, showing deformation by sinistral or dextral simple shear. Lower hemisphere projection. Solid line: orientation of dike walls. Dashed line: orientation of shear planes. Arrows: lineation defined by clinopyroxene orientations.

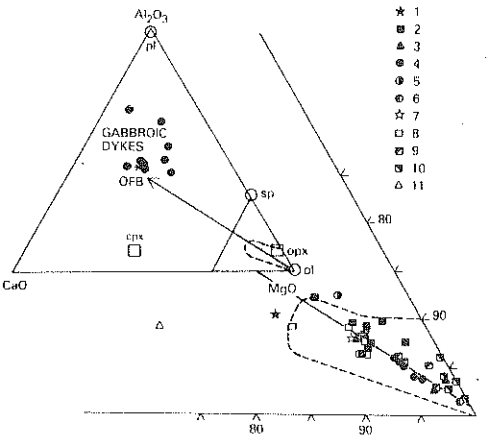

FIG. 9.—Al_2O_3-CaO-MgO triangle. (1) through (6) are peridotites from this paper: (1) pyroxenite; (2) lherzolite; (3) cpx-poor lherzolite; (4) harzburgite; (5) plagiferous dunite; (6) dunite; (7) three pyrolites (Green and Ringwood 1975). (8) through (11) are Corsican peridotites (Ohnenstetter and Ohnenstetter 1975): (8) plagioclase lherzolite; (9) cpx-poor lherzolite; (10) harzburgite and dunite; (11) olivine websterite. Dashed line: compositional field of oceanic peridotites (after Ohnenstetter and Ohnenstetter 1975). OFB: mean of ocean floor basalts (after Thompson 1973).

composition of oceanic gabbros (Thompson 1973) and the Mt. Maggiore gabbros. The fields occupied by the Corsican peridotites are similar to those defined by oceanic samples (Ohnenstetter et al. 1975) and also to that of other Mediterranean ophiolites (Nicolas and Jackson 1972; Menzies and Allen 1974).

Peridotite and gabbroic whole-rock compositions are compared with other western Mediterranean peridotite/gabbro assemblages in the Na_2O+K_2O-MgO-FeO triangle of figure 10. Refractory ultramafic compositions cluster on the MgO-FeO join, and the dikes occupy the field of oceanic samples with iron- and titanium-rich gabbros from Mt. Maggiore plotting at the inflection point of the curve. A leuco-ferrodiorite dike from the massif plots toward the alkaline apex. This differentiation pattern in the dikes is similar to the chemical evolution of the gabbroic cumulate rocks located within the overlying cumulate section (Ohnenstetter et al. 1975) and suggests that the gabbroic dikes may have fed these magma chambers.

Peridotite Mineralogy.—Microprobe analyses of minerals from peridotites and gabbroic dikes were obtained using the electron microprobe Cambax at the University of Nancy I, employing an acceleration voltage of 15 kv, a beam current of 8 nA, a beam diameter interior to 3 Mm, and a six second counting time for a group of three elements. Corrections were obtained through the MBX COR after Henoc and Tong (1978). There is no mineral zoning in these rocks and thus little variation in the composition of the constituent phases.

Plagioclase does not occur in all the peridotites but, when present, is often transformed to grossular, zoisite, prehnite, and zeolite. The pure jadeite found in some samples suggests that the peridotite was affected by the high pressure phase of alpine metamorphism. Interstitial plagioclase within dunites is An 88, while plagioclase from the gabbroic dikes is much more sodic (An 54). In all the peridotites, including the dunites and pyroxenite banding, olivine has a uniform forsterite content (100 × Mg/Mg + Fe^{++} + Mn) of Fo_{90-91} with nearly invariant nickel, calcium, and manganese contents. In the gabbroic dikes, olivine (Fo_{85}) lies within the same composition range (Fo_{90-95}) defined for olivines in troctolites from the overlying cumulate sequence (Ohnenstetter and Ohnenstetter 1975). Compositions of primary enstatite and Cr-diopside pyroxenes with exsolution lamellae are illustrated in the pyroxene quadrilateral of figure 11. The magnesium contents in the pyroxenes vary little with lithology, are typical of pyroxenes from alpine peridotites, and range from orthopyroxene $Ca_{1.8}$ $Mg_{90.7}$ $Fe_{8.9}$ to $Ca_{2.2}$ $Mg_{90.1}$ $Fe_{9.5}$ and clinopyroxene $Ca_{45.8}$ $Mg_{91.3}$ $Fe_{4.6}$ to $Ca_{46.7}$ $Mg_{90.1}$ $Fe_{5.2}$. Aluminum contents are low and typical of values reported from the western Mediterranean area (Boudier 1976; Cottin 1978; Ernst 1979; Ernst and Piccardo 1979). Chrome content is greatest in diopside from the most refractory lherzolites; clinopyroxene is enriched in Al, Ti, and Cr

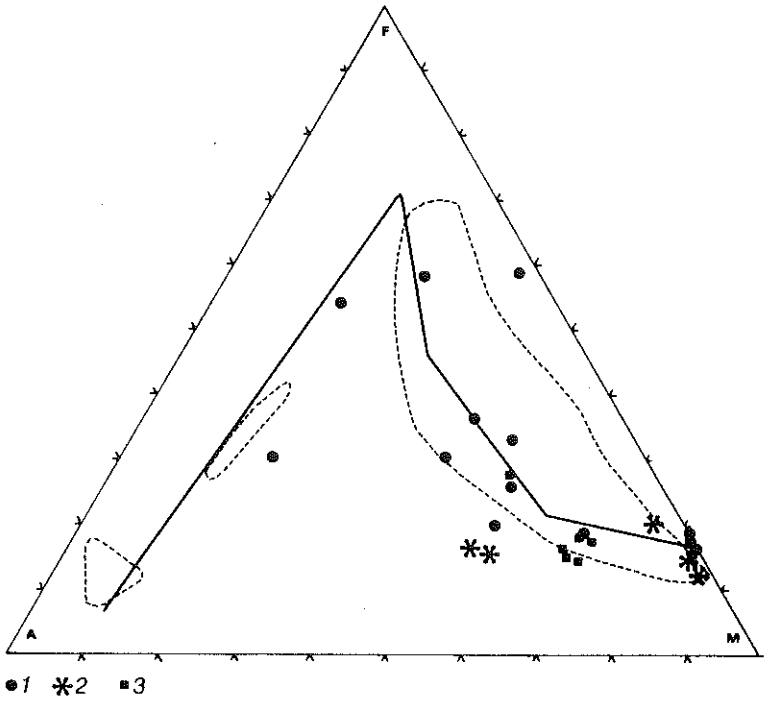

FIG. 10.— **A M F** diagram showing bulk chemical analyses of: (1) peridotites and gabbroic dikes of the Mt. Maggiore massif; (2) gabbroic rocks from the Bracco Zone (after Cottin 1977); (3) gabbroic dikes from the Lanzo massif (after Boudier 1976). Thick line: Corsican cumulate trend (after Ohnenstetter et al. 1975). Dashed line: compositional fields of oceanic plutonic rocks (M. Ohnenstetter, unpub. data).

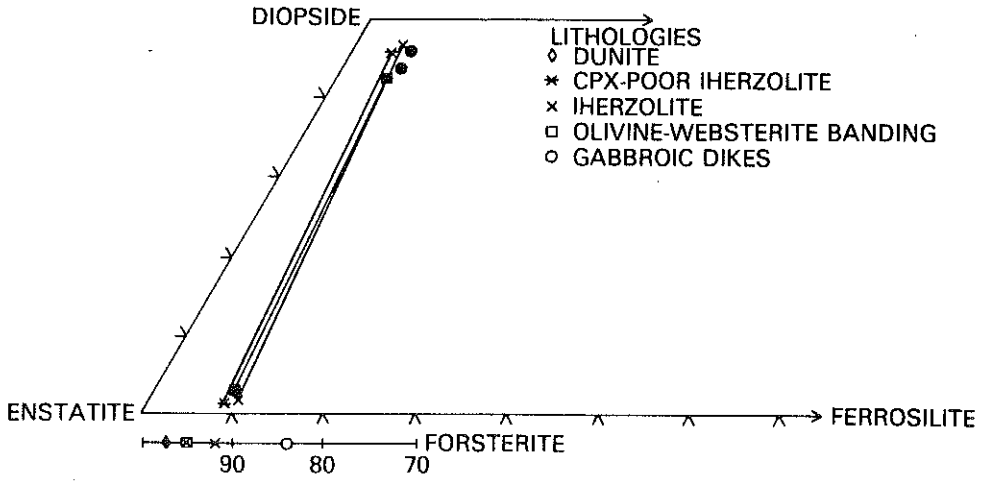

FIG. 11.—Plot of orthopyroxene, clinopyroxene, and olivine compositions of the Mt. Maggiore massif.

compared with coexisting orthopyroxene. Diopside from the gabbroic dikes is greatly enriched in Ti.

Unzoned brown chrome spinel, sometimes in simplectite form interfingered with orthopyroxene, occurs throughout the massif peridotite. Although chrome spinel is absent in the dikes, magnetite and ilmenite occur in the most differentiated Fe and Ti gabbros. The composition of the spinel varies with lithology, so that in the lherzolites and pyroxenite banding spinels are rich in chromium and in the dunites and harzburgites (which may contain relict Cr-diopside) the spinel is slightly richer in Al and Mg (fig. 12).

Dunites.—Elliptically shaped patches of dunite, up to 3 m in diameter, contain anhedral megacrysts of picotite spinel, often rimmed with plagioclase and elongated parallel to the foliation in the surrounding lherzolite. Mg/Mg + Fe ratios in the dunites range from 0.896–0.890, compared to about .885 in the lherzolite. If the dunites resulted from localized melting of the lherzolite and accompanying depletion in plagioclase and pyroxene (Boudier and Nicolas 1972), then they should have higher contents of Cr, Ni, and Mg and lower Ti, Fe, Mn, than the surrounding lherzolite. Whole rock chemical analyses (table 1) confirm these trends. Microprobe analyses of spinel (fig. 12) show more refractory Mg/Mg + Fe contents in the dunites but do not show a strong depletion in Al_2O_3 in spinel from plagioclase dunite. However, the networks of unzoned plagioclase which invade the undepleted lherzolite surrounding the dunite pods (fig. 7), as well as their more refractory compositions, strongly suggest a residual origin by in-situ partial melting of these rocks.

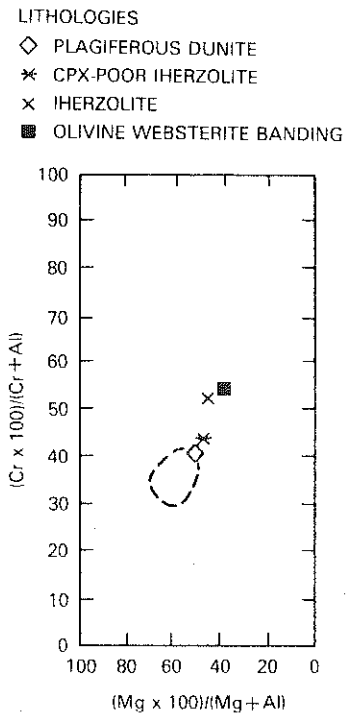

FIG. 12.—Plot of chrome spinel compositions from Mt. Maggiore peridotites showing the variation of major components. Dashed line: field of cessory chrome spinel in residual harzburgites from Canyon Mountain, Oregon (Dungan and Ave Lallemant 1977; Himmelberg and Loney 1980). Chrome spinel from metacumulate dunites at Canyon Mountain plot above this field.

Large dunite patches and pods containing euhedral chromite have been described in ophiolitic harzburgites at, for example, Troodos, Greece (Menzies and Allen 1974), Josephine peridotite, Oregon (Dick 1977) and Canyon Mountain, Oregon (Himmelberg and Loney 1980). Because of their nonrefractory chemistry, relict cumulate textures, and layering, they have been interpreted as the result of olivine and/or spinel crystal settling from pockets of basaltic magma at a spreading ridge (Menzies and Allen 1974) which have been, perhaps, infolded with residual harzburgites (Himmelberg and Loney 1980). Field and chemical evidence from Mt. Maggiore, however, indicates that these dunites differ from those previously mentioned and originated by processes of localized partial melting, similar to dunites in the Lanzo peridotite, Italy (Boudier and Nicolas 1972).

GABBROIC DIKE STRUCTURES

Intrusive gabbroic dikes tend to lie in one predominate orientation, trending 130–165° and dipping steeply to the east (fig. 5). When

TABLE 1

COMPOSITIONS OF PERIDOTITES AND GABBROS FROM THE MT. MAGGIORE MASSIF

	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)
SiO_2	38.44	41.16	42.30	47.08	49.12	48.41	47.75	53.45	36.04
TiO_2	.02	.05	.09	.12	.16	.19	1.22	1.95	4.15
Al_2O_3	2.70	2.00	2.61	4.03	16.71	17.99	15.67	20.72	18.15
Fe_2O_3	9.14	8.94	8.86	7.45	4.49	4.55	7.50	4.28	13.56
MnO	.14	.14	.14	.12	.05	.10	.10	.08	.23
MgO	38.98	37.74	35.37	29.92	15.03	10.54	9.73	2.59	7.99
CaO	1.54	1.69	2.05	5.09	9.72	10.01	9.73	7.82	15.62
Na_2O	.24	.07	.11	.08	3.51	3.75	4.45	6.12	.40
K_2O	.02	.02	.03	.04	.08	.05	.07	.08	.05
P.F.	6.75	7.46	7.81	5.53	6.14	3.89	3.48	2.78	3.45
Ni ppm	2235	1957		1850		410	140		280
Cr	1640	1922		2870		870	140		
Total	98.27	98.73	99.37	99.46	100.01	99.48	99.70	99.87	99.70
Mg/Mg + Fe + Mn	89.38	89.31	88.76	88.83	86.89	81.96	74.30	54.51	53.85

NOTE.—Bulk chemical analyses of peridotites and gabbroic dikes from Mt. Maggiore: (1) average of 2 dunites; (2) average of 7 harzburgites and cpx poor lherzolites; (3) average of 2 lherzolites; (4) websterite banding; (5) troctolite dike; (6) average of 3 olivine gabbro dikes; (7) leucogabbro dike; (8) and (9) Fe and Ti gabbro dikes.

plotted in stereographic projection (fig. 13) against a structural reference of the foliation oriented vertically NS and the lineation oriented horizontally within this plane, poles to the principal family of dike planes form a partial girdle with a point maxima lying about 45° oblique to the foliation. Commonly, the dikes outcrop as complex, synchronously injected ensembles which deviate in orientation from the more linear dike planes (fig. 5). Individual dikes may be perpendicular to each other locally (fig. 14) and might have formed as fillings of conjugate fractures, more or less symmetric about the foliation plane.

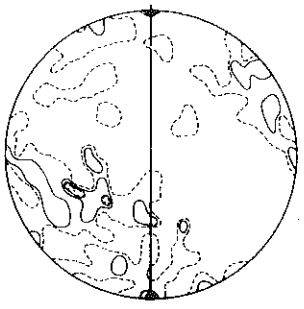

FIG. 13.—Orientations of linear gabbroic dikes in a structural reference system of the foliation vertical, NS and the lineation horizontal, NS (dots). This plane corresponds to the XZ plane of the infinitesimal strain ellipse. Contours 1%, 2%, 4%. 102 measures.

Folding.—The olivine websterite banding has rarely been folded into outcrop sized, diffuse isoclinal folds (fig. 6) with fold axes plunging subhorizontally, parallel to the mineral lineation in the enclosing lherzolite. Otherwise the gabbroic dikes are neither folded nor boudinaged, with the exception of two localities where the dikes are gently warped by concentric folds with steeply plunging axes.

Structural Analysis.—Analysis of the geometry and chronology of the gabbroic dike injection presents some difficulties. In the peridotite, the presence of a homogeneous, penetrative foliation accompanied by high-temperature olivine deformation microstructures and the parallelism of mineral lineations with the axes of small scale folds in the websterite banding indicate that solid, banded peridotite underwent a weak but pervasive high temperature (>1100°C) and

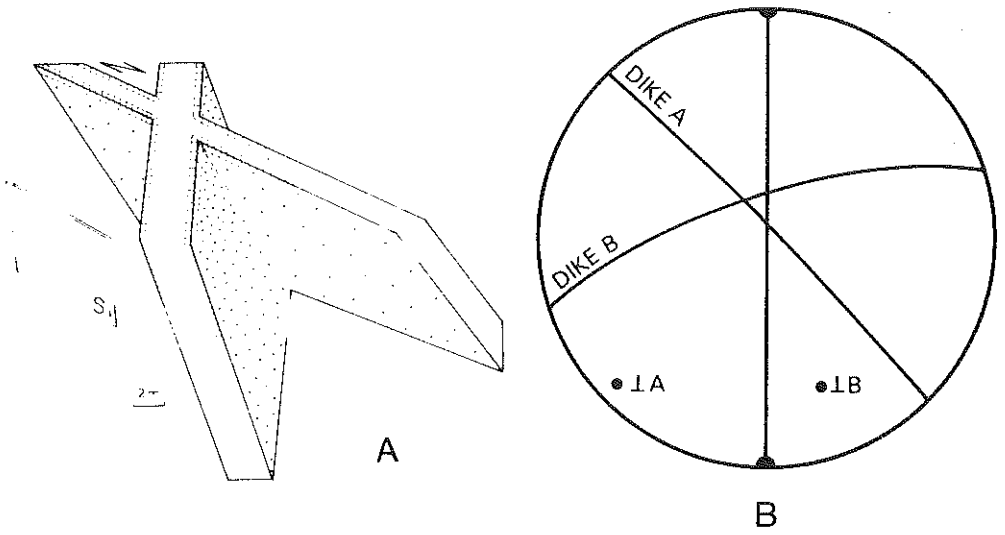

FIG. 14.—Conjugate gabbroic dikes drawn from outcrop. *A:* Block diagram of field orientations. *B:* Stereogram with structural reference system of the foliation vertical, NS, and the lineation NS. A continuous 2 cm-wide band of clinopyroxene forms an internal border around the dikes' intersection.

pressure (≈ 9 kb) deformation accompanied by in situ plagioclase melting. If gabbroic dike injection postdated cooling and cessation of solid state plastic flow in the peridotite, the dikes should show no folding, boudinage, or other geometric relation to this deformation. Raleigh (1965) has shown that pyroxenite veins in the Cypress Island alpine peridotite, Washington, USA, did not participate in the penetrative deformation which folded the chromitite layering in the peridotite, and that fold axes in the veins bore no relation to fold axes in the peridotite. He concluded that the dikes were emplaced after the layering was folded and that a weak, local deformation may have produced minor, open folds in the veins. Since the Mt. Maggiore gabbroic dikes show no folding or boudinage related to deformation in the peridotite, they could be interpreted as postdating the peridotite high-temperature deformation.

If the dikes were emplaced during the plastic deformation, there should exist some relationship between the orientation of the dikes and the strain ellipse deduced from the peridotite fabric. In addition, the undeformed nature of the dikes would have to be accounted for. In the Lanzo, Italy, tectonized lherzolite massif, Boudier and Nicolas (1972, 1977) describe feldspar lenses corresponding to about 5% incipient melting in the peridotite, and gabbroic dikes whose emplacement apparently accompanied the high-temperature plastic flow. The feldspathic lenses there form veinlets oriented about 15° oblique to the foliation and parallel to the flow, or shear plane deduced from petrofabric analysis. These veinlets grade into gabbroic dikes, some with dunitic walls, which commonly lie perpendicular to the peridotite lineation. The differences in orientation between the feldspathic veinlets that puddled at olivine grain boundaries and the more intrusive gabbroic dikes could be due to varying fluid pressures in the dike magma (Jackson 1979).

Quantitative evaluation of the principal strains undergone by the Mt. Maggiore peridotite is impossible due the absence of measurable strain markers, including boudinaged dikes, in these rocks. However, it is

possible to estimate the general shape of the strain ellipse, if we assume coaxial strain (strain axes parallel to stress axes) and infinitesimal strain (< 10% shortening). These are reasonable assumptions for the weakly deformed Mt. Maggiore rocks, where only one pulse of dike injection can be determined from cross-cutting dike relations. The stereogram in figure 13 shows the orientations of the gabbroic dikes with regard to the principal axes, X, Y, Z, of the infinitesimal strain ellipsoid. The XZ strain plane is assumed to be parallel to the peridotite foliation plane containing the stretching lineation.

Although poles to the dikes tend to group around a weak point maxima and partial girdle about 45° oblique to the XZ strain plane, their orientations vary considerably throughout the massif. Even so, individual dikes are often perpendicular to each other locally and might have formed as fillings of conjugate fractures, more or less symmetric about the peridotite foliation plane. The dikes sketched in figure 14 lie about 60° apart and, within measurable error, intersect at the Y axis of the strain ellipse. A continuous band of clinopyroxene around the dikes' intersection indicates that they crystallized simultaneously. In another example, left lateral shear operative on WNW trending dike planes (fig. 15) and the foliation in the dikes (figs. 8, 15) suggest that intrusion of the dike magma may have followed shear fracture planes oblique with respect to the foliation. These data suggest a geometric relation between gabbroic dike orientations and the peridotite foliation plane. If the dike injection occurred as the solid state flow of the cooling peridotite began to cease, then the only such deformation recorded by the dikes would be the sheared and broken clinopyroxene crystals forming an internal foliation within the gabbro.

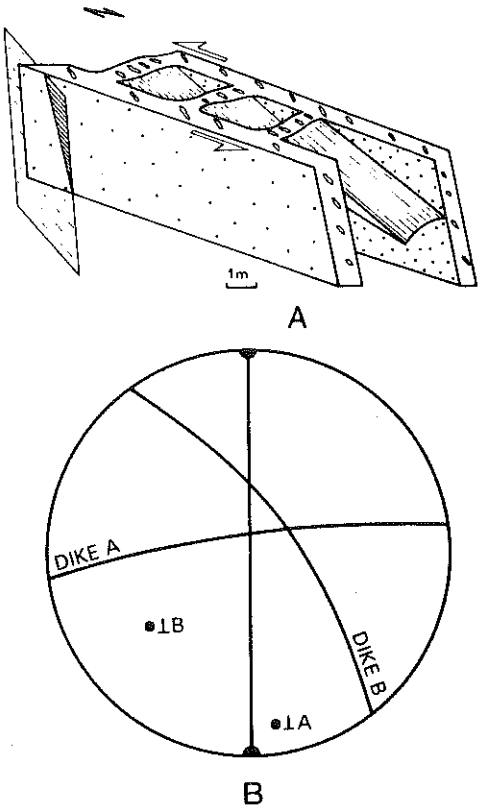

FIG. 15.—Gabbroic shear dikes with associated magma filled tension fractures. *A:* Block diagrams of field orientations. *B:* Stereogram with structural reference system of the folia on vertical, NS and the lineation horizontal, NS.

CONCLUSIONS

The Mt. Maggiore peridotite forms the tectonized base of a reconstructed ophiolitic section outcropping in eastern Alpine Corsica (Ohnenstetter 1975). Within the lherzolite, high-temperature peridotite flow structures indicate deformation under rotational stress conditions with a sinistral sense of shear. However, the lherzolite exhibits coarse-grained protogranular textures (Mercier and Nicolas 1975) and is only slightly deformed.

Within the lherzolite, plagioclase frequently penetrates along grain boundaries and at olivine triple points. Diffuse, discontinuous, 1 cm-thick plagioclase veinlets are common but show no preferred orientation, probably reflecting the relatively undeformed nature of the lherzolite, which failed to align the plagioclase. Lace-like networks of plagioclase also surround dunite pods and in-

vade the neighboring lherzolite. The presence of "in situ" plagioclase blebs indicates requilibration of the massif peridotite near the spinel-plagioclase lherzolite phase boundary that occurs at a pressure of about 9 kb, as well as associated partial melting of the lherzolite as the peridotite crossed the lherzolite dry solidus at 1200°C. Similar plagioclase-bearing peridotites have been described, for example, at Ronda, Spain (Dickey 1970), Lanzo, Italy (Boudier and Nicolas 1972, 1976), Troodos, Cyprus (Menzies and Allen 1974), the Josephine peridotite, Oregon (Dick 1977), Red Mountain, New Zealand (Sinton 1977), and Canyon Mountain, Oregon (Himmelberg and Loney 1980). They have been interpreted to be the result of trapped mafic melt reacting with the residual peridotite at pressures less than 9 kb (Menzies and Allen 1974; Sinton 1977).

Within the Mt. Maggiore massif, peridotite and gabbroic dike compositions compare closely with samples from the deep oceanic crust (Sinton 1978). Olivine mineral analyses range from Fo_{90-91} in the peridotites to Fo_{85} in the olivine troctolite intrusive dikes. Interstitial plagioclase contained in dunite is An_{88}, becoming more sodic (An_{54}) in the gabbroic dike sample. Spinel compositions range from chrome-rich in the lherzolites and pyroxenite banding to Al, Mg-rich in the depleted dunites.

Intrusive gabbroic dikes, ranging in mineralogy from troctolites to Fe and Ti gabbros, cross-cut the plagioclase segregations. The dikes are often perpendicular to each other locally and might have formed as fillings of conjugate fractures, more or less symmetric about the peridotite foliation plane.

These peridotite and gabbroic structures and textures suggest that the emplacement of the dikes was contemporaneous with the high-temperature deformation of the peridotite, perhaps as it flowed away from a spreading oceanic ridge (Nicolas *et al.* 1980). The differentiation pattern in the gabbroic dike compositions is similar to those of gabbroic cumulate rocks located within the overlying Corsican ophiolitic section (Ohnenstetter et al. 1975) and suggests that the Mt. Maggiore gabbro dikes may have fed these magma chambers.

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MAGMATIC AND TECTONIC EVENTS IN AN OCEANIC DIAPIR : EVIDENCE FROM THE MONTE MAGGIORE MASSIF, CORSICA.

Maryse OHNENSTETTER (first draft)

ABSTRACT

The lherzolitic massif from Monte Maggiore, Northern Corsica, France (3 km long, 1.5 km wide) which belongs to an ophiolitic complex rests tectonically on a pile of thrust slabs. Structural, petrological and mineralogical studies were undertaken to determine the chronology of coupled magmatic and tectonic events during diapiric uprise of oceanic mantle.

The oldest rocks belong to well layered and folded Cpx-rich peridotites equilibrated in an Al-rich pyroxene facies (layering I). The fold axis plunges towards the South. These peridotites have suffered limited to extensive partial fusion in the Spinel facies giving a residue of clinopyroxene-poor lherzolite. This partial fusion was probably initiated by pervasive injections of some alkali and incompatible-element rich liquids migrating along shear zones and extensive area related to a thermal channel. During progressive uprise of the mantle, the lineation rotated from an inclined to a sub-horizontal position and the deviatoric stress increased. A mappable layered plagioclase-rich peridotite, occurring towards the centre of a lens of depleted peridotites (0.8-0.5 km), indicates fractionation of olivine-spinel-orthopyroxene-clinopyroxene and plagioclase during magma ascent depending upon the pressure : orthopyroxene disappears as a liquidus phase in the least evolved rocks as pressure decreases. The extrated peridotites (layering II) have higher content of incompatible and fusible elements thant the residual peridotites of the layering I. A process of dissolution-recrystallization occurred during partial fusion and high-pressure fractionation.

The circulation of alkali and incompatible-element rich liquids persisted after partial fusion and during high and low-pressure fractionation. It leads to local enrichment of Cr, Ti, Na, K and P in the residual and fractionated peridotites. Shear zones in peridotite and some dunitic dykes are related to the crystallization from an evolved magma rich in alkali and incompatible elements. This magma reacted with the hot enclosing peridotites.

As pressure decreased, the temperature of the enclosing peridotite decreased. Dykes with straight boundaries are filled by plastically deformed gabbro. In the dyke, the rotation of the clinopyroxene lineation recorded the displacement of peridotite blocks during magma injection.

I - INTRODUCTION

Petrological and structural features related to oceanic mantle formed at an accreting plate boundary may occur in ophiolites. In the Western part of the Alpine Mediterranean area, ophiolites show close petrological affinities with oceanic crust whatever kind of discrimination is chosen. The structural and petrographical features related to diapiric intrusion have been analysed in the Monte Maggiore massif. This massif escaped important late deformation during secondary emplacement on a continental margin.

THAYER (1972) divided Alpine-type peridotites into two groups based on the proportion of clinopyroxene : the lherzolite-type and the harzburgite-type. Both groups exist in the Mediterranean area with the lherzolitic type in the western part (NICOLAS et JACKSON, 1972). Some lherzolitic massifs are associated with continental crust, whereas others may belong to an ophiolitic association (VUAGNAT, 1968) like harzburgite in the eastern Mediterranean ophiolites (ROCCI et al., 1975). Ophiolitic peridotites are generally equilibrated only in the spinel (Sp) and/or plagioclase (Pl) facies. They have generally been deformed both during their primary emplacement in an accretion zone and during their secondary emplacement onto a continental margin (NICOLAS et LE PICHON, 1980) or during strike-slip movements (SALEEBY, 1977). Mylonitic and porphyroclastic textures often obscured diapiric textures (protogranular to poeciloblastic textures) which persist in some massifs like Vourinos (ROSS et al., 1980). Due to the presence of lavas and peridotites in the ophiolitic association, partial fusion and/or some fractionation may occur in the diapir during flowage of the peridotite when P and T decrease. Both processes may lead to coarse-grained assemblages which are difficult to identify. George (1977) has reviewed the kind of flow which may exist in peridotites. Besides syntectonic recrystallization and plastic flow, he mentioned mechanisms of dissolution and recrystallization when solid and liquid are present (DEN TEX, 1969) as may be expected during the diapiric uprise of mantle.

The purpose of this paper is to indicate the chronology of the main petrological events which occured during the deformation of the Monte Maggiore massif. The chronology is based mainly on cross-cutting relationships. Evident features are dykes with more or less straight boundaries. Other elements may be

obtained from successive parageneses, from mineral facies, and from the proportion, size and crystallization order of the essential phases. Maps and structural features allow one to show areas where partial fusion and fractionation processes occurred. Major element variation of bulk-rock analyses and mineralogical evolution will help in the recognition of residual mantle from fractionated and contaminated peridotites during the upward migration of liquids.

II - REGIONAL SETTING

It is now widely accepted that ophiolitic fragments from Corsica, the Alps and the Apennine once formed part of an oceanic crust closely resembling present mid-oceanic ridge products (OHNENSTETTER et OHNENSTETTER, 1975, 1976; BECCALUVA et al. 1977). These ophiolites form remnants of the Liguro-Piedmont ocean (DAL PIAZ, 1974) which closed during Jurassic times. Zircon or whole-rock dates as well as sediment ages indicate that accretion might have occured during 40 Ma (from 180 to 140 Ma) (OHNENSTETTER et al. 1981). Consequently the petrographical and structural features of ophiolites, especially those in the Monte Maggiore massif, may be related to the begining of an oceanization process which may differ slightly from steady-state mid-oceanic ridge evolution (ROOTS et al. 1979). Early volcanism in anomalous ridge segments also exhibits a transitional character (VENTURELLI et al. 1981) later followed by normal-type MORB. In the Liguro-Piedmont ocean, ridge volcanism is interrupted by transform-fault activity. The related ophiolitic series indicates that magmatism existed during the tectonic and stratigraphic processes (OHNENSTETTER M., 1979). Low-pressure fractionation is controlled by olivine (Ol), plagioclase (Pl) and clinopyroxene (Cpx) giving troctolite and associated dunite and anorthosite, olivine-gabbro, gabbro, ferrogabbro and some leucoferrodiorite and albitite (OHNENSTETTER et al. 1975). The peridotites, the deepest part of the ophiolitic section are mainly of lherzolitic character. They may have arisen in a more or less evolved ridge related to the age of the basin (from continental to oceanic crust) and/or through transform-fault tectonics. Thrusting linked to the closure of the oceanic basin occurred at least from the Middle Cretaceous (CARON, 1977; COHEN et al. 1981). During the upper Cretaceous, possible intraoceanic subduction ended with the blockage of a continental block at depth which initiated the obduction of the ophiolitic remnants on the western shield (MATTAUER and PROUST, 1976). Successive tectonics and metamorphism led to

complex imbricate tectonic slices of ophiolite and continental blocks with their respective sedimentary covers. In the cumulates, three successive metamorphic stages were observed; a greenschist to amphibolite facies one linked to an oceanic event, a high-pressure phase of upper Cretaceous age and a terminal greenschist facies developed during Eocene to Oligocene times (OHNENSTETTER et al. 1976). Important tectonic movements after the main metamorphic phases were responsible for the stacking of various structural and metamorphic units (AMAUDRIC DU CHAFFAUT et SALIOT, 1979).

III - THE MONTE MAGGIORE AREA

In the Northern Cap Corse, the Monte Maggiore peridotitic massif structurally occupies a high position, resting tectonically on strongly deformed and metamorphosed ophiolitic gabbros, lavas and sediments. Peridotites and the ophiolitic rock constitute two different slices thrust on to a large continental block made up of isoclinally folded gneiss and micashists (LAPORTE, 1954) partly enclosed in the deformed ophiolitic nappes. According to GUILLOU (1962) and GUILLOU et al. (1962), the tectonic pile may belong to the normal flank of a large fold. The relative freshness and absence of important Alpine tectonic may also be explained by a Ligurian origin and thrusting towards the East (GUILLOU, 1962; GUILLOU et al. 1962; DURAND DELGA, 1978) or South-East (LAPORTE, 1955). The peridotite body may also have sliced towards West (MATTAUER et PROUST, 1975) in agreement with the Austro-Alpine character of the shield (CABY, 1981). A more recent interpretation relates the thrusting to a southerly movement during the Upper Cretaceous. It occurred during strike-slip motion along the North Pyrenean fault blocked by a rigid continental edge which modify the general westwards overthrusting (MALAVIELLE, 1982). Microstructures, elongation lineations and the quartz petrofabrics in a quartzite band, 20 m below the southern part of the massif indicate a westerly movement of the ophiolite and its sedimentary cover (JACKSON et OHNENSTETTER, 1981). The lherzolitic peridotite massif is one of the freshest massifs in Corsica. Near the cool tectonic contact, serpentinites (antigorite and chrysotile) are developed as well as along the main faults (see map; fig 1). Gabbroic dykes generally metamorphosed in the greenschist facies are locally cut by veins of serpentine, zeolite and prehnite (GUILLOU 1962; AZAIS et al. 1968). The presence of pure jadeite may testify to a limited high-pressure phase of Alpine metamorphism (JACKSON et OHNENSTETTER, 1981). Ni mineralization varies with the degree of serpenti-

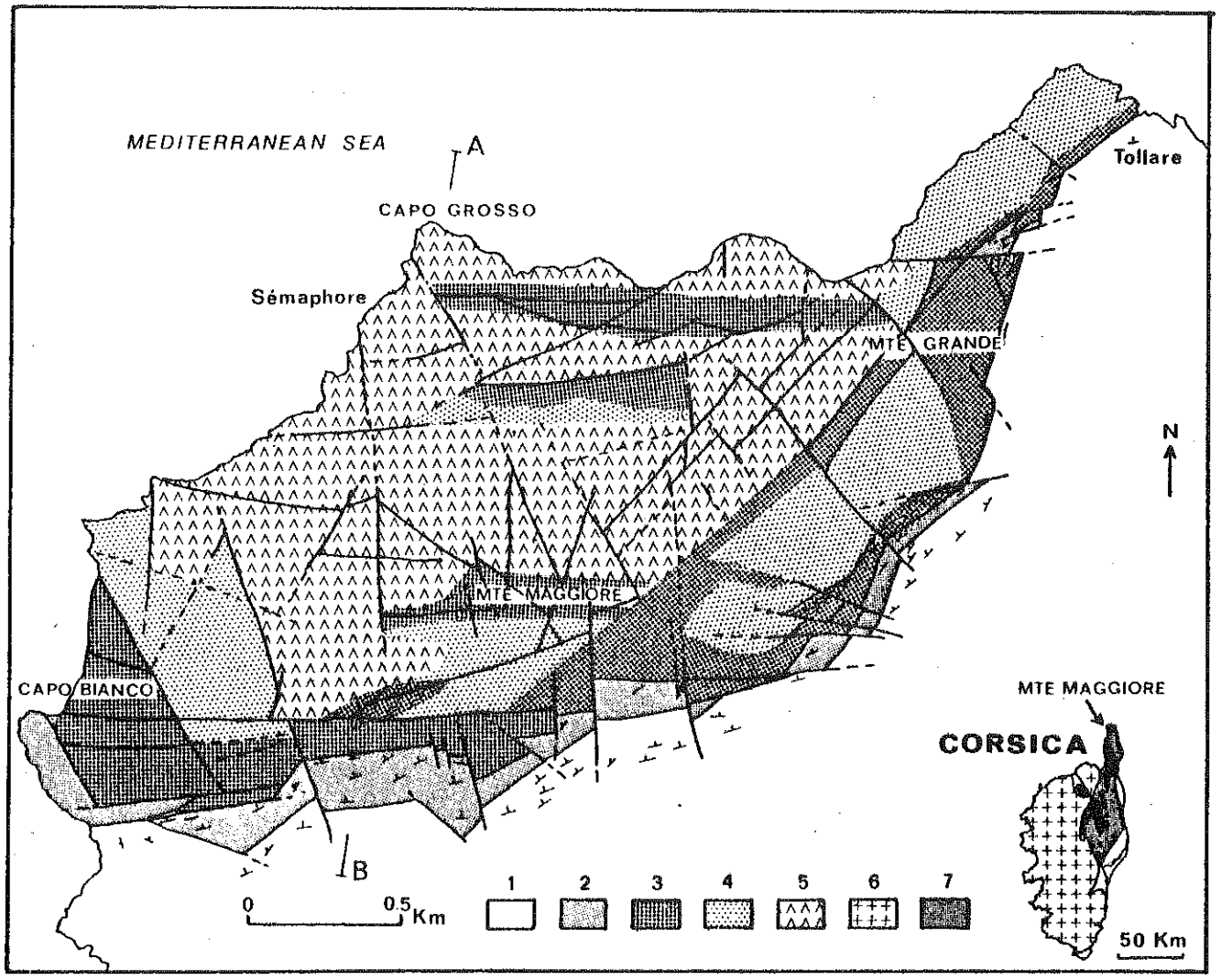


Fig. 1

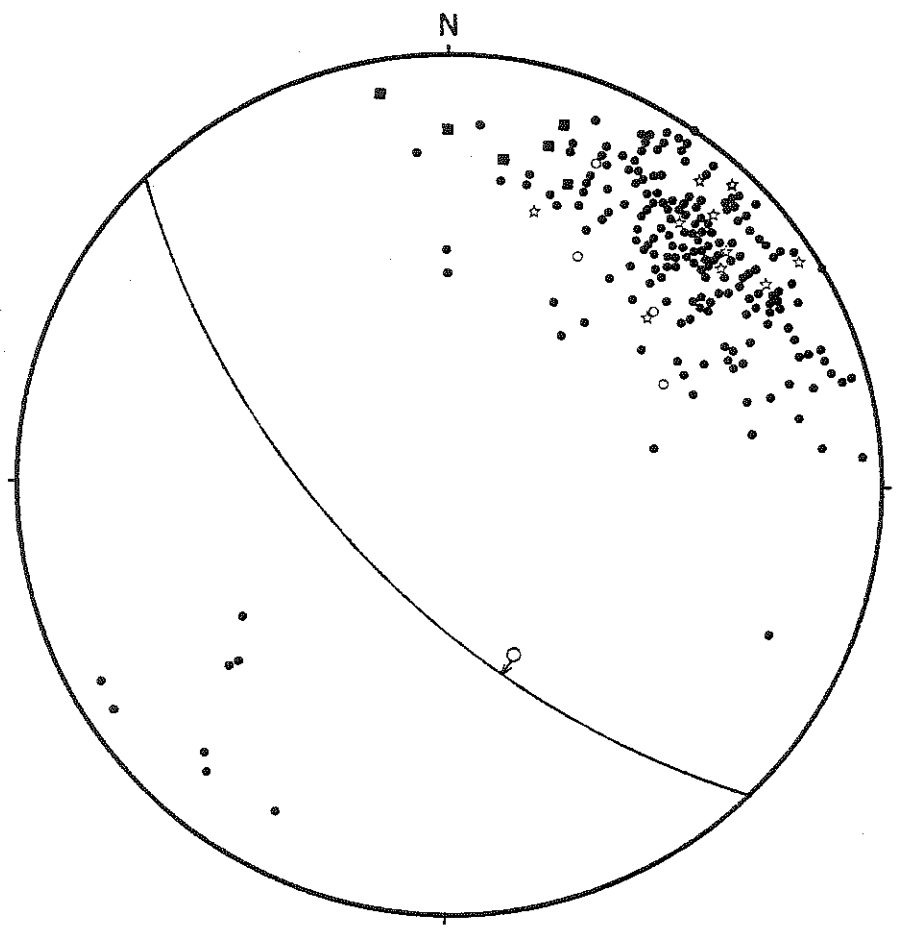
Fig. 1 The Monte Maggiore peridotite complex.
1; shield gneiss of Centuri, 2; metamorphosed ophiolitic units (gabbros, lavas and metasediments) , 3; massives or sheared serpentinites, 4; highly serpentinized peridotites, 5 peridotites slightly serpentinized. --- presumed faults, ——— faults.

nization of the massif (AZAIS et al. 1968). The relative abundance of Cpx in the peridotite has always been noted. The layering of magmatic origin is made up of wehrlite, lherzolite, some harzburgite, diallagite and websterite (GUILLOU, 1962). Dunites are later than the layered peridotites as well as the gabbroic dykes (GUILLOU, 1962; OHNENSTETTER et al., 1975). The deformation of these gabbros is thought to have occurred during magmatic injection or slightly after (OHNENSTETTER et al., 1975). In the peridotites, aggregates of spinel testify to some deformation before serpentinization (AZAIS et al., 1968; OHNENSTETTER et al., 1975). Because of the low deformation, a cumulate terminology has been applied. Detailed study

revealed the importance of poikilitic minerals, mainly Ol, Pl and some Opx and Cpx during the crystallization of peridotites (OHNENSTETTER et al.,1975).In the whole massif the texture is protogranular (OHNENSTETTER et al., 1975; JACKSON et OHNENSTETTER, 1981). A structural study showed that the peridotites were penetratively deformed under rotational stress conditions with a sinistral sense of shear (JACKSON, 1979; JACKSON et OHNENSTETTER, 1981); the petrofabrics of olivine and enstatite indicate high-temperature (>1200º) intracrystalline slip following the (010) [100] system (JACKSON, 1979). The peridotitic flow was contemporaneous with the injection of dykes occurring in conjugate shear fractures as well as in a direction perpendicular to σ_3 due to lowering of the effective stresses (JACKSON, 1979; JACKSON et OHNENSTETTER, 1981; NICOLAS et JACKSON, 1982). Some structural similarities, especially the relationships between the deformation of the peridotites and the dyke attitude, exist between the Lanzo massif and the Monte Maggiore massif (JACKSON et OHNENSTETTER, 1981). Feldspar-rich veinlets later cut by dykes may be considered as magma formed during partial fusion at the spinel-plagioclase phase change around 9 kb (JACKSON, 1979; JACKSON et OHNENSTETTER, 1981).

IV - RELATIONSHIPS BETWEEN THE STRUCTURAL AND LITHOLOGICAL PLANES

Peridotites are generally of uniform composition and setting. The Monte Maggiore peridotites have five phases. Besides Sp and Pl which have a pressure dependence, the proportions of the three other phases are generally related to partial-fusion or to fractionation processes. As peridotite is a metamorphic rock, minerals may be deformed and aligned in the plane of foliation. Because of the slight deformation shown by the Monte Maggiore peridotite, aggregates and spindles of Sp have been used in the field to map the foliation plane and lineation. Other remarkable planes generally cut the layering and foliation planes. They are related to lithological planes (fig.3) marked by variable proportions of rocks or minerals : dunite, Pl-rich veins with more or less Ol-Opx-Cpx. The Pl-rich rocks were classified in three groups; the first one is formed by Pl patches (2 cm long-2 mm large) between the other minerals. An increase in the size of the Pl patches is also accompagnied by the presence of other phases. They form veinlets (second group) having diffuse boundaries and variable size : 5 cm to 7 m long, 2 cm to 50 cm large. The third group is made up of coarse-grained Pl-rich veins having well defined walls (0.4-3 m long - 2 cm large). The most evident feature in the



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Variations of lithologic planes

- pyroxenes
- plagioclases
- dunites, ± depleted harzburgites
- plane-axis

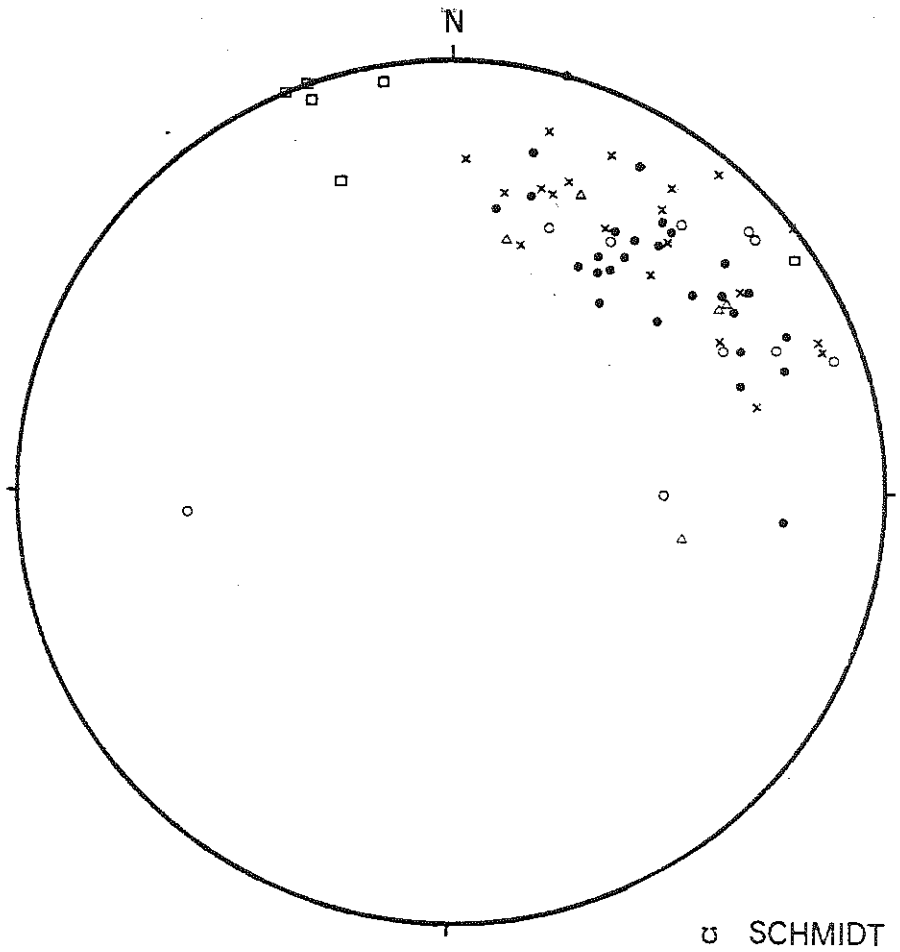

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Fig. 5 Dunitic pods

- pod border
- pod S_o
- pod axis
- S_1
- ± dunite dyke

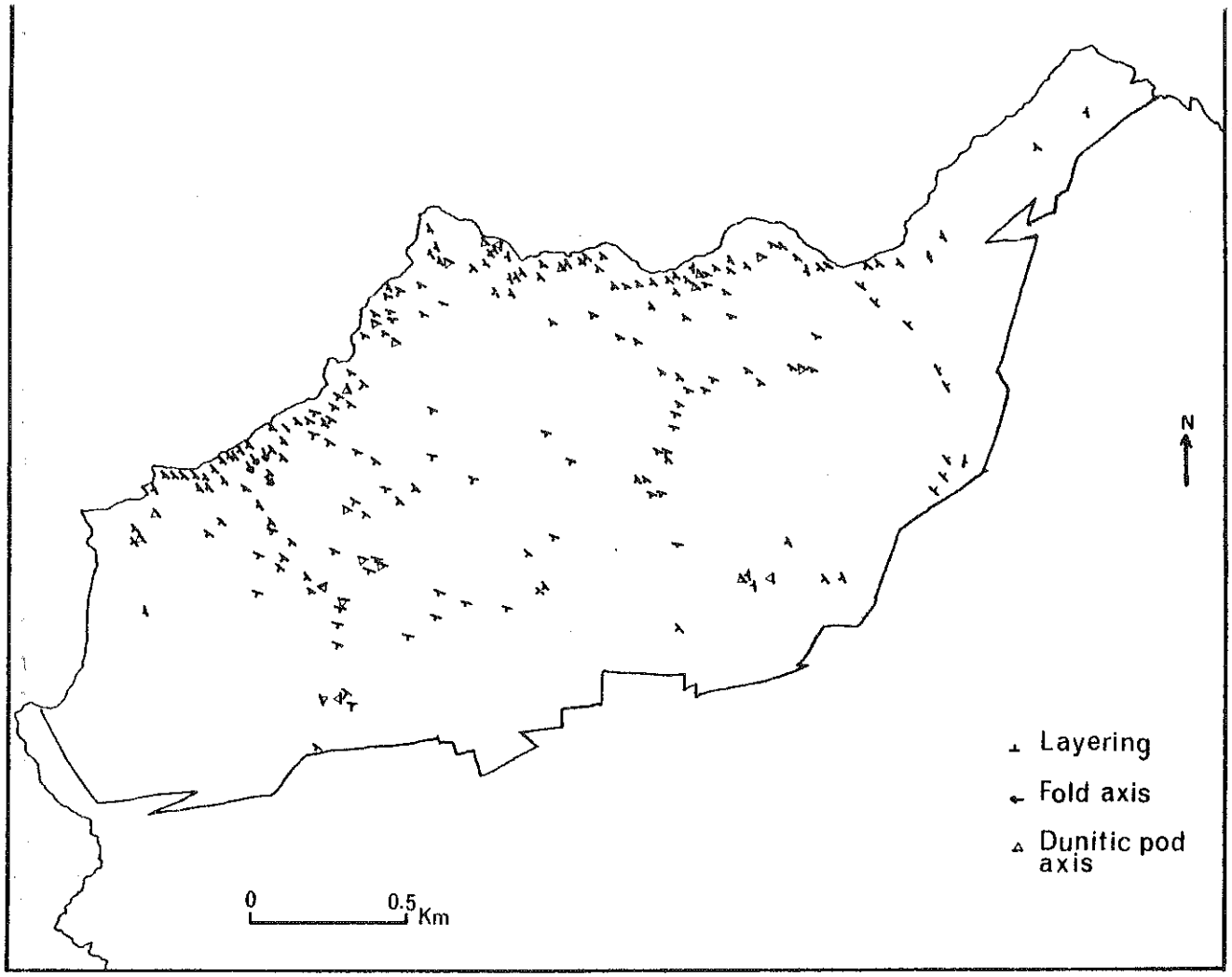

Fig. 2

Fig. 4 Petrographic facies.
1) layered peridotite (lherzolite, Ol-websterite, Harzburgite, dunite)
2) Cpx-poor peridotite
3) highly depleted peridotite
4) Opx enlargment in every kind of rock ± spinel

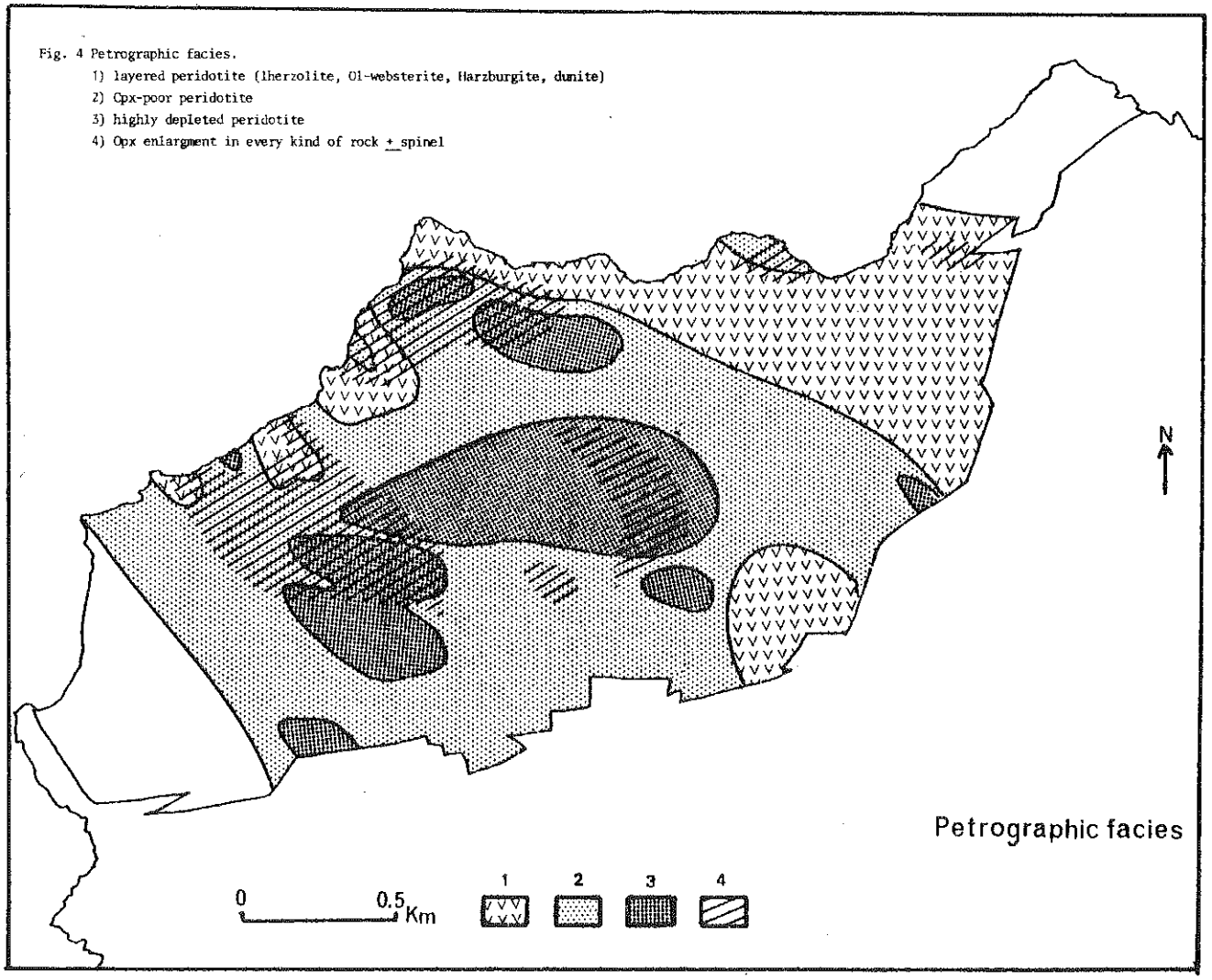

Fig. 4

massif is the presence of secant dykes of gabbroic composition (troctolite to ferrogabbro), sometimes bordered by dunite, having diffuse contacts with the enclosing peridotites.

The final purpose is to draw a map which relates the petrographical variation (fig.4) to the structural planes and to establish the chronology of the different magmatic and tectonic events. Special attention was paid to the position of the different minerals, especially Sp and Pl.

A - LAYERING

Layering marked by variable proportions of Olivine, orthopyroxene et clinopyroxene is more or less evident dependant upon the area considered. Along the Northern and Western coast, the massif is well layered (fig. 2). Pyroxenitic bands alternate with lherzolite, harzburgite and dunite showing coarse grained minerals. The layering is locally isoclinally folded. The fold axis is inclined towards South (N 140º-500º S). In the North Eastern area, some elliptical patches of clinopyroxenite may be the remnants of a hinge fold. Clinopyroxene-rich layers may be interrupted by elongated pods of spinel-rich dunite. Some coarse grained orthopyroxenite occurs in the banding. In the layers, clinopyroxene is of variable size and form. Orthopyroxene is large. It may also fill the interstices between Olivine and clinopyroxene giving to the rock a peculiar aspect. In some places along the coast and towards the centre of the massif, the layering diseappears and folds can no longer be observed. Some aggregates of two pyroxenes mark relic layering. These occur in a depleted unlayered clinopyroxene-poor lherzolite. In such rocks clinopyroxene decreases in size and proportion giving medium-grained or fine-grained peridotite. When it is medium-grained, clinopyroxene of variable size and form is scattered in an olivine-rich matrix. Sometimes, in this matrix, large Opx (0.5-1 cm) form patches. Peridotites are considered to be fine-grained when clinopyroxene is small (1-2 mm) and round. They are also devoid of pustulous Orthopyroxene (See plate 1). Clinopyroxene, sometimes associated to orthopyroxene, is generally enclosed in olivine of large size as revealed by microscopic observations (see plate1).The most frequent orientation of the pyroxenitic layerign (fig. 2) is about N 135º - 75º S. The layering is nearly parallel to the axial plane of folds. They occupy a more meridian position at Capo Grosso, Capo Bianco and towards the southern part of the massif (see map, fig. 2).

This type of layering constitutes the oldest recognizible features of the massif. As spinel, olivine and plagioclase locally interrupt the pyroxene-rich layers, it may be suggested that the lherzolite and pyroxenite layering may be older than the development of spinel. No garnet nor pseudomorph after it have been observed in these bands.For clarity, this layering will be refered to Layering I. In thin section, Opx + Sp $\pm$ Ol developed at the expense of Cpx (plate 1). Generally, Ol + Pl $\pm$ Sp $\pm$ Opx replace earlier Cpx. In some places, peridotites are destabilized. Minerals were cut by micro-veins (plate 3). In one case amphibole (Amph) + Pl + Opx + Ol is associated with apatite and Ti-rich Sp (contaminated peridotites) (plate 3).

On the maps (fig. 2) it may also be noticed that in some rocks the layering trends towards E.W. These rocks are genarally coarse-grained and composed of dunite, harzburgite, Cpx-poor lherzolite and some lherzolite. Pl and subhedral Sp are in variable proportion. The setting and the nature of these layered rocks indicate that they do not correspond to a more pronounced rotation of Layering I. This new layering is called Layering II. (Plate 2).

A schematic petrographical map has been drawned (fig. 4). It indicates the areas where peridotites are well layered (Layering I). Zones of high depletion are restricted to elliptical masses enclosed in slightly depleted peridotites, where patches of two pyroxenes may be recognized. Knots of large Opx may be superimposed in all the previous areas and in every kind of rocks. Layering II may be found everywhere but occurs mainly in the depleted to slightly depleted area. (fig. 3).

B - THE SETTING OF THE DUNITES

They may be found as thin bands of variable thickness 10 cm large and are associated with rocks with Layering I and II. Dunites also occur in spherical to elliptic pods (10 cm to 5 m for the greatest axis) with no sharp contact with the lherzolite. Remarkable planes are constitued by the borders, the axial plane of the pods, layering inside the pods marked by Sp and foliation (fig. 5). Borders deviate slightly from layering I as well as the axial plane (N 125º $\pm$50ºS). Their dip is also less inclined than that of Layering I. Layering inside the pods is sometimes sub-parallel to the borders. Generally it is at a small angle to the border ($<$15º) and trends E-W like layering II. The foliation inside, marked by aggregates of Sp rimmed by Pl, varies in trend from that of the foliation associated with Layering I towards a more E-W direction.

Dunites also occur in subvertical dykes trending E-W to N 70º. In one case a dyke is subparallel to Layering I. The contact between dunite and peridotite, evident in the field is not so sharp in thin section. It was not displaced by late tectonic movements. Towards the centre, discontinous zones are locally occupied by interstitial Sp, Pl and Cpx. In some places, coarse-grainde dunite and/or Pl-rich dunite separate the gabbroic dykes from the peridotite (dunitic walls). Contacts are gradational with the peridotites (2-3 cm) and sharp with the laminated troctolite inside. Dunites found in xenoliths inside the Ol-gabbro.

C - THE STRUCTURAL ATTITUDE OF PLANES MARKED BY SPINEL

Sp is an important accessory phase everywhere but its proportion varies from place to place (fig. 6). Nearly always present in the Northern and Western part of the massif, it becomes rare towards the central part (fig. 7). In this area, it

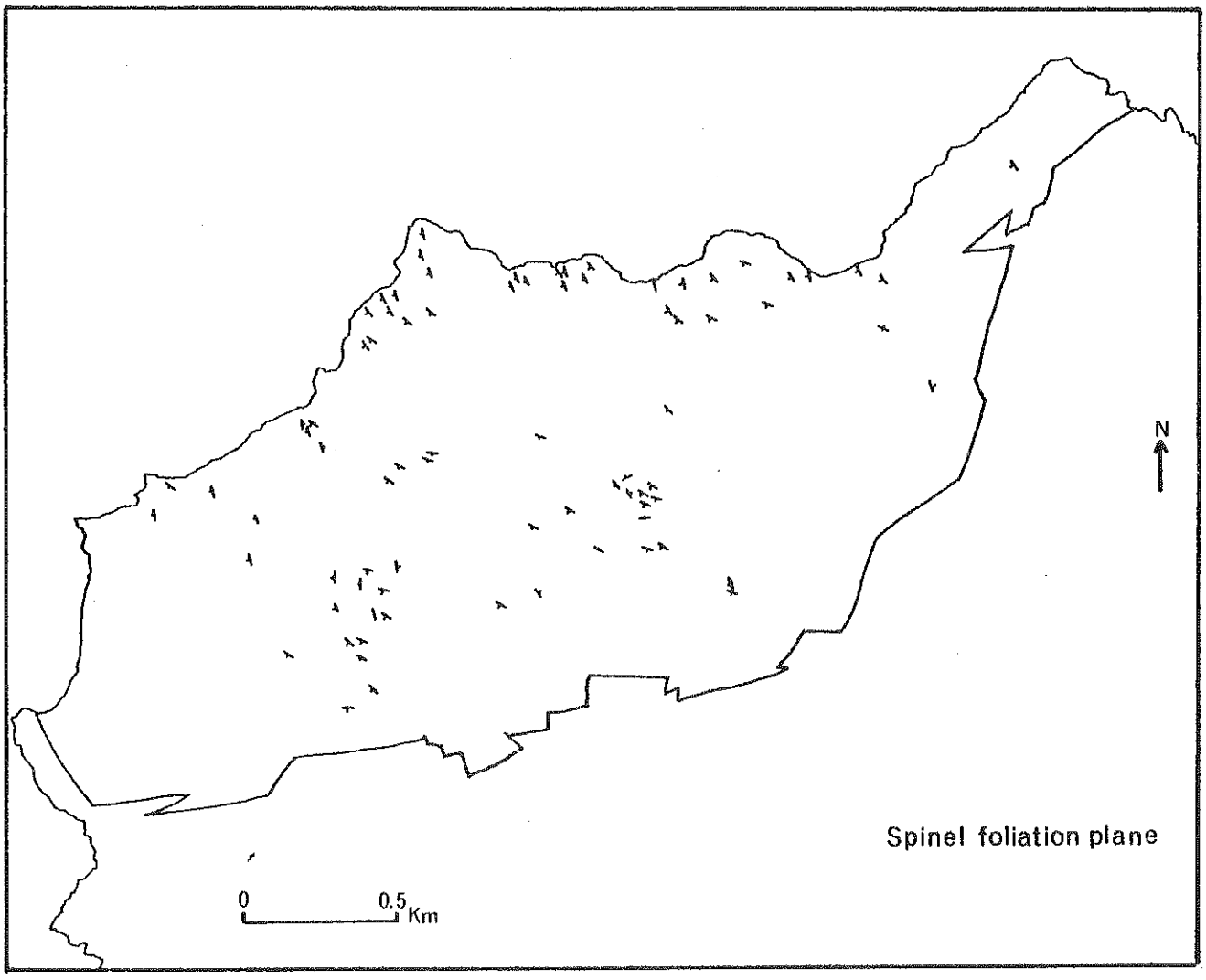

Fig. 7

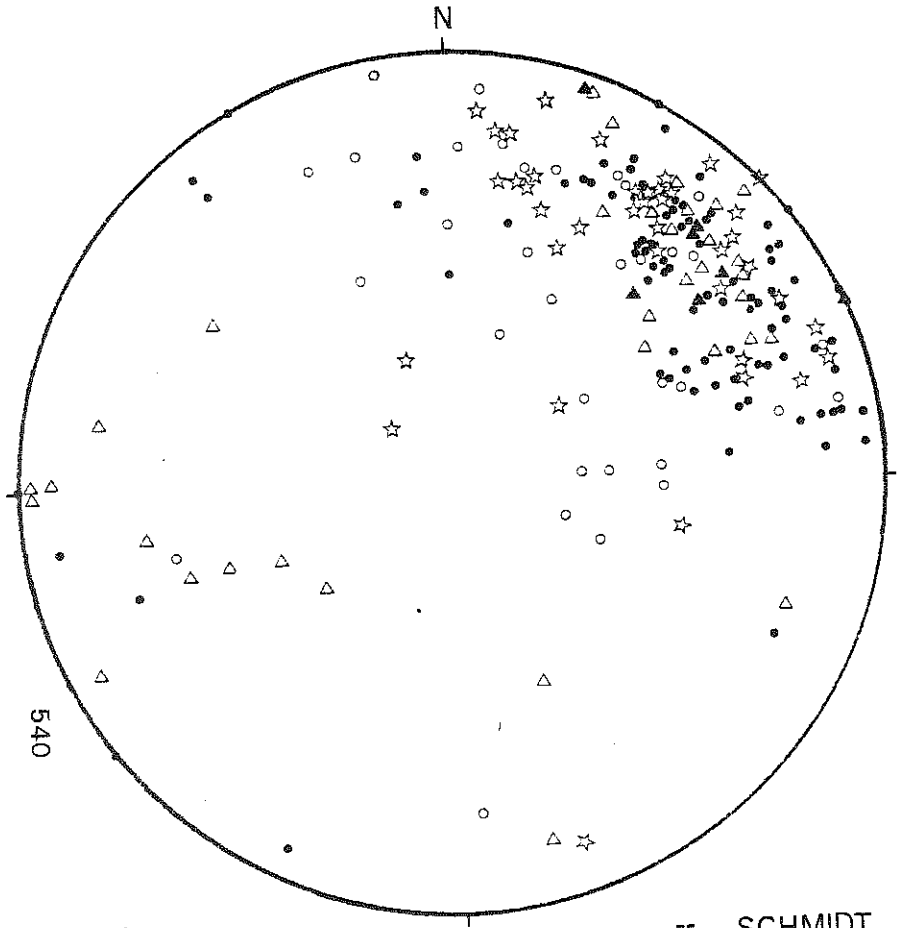

Fig. 6

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Planes marked with spinels

- • spinels
- △ spinels lodes
- ○ spinels with plagioclases
- ☆ spinels in or with olivines
- ▲ spinels with orthopyroxenes

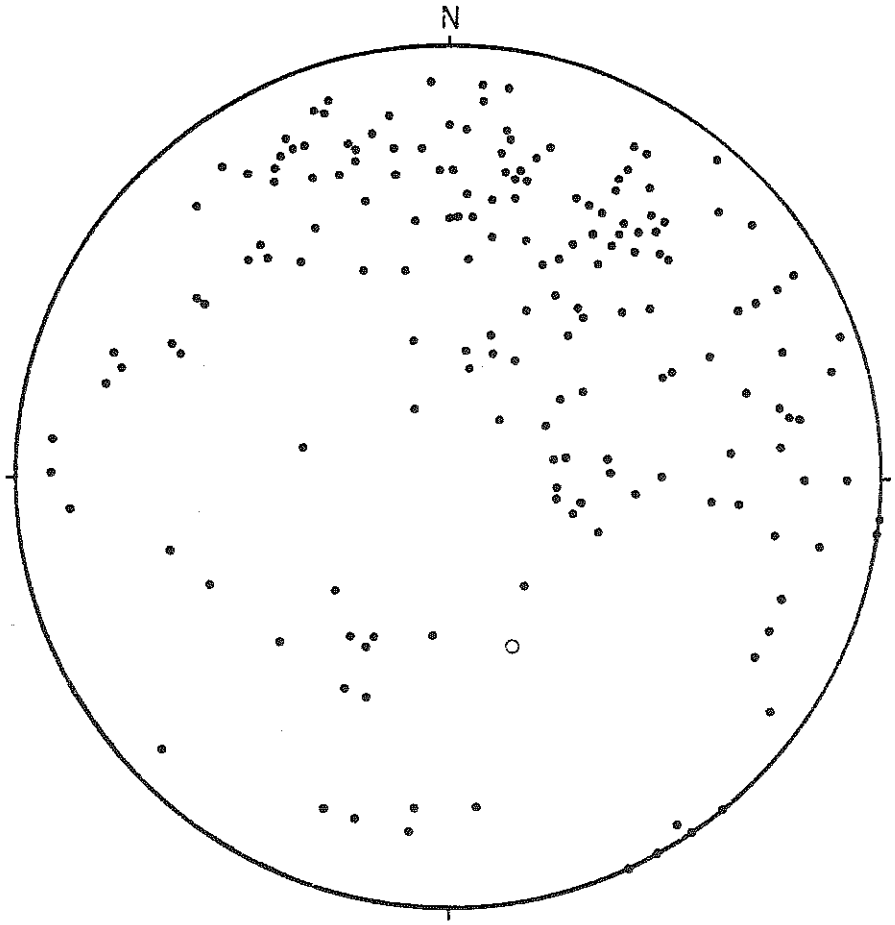

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Fig. 8 Collections of plagioclases

○ Fold axis

may reach a high proportion in dunites, some harzburgites and Pl-rich veins. Aggregates and spindles of Sp were used to find the foliation plane (fig. 6). The maxima of Sp aggregates occupy a more southerly position than the layering. They are of variable size : coarse-grained to the North and fine-grained towards the centre. They may contain only 2 or 3 minerals. Lineations are slightly inclined towards South. In one case, the aggregates spread out parallel to the foliation plane, whereas each mineral grain developed in a direction perpendicular to the extension.

Planes which contain Sp are of more variable attitude than the pyroxenitic planes. Planes with poikilitic Opx around Sp are mainly restricted to Layering I and sometimes to Layering II on the Monte Maggiore slopes. Planes related to Sp included in Ol or associated with Ol are more scattered in orientation. They belong to the foliation plane, to the layering plane I and II and also to the elliptical dunite pods. They are also parallel to the secant dunitic dykes. Sp with Ol and aggregates of Sp may define some slightly inclined plane like Pl-rich planes containing coarse Sp. In this case, the planes vary around an E-W position. It may nevetheless be noticed that the Sp and Pl association may be found in Layering I and II.

D - THE STRUCTURAL ATTITUDE OF PLANES MARKED BY PLAGIOCLASE

In the field, Pl lies along numerous diffuse planes, bordering all the other minerals or forming veins and dykes with more or less diffuse boundaries. The map (fig. 9A)shows where Pl-rich veins form better defined planes along the coast and on the slopes of the Monte Maggiore area. A first projection is related to tiny discontinuous Pl-rich zones (2 cm-2 mm). In the field, one or two directions may be measured at any one place, besides some Pl-rich planes sub-parallel to the layering. Veins vary around and E-W trending zone (N 130º - N 60º). Most of them may belong to two girdles whose axes are respectively related to the fold axis and the stretching lineation (fig. 8). Most of the planes dip towards the S, S-W, and W. Some planes of variable direction are slightly inclined to or nearly horizontal. Low dips have also been noticed for some layered rocks in the elliptical dunitic pods.

When Pl associated with pyroxene, Ol and some Sp belong to longer veins, they occupy more restricted planes cutting the previous layering I. They fill distansion zones compatible with a major compressive axis near a N 110º direction.

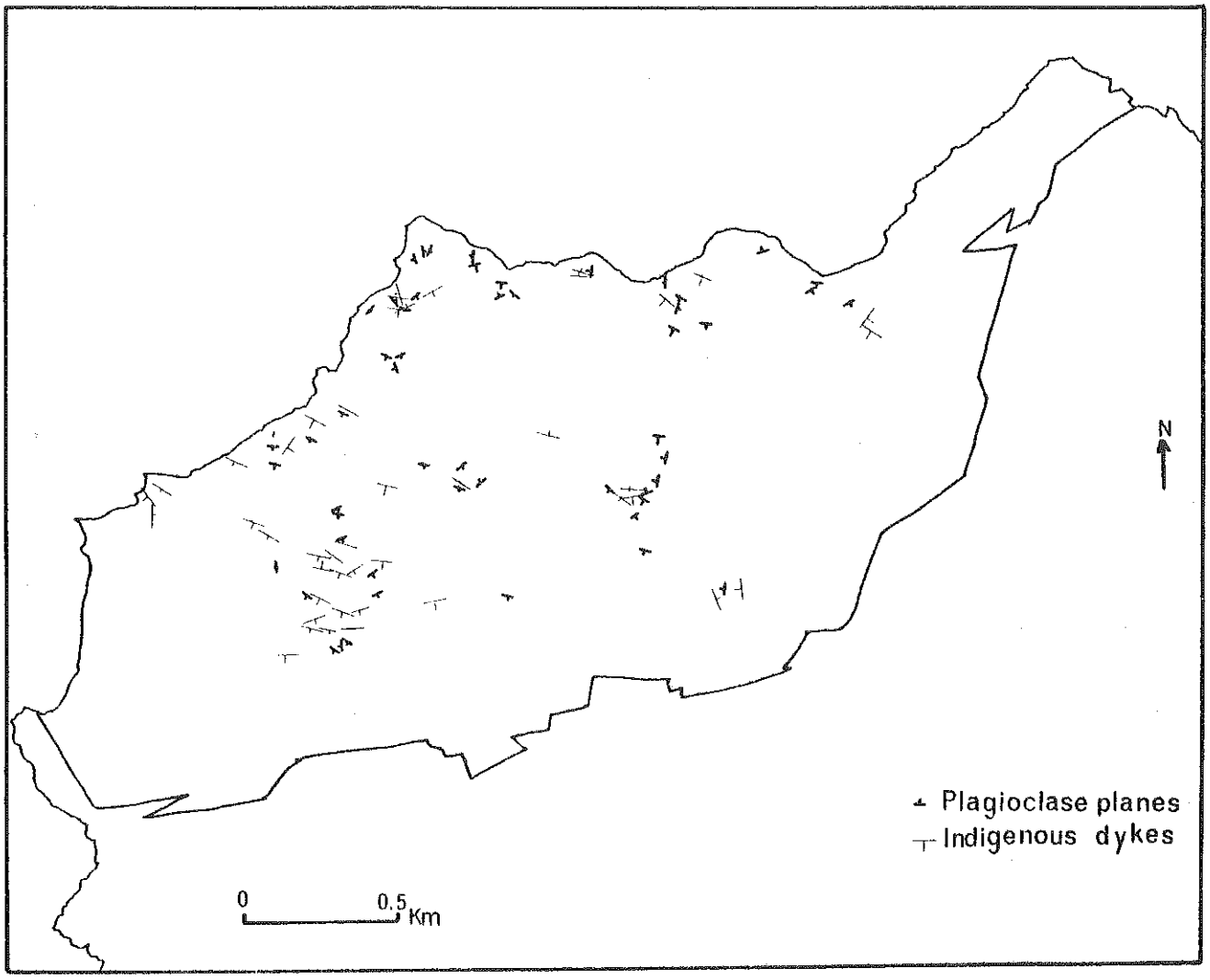

Fig. 9 A

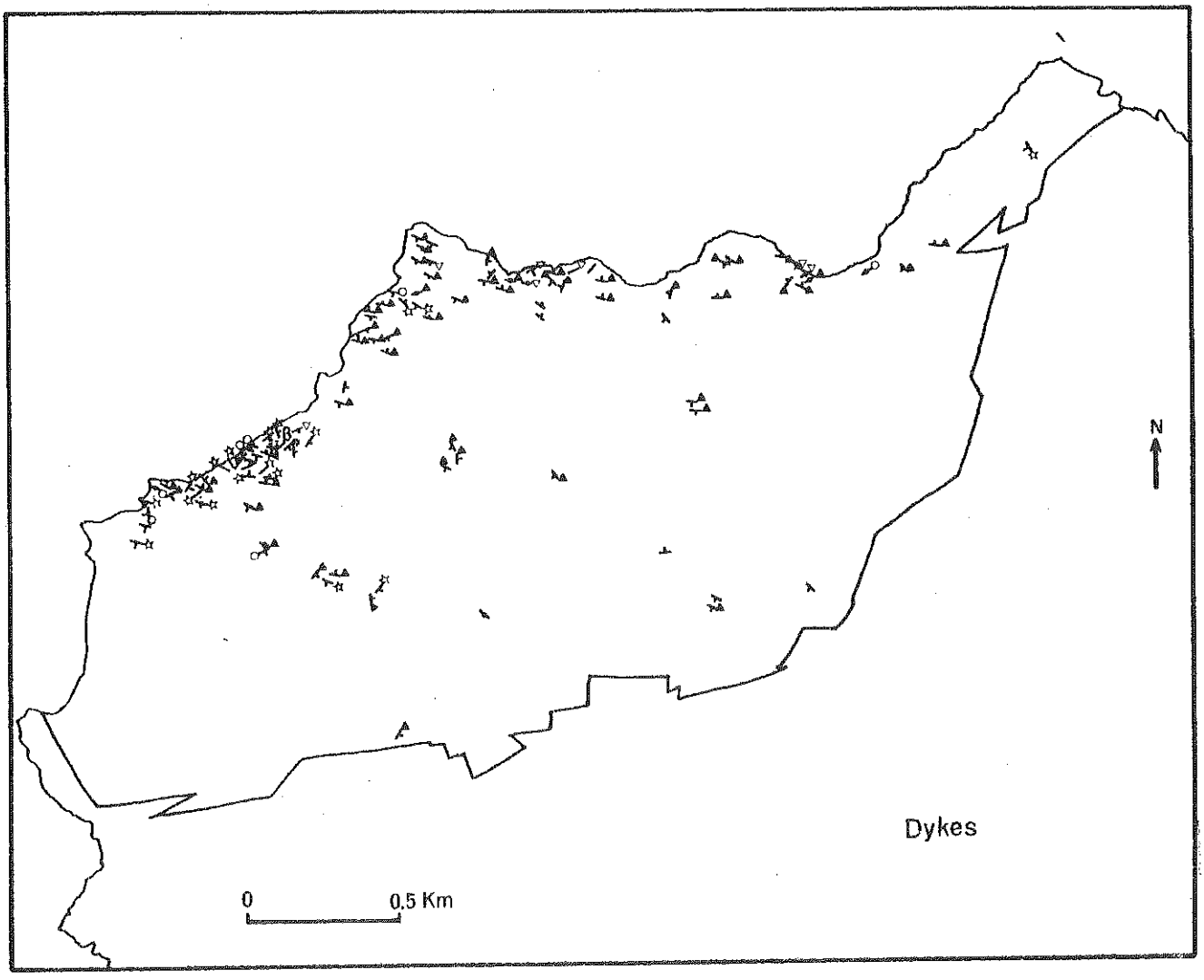

Fig. 9 B

The gabbroic dykes cut straightly the peridotite (fig. 9B). Generally the borders (except the contact between dunite and peridotite) have moved after magmatic injection during the serpentinization and metamorphism of the gabbros. Generally, all petrographical types occupy all dyke directions (fig. 10). Ferrogabbro occurs mainly in subhorizontal planes along the western coast. The maxima are about N 110º as for the plagiferous veins but they are steeply inclined. Gabbroic dykes occur in similar fracture zones as the veins and also the tiny Pl-collections. Their opening may thus be related to reactivation of old fractures. During emplacement of the veins, no important displacement of the wall occurred During dyke injection, important movements occurred (OHNENSTETTER and GAGNY, 1982, in prep.). They are recorded by the rotation of Cpx which nucleated and grew subperpendicularly to the wall. A structural analysis in each case shows that the growing Cpx defines a lineation with a direction between the stretching lineation and the normal to the wall. Therefore the opening of extension zones may be slightly oblique to the maximum stres direction, a fact already noticed in metamorphic terrains (FYFE et al. 1978). During dyke injection, Cpx rotates because of some horizontal and vertical movements of the wall, leading to plastic deformation of the gabbroic dyke. They get tabular, granular and porphyroclastic textures which are not seen in the enclosing peridotites. Along N 110º vertical dykes, generally the northern block of peridotite is displaced horizontally to the west. This sinistral shearing is compatible with that noted in the peridotite (JACKSON et OHNENSTETTER, 1981). When displacement along horizontal minus stress is no longer possible, boudins of gabbros are vertically displaced. Along sub-horizontal planes, the upper block moves towards North. Some platy faults are relatively early : they cut dunitic pods and have not been displaced during subsecant serpentinization. Tectonic activity during magma emplacement is indicated by the rotation of Cpx, the high temperature of deformation, the assymmetric features related to secant shear planes. These shear zones control the size, morphology and distribution of the magmatic phases. Deformation increases when the dykes were injected in the vicinity of the flow plane which affected the peridotite (N 120º - N 130º). These dykes may have been injected later than underformed or slightly deformed dykes trending N 80º.

The main chronological points may be resumed from cross-cutting relationships. The oldest pyroxene-rich assemblage belonging to layering I is folded. The fold axis is more inclined than the later stretching lineation indicated by the Sp or Ol fabric. A secant layering II occupies more E-W trending planes which correspond to extension zones. Layering and foliation planes and extension zones

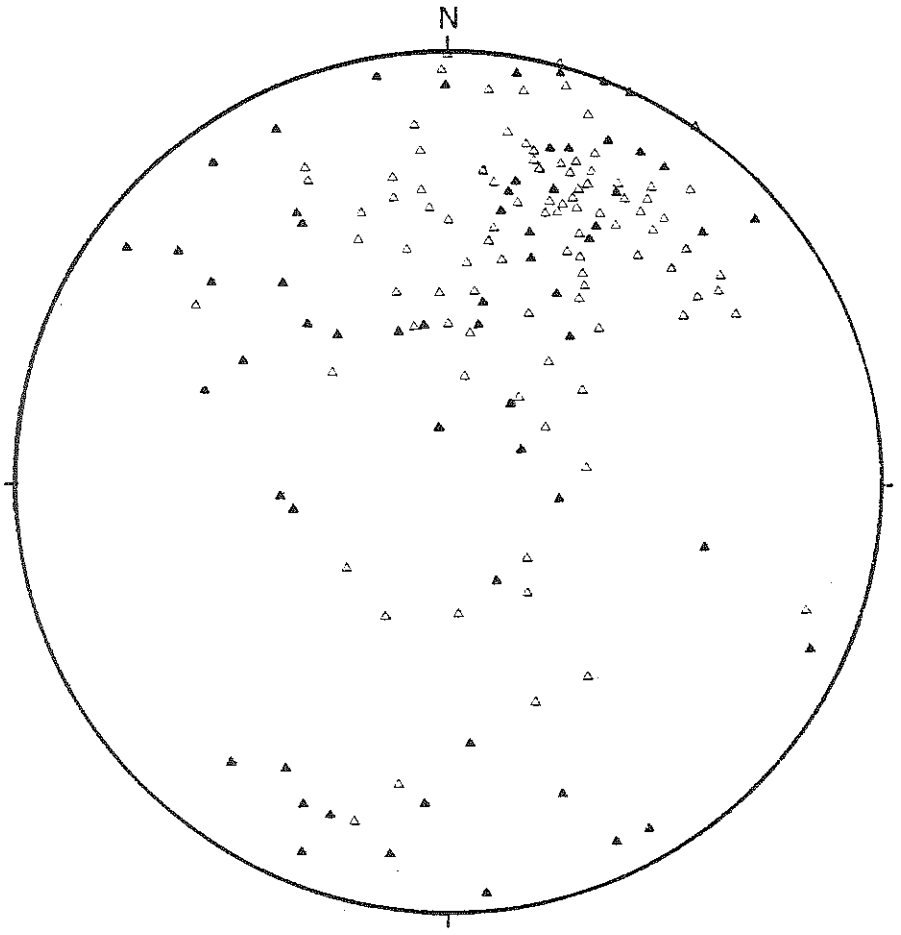

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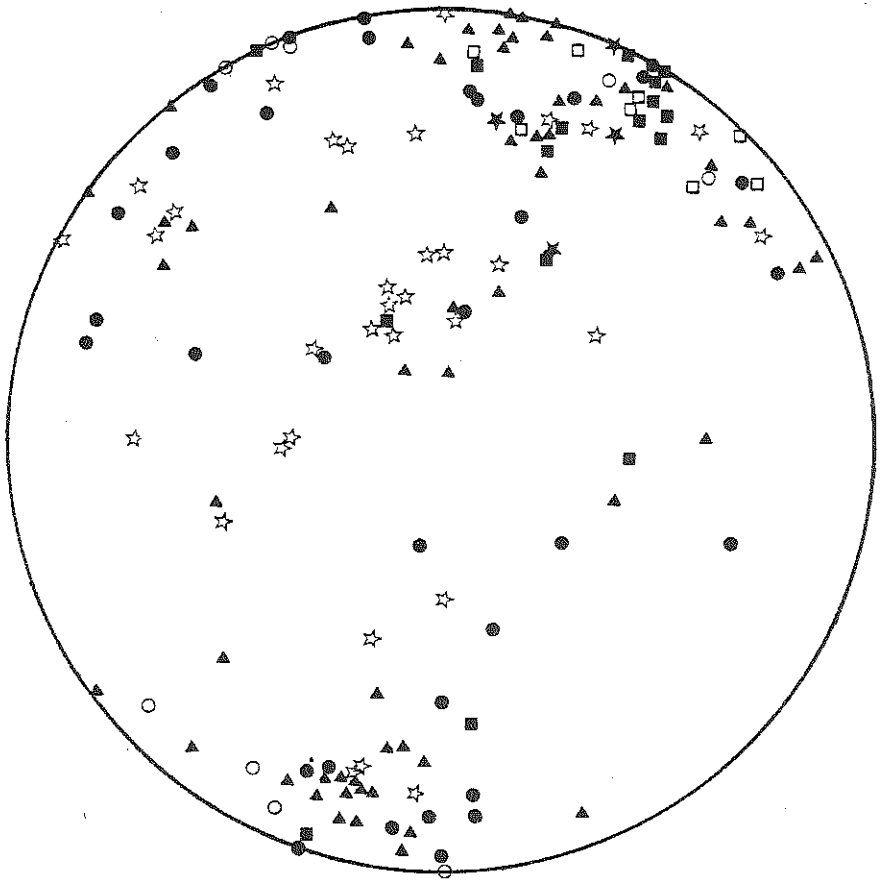
Fig. 10

△ discontinuous veining

▲ dykelets



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Dykes

○ dunite

■ troctolite

□ olivine – gabbro

●

☆ ferrogabbro ; leucoferrogabbro

▲ gabbro ± olivine ± orthopyroxene

★ norite

are also occupied by plagioclase concentrations and Pl-rich veins with more or less diffuse boundaries. Dykes of dunitic and gabbroic composition cut all the previous rocks. They were deformed during magma injection. (OHNENSTETTER et GAGNY, in prep.).

IV - MAJOR ELEMENT VARIATION

A - VARIATION OF FeO and MgO

Recalculated analyses (without water) of the Monte Maggiore peridotites (fig.II) are in an FeO-MgO diagram in two ways. The first one is based on petro-

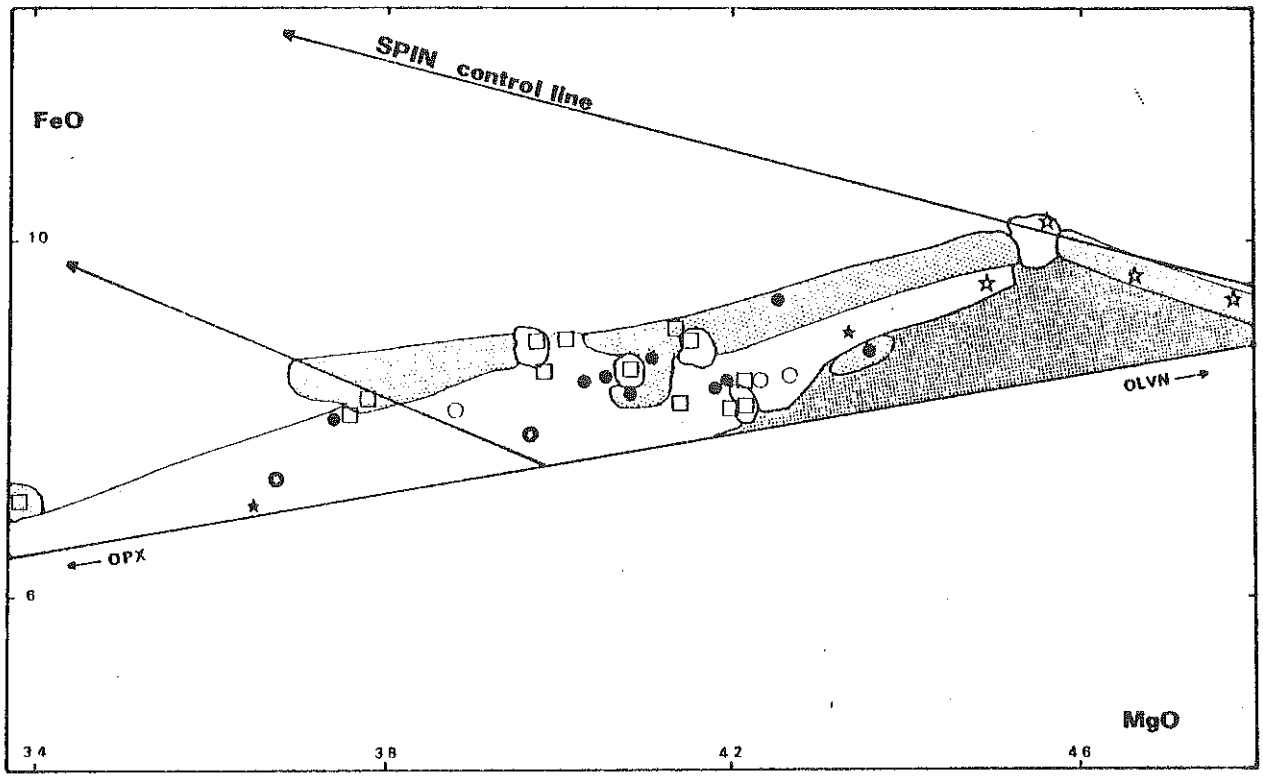


Fig. 11 FeO-MgO diagram of mantle rocks and minerals. triangle: dunite, crosses: harzburgites, circle: lherzolite, square: depleted lherzolites, star: orthopyroxenite.

graphical variation -pyroxenite, orthopyroxenite, lherzolite, Cpx-poor lherzolite, harzburgite and dunite which reflect different clinopyroxene/orthopyroxene ratios and two-pyroxene/olivine ratios. A special subscript indicates Pl-rich types, for example- Pl-dunite, Pl-harzburgite or Pl-lherzolite. The second plot refers to layering which for clarity are named the old young and intermediate one. They correspond to different layered rocks defined from petrographical and structural relationship in the fields. Most of the samples from the old layering come from the Northern and Western coast. They consist of pyroxenite, more or less spinel-rich orthopyroxenite and variously depleted lherzolites. Samples from younger layering correspond to some Pl-rich lherzolite, harzburgite and dunite.

In the FeO-MgO diagram, samples are spread out along a line which is also the Pl control line. The mean and possible range of composition of the main phases (Sp-Ol-Opx-Cpx) are also plotted. Means of silicate compositions are well aligned but below the rock trend except for some samples. This is in agreement with the fact that most of the sample bear some accessory spinel (with or without Pl) which increases the Fe/Mg ratio of the bulk-rock. Addition of both Sp and Pl to a sample moves it to the left, towards lower MgO content for slight FeO variation. It may be noted that some pyroxenites, orthopyroxenites and depleted lherzolites are close to the Ol-Opx-Cpx line. Sp-rich samples belonging to the old layering are strongly displaced towards the Sp field. Theoretically, in an FeO-MgO diagram, if the bulk-rock composition is linked to a partial fusion event, we should observe an FeO decrease and a corresponding MgO increase when Cpx and accessory aluminous phases disappear in samples. At first, this is not the case. Moreover, the destabilization of Cpx at the expense either of Opx or Ol moves the bulk composition to the right. In the present case, if most dunites and pyroxenites plot normally at higher and lower MgO values, the positions of the plots for lherzolite to harzburgite do not depend strongly on the proportion of Cpx. For example, most harzburgites are at lower MgO values than some lherzolites and Cpx-poor lherzolites. Samples from the new layering plot generally at high MgO values and far from the Ol-Opx-Cpx control line. The higher FeO-rich rocks are generally Pl and or Sp-rich. Similarly, if we always consider the partial fusion hypothesis, we must observe a decrease of fusible and incompatible element with MgO increase. Let us now consider the fields of equal CaO, Al_2O_3 in the FeO-MgO diagram.

B - VARIATION OF CaO

The concentration of CaO is linked to the proportion of Cpx (degree of partial fusion and to the proportion of accessory Pl). Broadly speaking from low to high MgO, a CaO decrease(fig. 12) is obtained but the line of equal CaO is not perpendicular to the Cpx or Pl control line except for higher FeO content at a given MgO value. We may consider what petrographical type covers the area between the equal-CaO lines : the higher content (3-4 % CaO) field consists of pyroxenites, a lherzolite which is compatible with the starting material, Sp-rich harzburgites locally found in the old-layering and some Pl-rich rocks (dunite, Cpx-poor lherzolite and pyroxenite) belonging to the younger layering. The 2-3 % CaO field is mainly of harzburgite, depleted lherzolite and minor lherzolite, all of them plagioclase-bearing or not. They belong generally to the

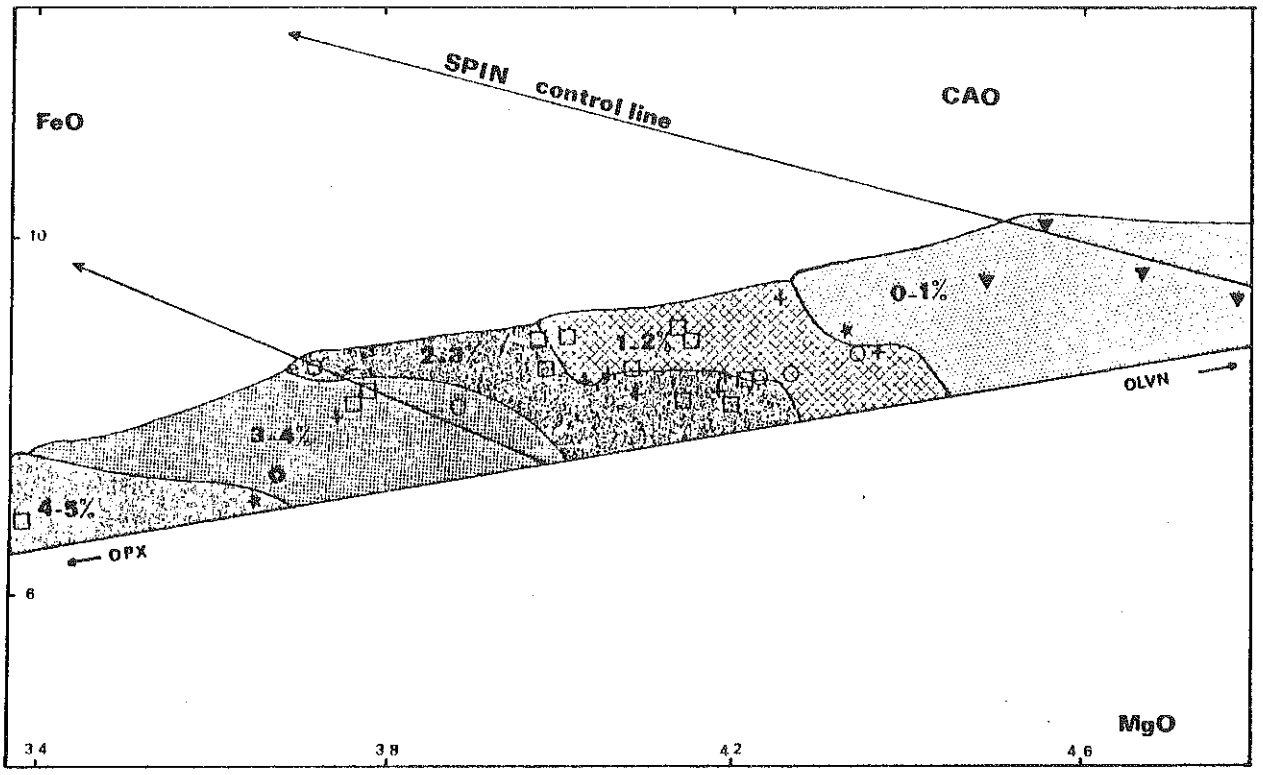


Fig. 12 Equal concentration of CaO in the FeO-MgO diagram. Cpx-poor rocks are generally Ca rich.

older layering except for the plagioclase-rich samples. The 1-2 % CaO field possesses mainly plagioclase-poor samples-olivine-pyroxenite, lherzolite, depleted lherzolite and harzburgite. They may be found in both types of layering. The lower CaO content occurs in dunite from the older layering.

C - VARIATION OF Al_2O_3 and TiO_2

Similar fields may be distinguished (fig. 13). The equal-Al_2O_3 lines are better defined and are slighly more inclined than for the CaO lines. The petrographic types are nearly similar to those of the CaO diagram. The inflexion curve for the CaO line compared to the Al_2O_3 line result from lower Ca/Al ratios in spinel-rich samples. TiO_2 is an incompatible element migrating into the liquid during partial fusion. Therefore, TiO_2 (fig. 14) should decrease from lherzolite to harzburgite. The isocontent lines reveal that TiO_2 is abundant in the starting material (pyroxenite and some lherzolite), in some spinel-rich orthopyroxenite and harzburgite equilibrated at high pressures. The lower concentrations are found in some olivine-pyroxenite, lherzolite, Cpx-poor lherzolite, dunite and some Pl-rich peridotites belonging to the younger layering. Intermediate values are found mainly in harzburgite. The main differences with the CaO and Al_2O_3 isocontent lines are in the respective positions of the spinel and plagioclase-rich samples having very low Ti-content at relatively low FeO and high CaO and Al_2O_3.

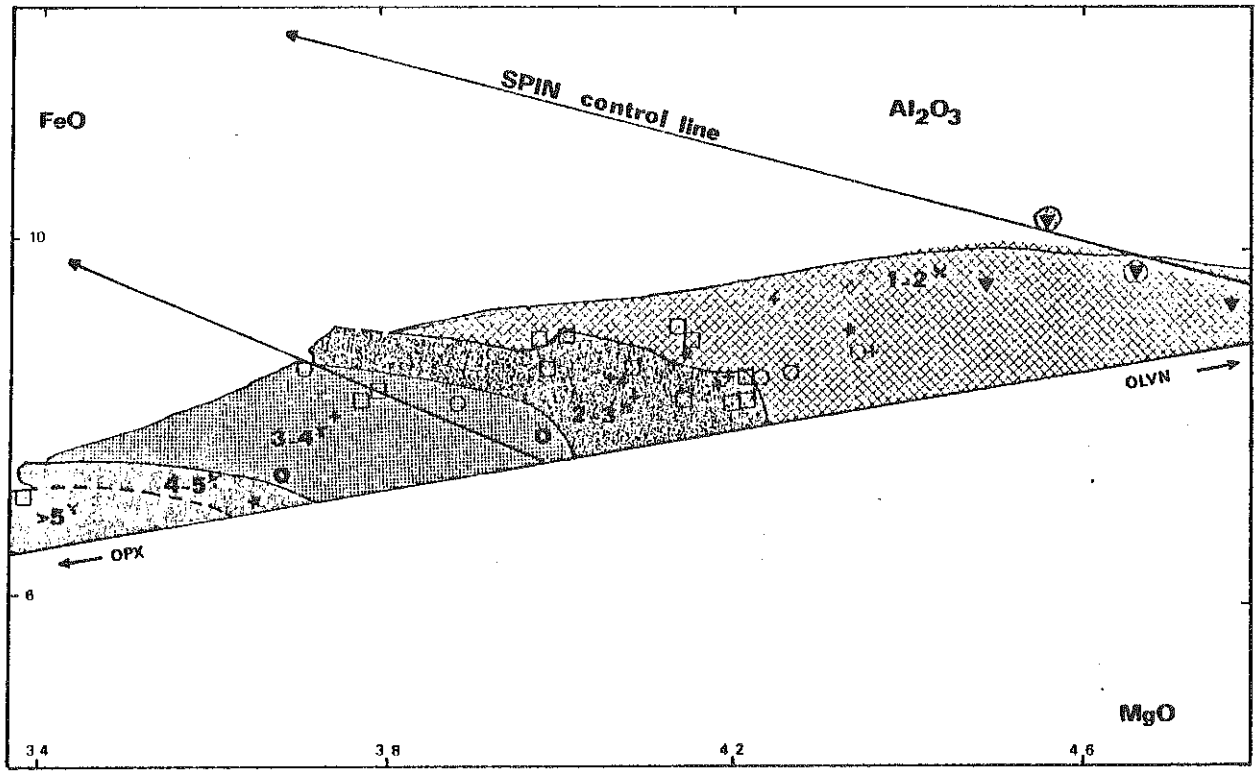


Fig. 13 Equal concentration of Al_2O_3 in the FeO-MgO diagram.

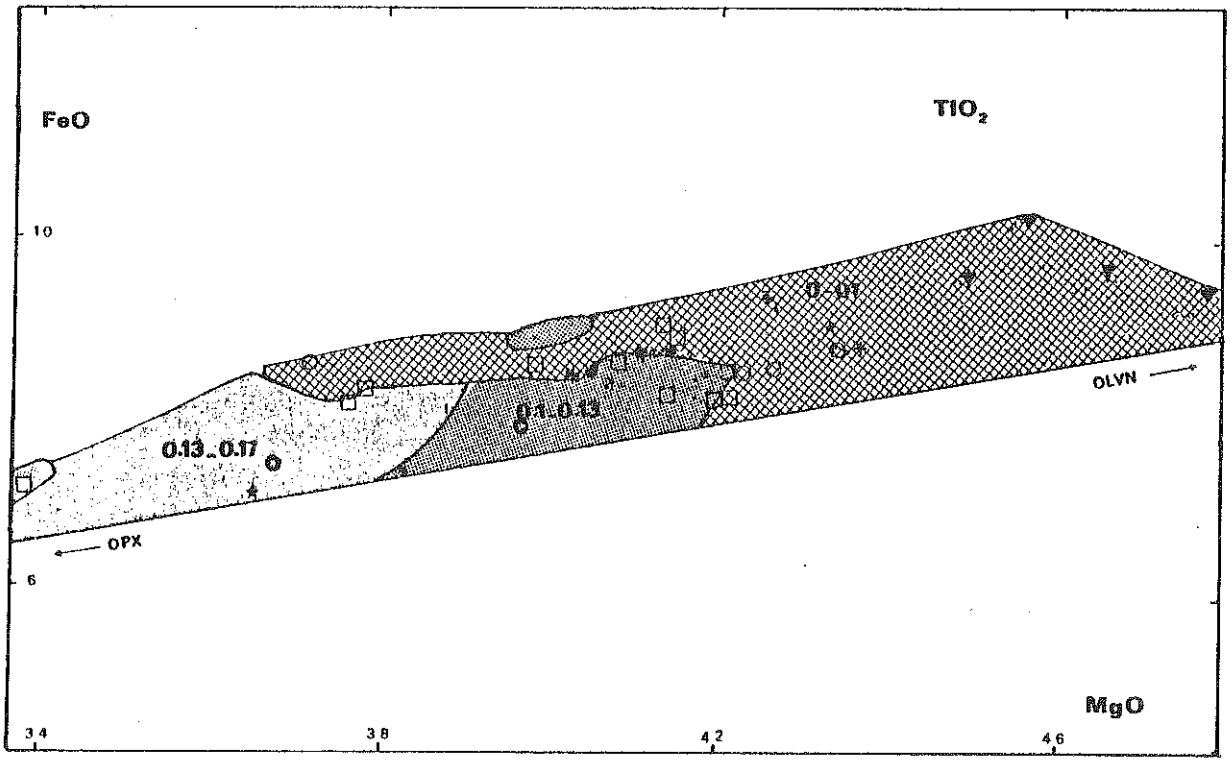


Fig. 14 Equal concentration of TiO_2 in the FeO-MgO diagram.

A comparison of FeO-MgO diagrams showing the distribution of the more fusible and incompatible elements leads to some conclusions. Firstly, MgO increase does not correspond necessarily with the progressive disappearance of Cpx. When it occurs, Cpx-poor rocks have nearly the lowest content of fusible elements, a fact compatible with a partial fusion hypothesis. Secondly, samples are displaced towards Sp $\pm$ Pl enrichment lines indicating that subsolidus reaction is not only responsible for the abundance of aluminous phases. The crystallization of the accessory Al-rich phases occurred rather during the percolation of an Al-rich magma. The migrating liquid may be also responsible for the crystallization of Opx-rich rocks which gives some harzburgite and Cpx-poor lherzolite. Increase of Opx in the samples explains the displacement towards the left, even if no Sp or Pl accompany this phase. Besides partial fusion events, several stages of crystallization may have occurred at different pressures. In layering I, the crystallization of Sp-harzburgite and some orthopyroxenite may have occurred in the Sp-lherzolite facies. In the Seiland facies (O'HARA, 1969), extraction of Opx, Sp and Pl may occur. The fractionated products occupy generally E-W trending planes and some slighly inclined planes. In the Pl-facies, Sp is less abundant compared to Pl. These fractionation processes occurred before injection of some dunitic and gabbroic dykes. The relative importance of partial fusion and fractionation may also be seen in the CaO-Al_2O_3 diagram. Most of the old layering fall close to the Cpx control line which may be related to the extent of partial fusion. In the absence of an aluminous phase, the initial composition of the Cpx should be Al-rich ($\simeq$ 1[illegible] %). Younger layering is at higher MgO values or is deviated towards the feldspar or spinel line. This indicates the refractory character of some cumulates and the relative enrichment in Sp, Pl or both.

V - MINERALOGICAL VARIATION

Mineralogical variation is less developed in alpine-type peridotites than in stratiform complexes. Nevertheless, a mineralogical study has been done to characterize the different rock types belonging to layering I, layering II, contaminated paridotites, dunitic pods and wall, gabbroic dykes.

A - OLIVINE

Olivine ranges from Fo_{92} to Fo_{90} in all the peridotites and in the Pl-rich veins and some early dunitic dykes. (fig. 15). These compositions are si-

milar to those of oceanic peridotites found, for example, at DSDP site 334 (HODGES and PAPIKE, 1976) and Leg 45, site 395 (SINTON, 1978). They are also similar to those of the Lanzo massif, of the Piedmont Alps (BOUDIER, 1976) and of the Erro-Tobbio area, Western Liguria, (ERNST and PICCARDO, 1979). In dunitic walls, Ol composition varies rapidly from the border towards the centre of the dykes (Fo_{80} to Fo_{70}). It is much more Fe-rich than in the enclosing peridotites, dunitic pods or Pl-rich veins (fig. 16). When they are cut by troc-

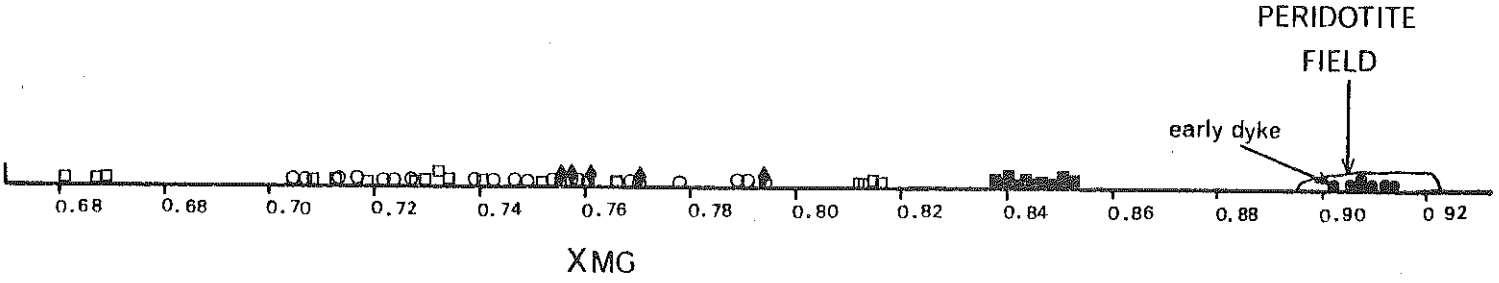


Fig. 15 Olivine composition of gabbroic dykes and peridotites.
Low Mg/Mg+Fe of gabbroic dykes and indigeneous dykes except for a sample in a conjugate plane related to the crystallization of layering II. Dunitic walls often crystallized from an alkali-richer magma (existence of mica and ilmenite).
Symbols are: + harzburgite; ○ dunite from dunitic walls, ▽ dunites ▼ plagiferous peridotite, △ contaminated peridotite, ◆ shear zone, ◊ spinel peridotite, ■ troctolite , □ ol-gabbro, ☆ Fe-gabbro, ★ norite, ▲ gabbro ± Ol ± Opx, ● early dykes, ✪ orthopyroxenite

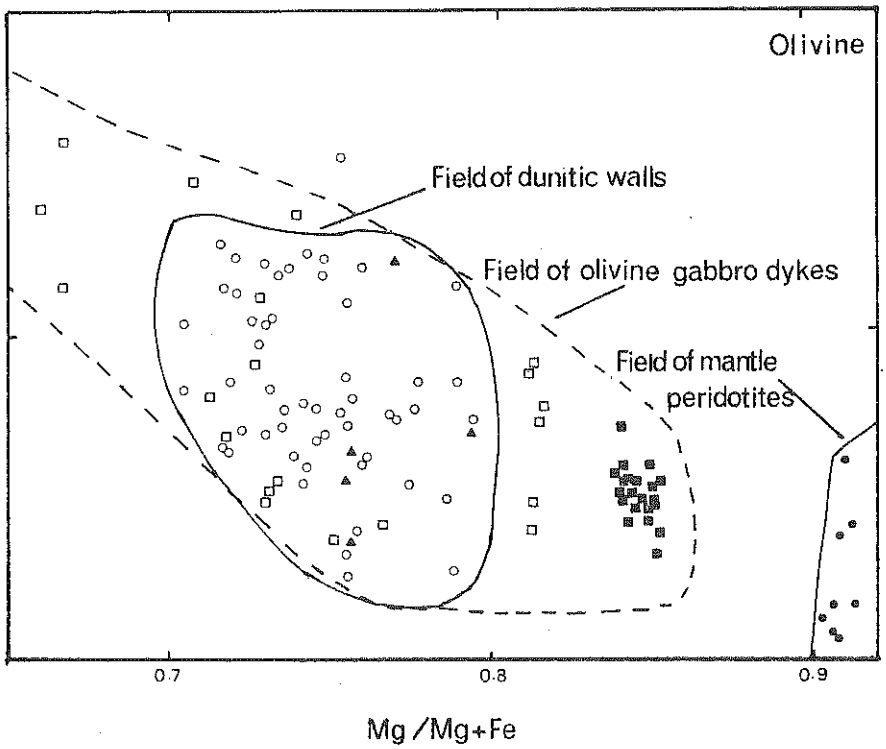


Fig. 16

Fig. 16 Mn content in olivines

tolitic gabbro, the forsteritic content of Ol increases : Fo_{85} in troctolite to Fo_{80} in Ol-gabbro. In the Monte Maggiore, a shear zone, 1 to 2 mm wide, cuts the peridotite before the injection of an Ol-rich gabbro. Once more the Ol composition is richer in Fa like in the dunitic walls. From the value of the distribution coefficient of iron and magnesium between Ol and liquid of 0.3 (ROEDER and EMSLIE, 1970) and from the temperatures estimated from coexisting pyroxenes (WOOD and BANNO, 1973, MORI, 1977, HERZBERG, 1978, POWELL, 1978), the composition of coexisting liquid were calculated. Layering I must have reequilibrated with a picritic basalt (MgO≃19 %) and layering II as well as the early Pl-rich veins with a magnesian basalt. Dunitic walls and the shear zone must have crystallized from an alkali and Fe-rich basalt (FeO = 10 % ; MgO = 11.5 %), whereas troctolite and Ol-rich gabbro were extracted from a tholeiitic basalt. Increase of Fe and Mn in these rocks are also compatible with a fractionation (fig. 16).

B - ORTHOPYROXENE

Opx is recognized in the earlier asemblages where it is associated with Cpx. During successive petrological events, it may develop and enclose Cpx, or Sp. It crystallizes together with Pl. In Pl-rich veins it is euhedral and enclosed in Pl. The mg^* ($Mg/Mg+Fe^{2+}$) ratio ranges between 0.9 and 1 (fig. 17A)for Layering I and II and Pl-rich veins, the highest values being in layered II rocks. In some early troctolitic dykes, Opx of mantle composition is included

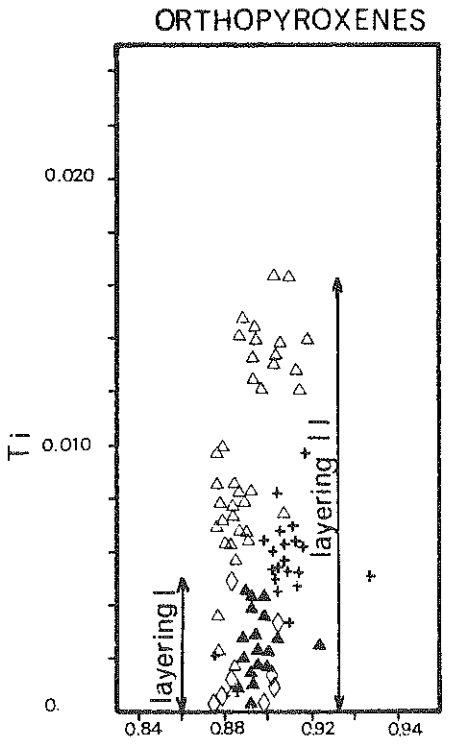


Fig. 17 A

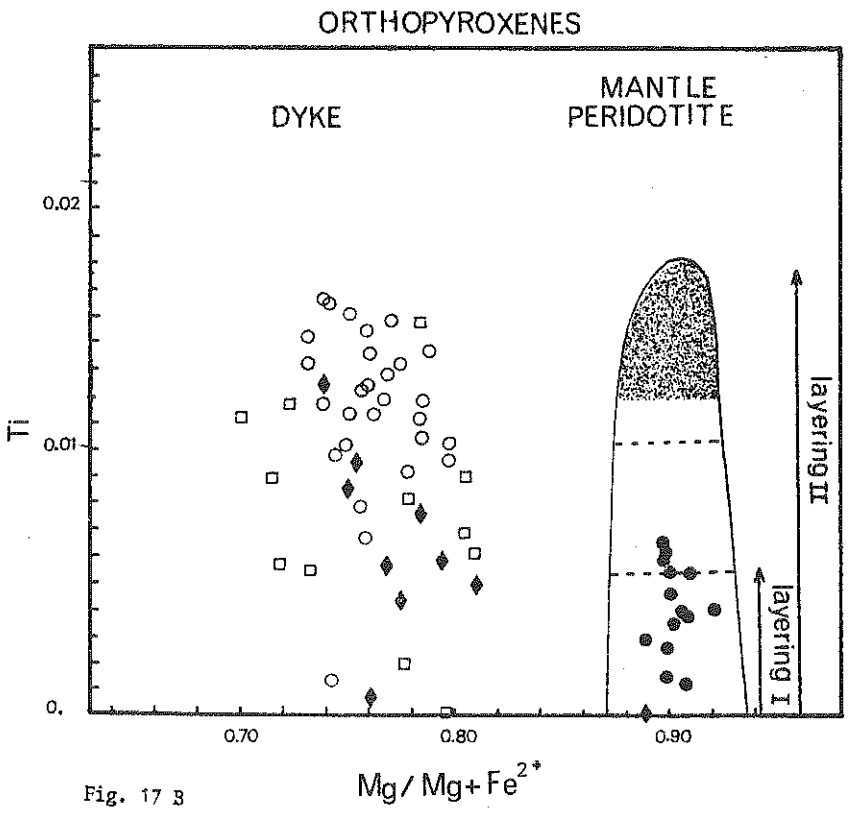


Fig. 17 B

Fig. 17 A and B Ti vs mg^* in orthopyroxenes.

in spinel. Opx exists only interstitially in Ol-gabbro. Its mg^* ratio ranges from 0.82 to 0.7 for Ol-gabbro, dunitic walls and shear zones. A plot of Ti versus mg^* (fig.17B) shows that layering II samples are generally richer in Ti than those from Layering I. Higher Ti contents are obtained from contaminated peridotites and some dunitic walls. Low Al_2O_3 content (fig. 18) is recorded from contaminated peridotites (2 % against 4 % in other peridotites). The proportion of the jadeite molecule decreases from layering I to layering II and contaminated peridotites. Opx from gabbros has the lowest Al_2O_3 contents.

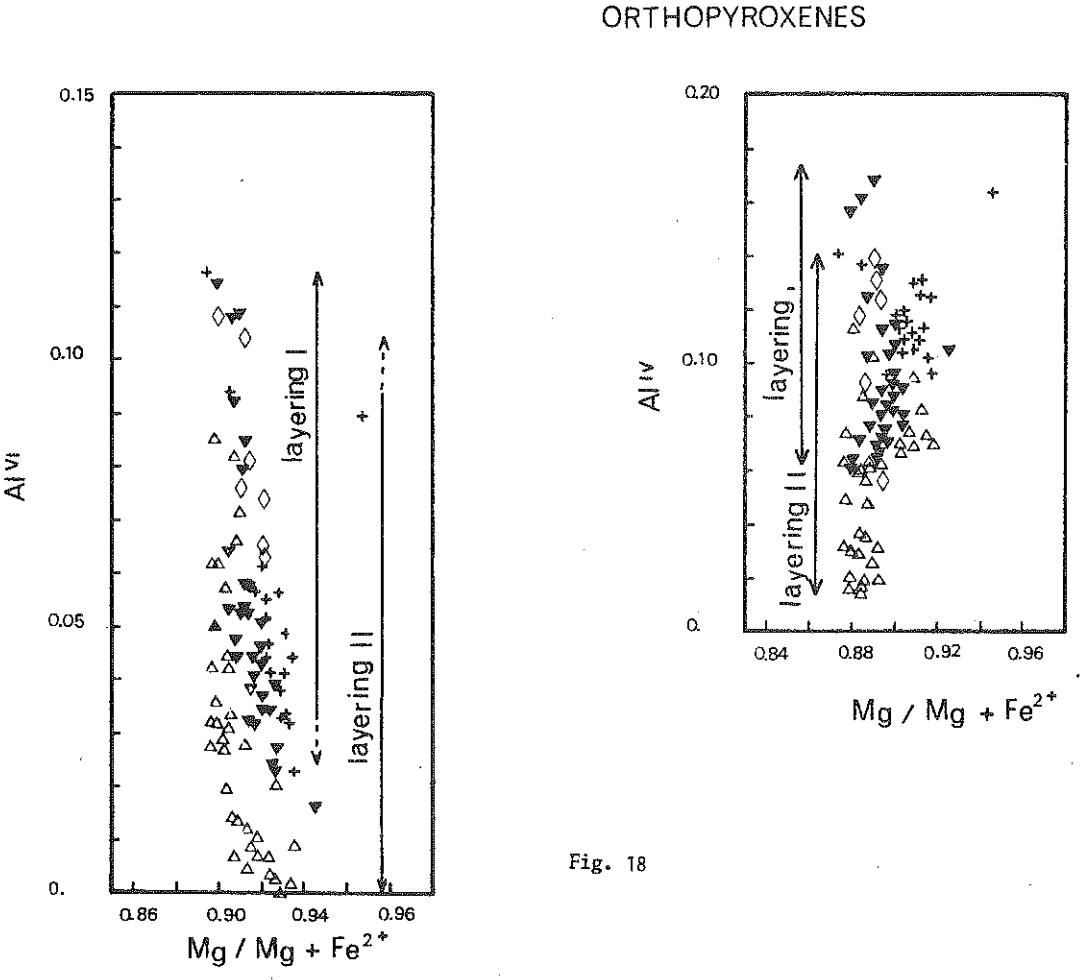


Fig. 18

Fig. 18 Al^{IV} and Al^{VI} vs mg^* in orthopyroxenes.

C - CLINOPYROXENE

Cpx is relatively abundant in Layering I. Its proportion decreases towards the centre of the massif where it is round and small. It occurs locally in Layering II and Pl-rich veins, where it may be euhedral. The mg^* ratio is similar to that in Opx. Al_2O_3 decreases in contaminated samples at the expense of TiO_2. A coupled increase of Na and Ti (fig. 19) is recorded from Layering I to Layering II and contaminated samples. In Layering I some Cpx are rich in the jadeite molecule. Al and Cr are lower in dunitic (fig. 20 – 23) wall, Ol-gabbro

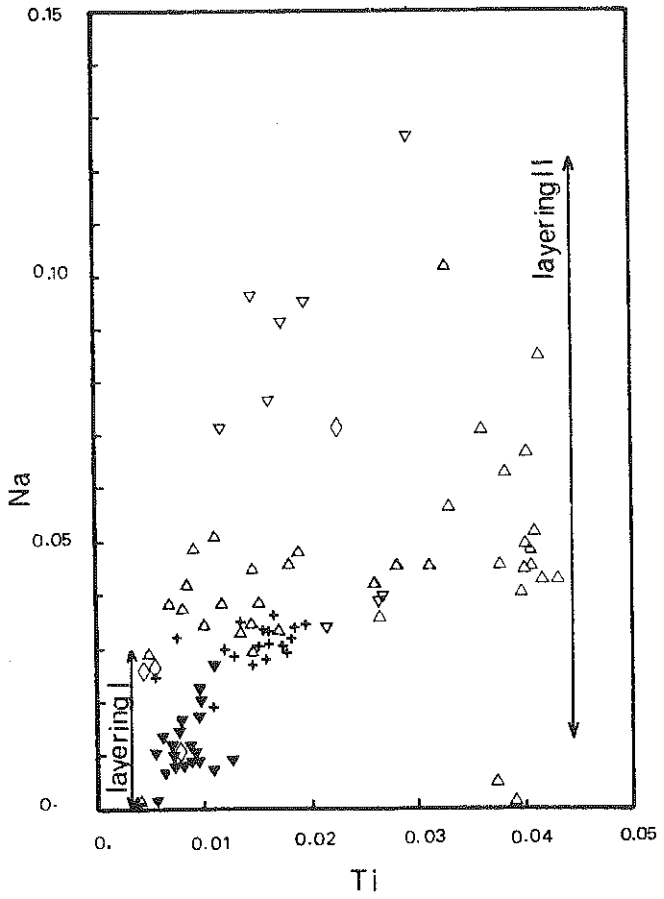
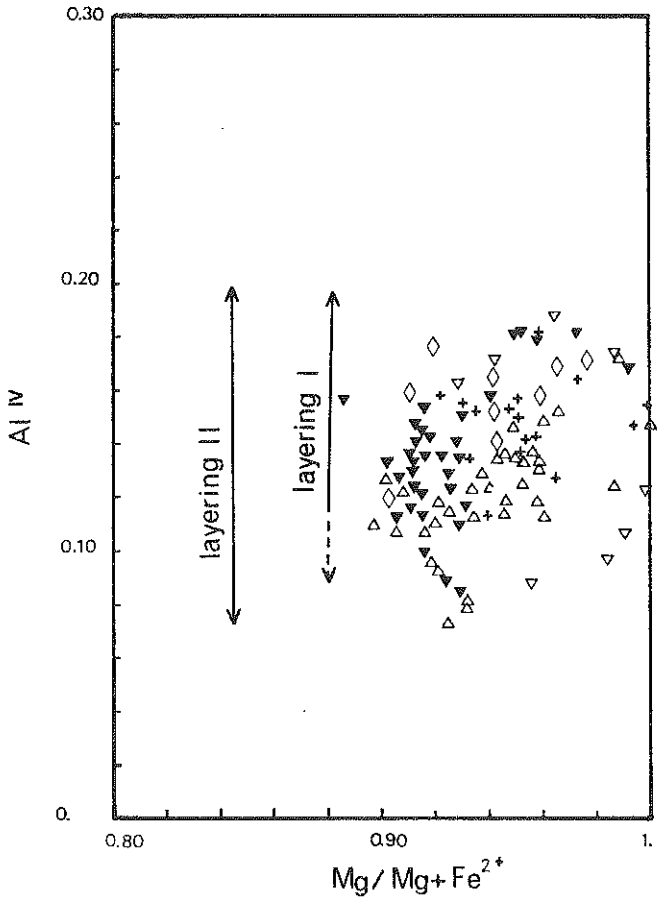


Fig. 19 Ti vs Na in clinopyroxenes.



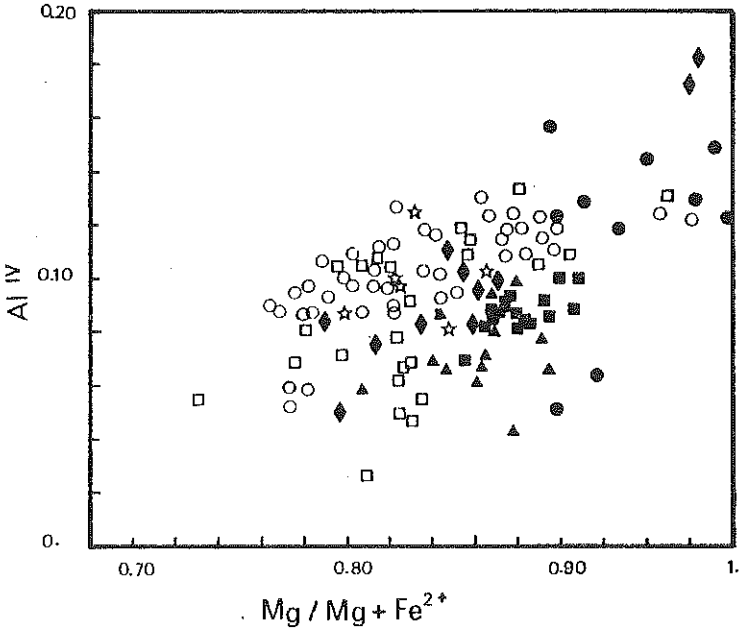
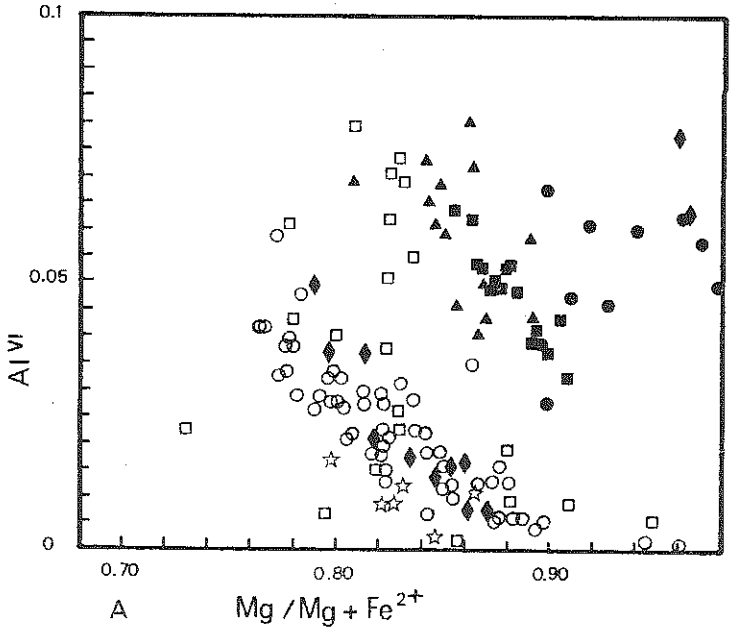


Fig. 20 Al^{IV} vs Mg/Mg+Fe in clinopyroxenes from gabbros and peridotites. Same symbols as in Fig. 15



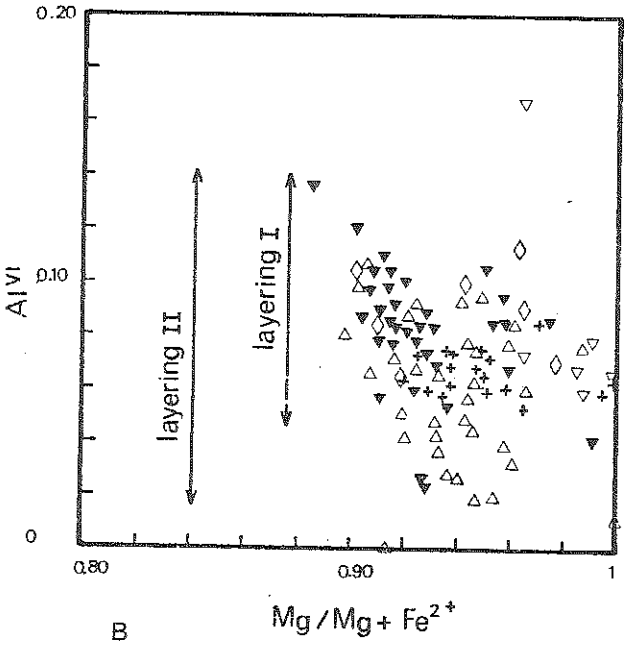


Fig. 21 Al^{VI} vs Mg/Mg+Fe in clinopyroxenes from gabbros (A) peridotites (B).

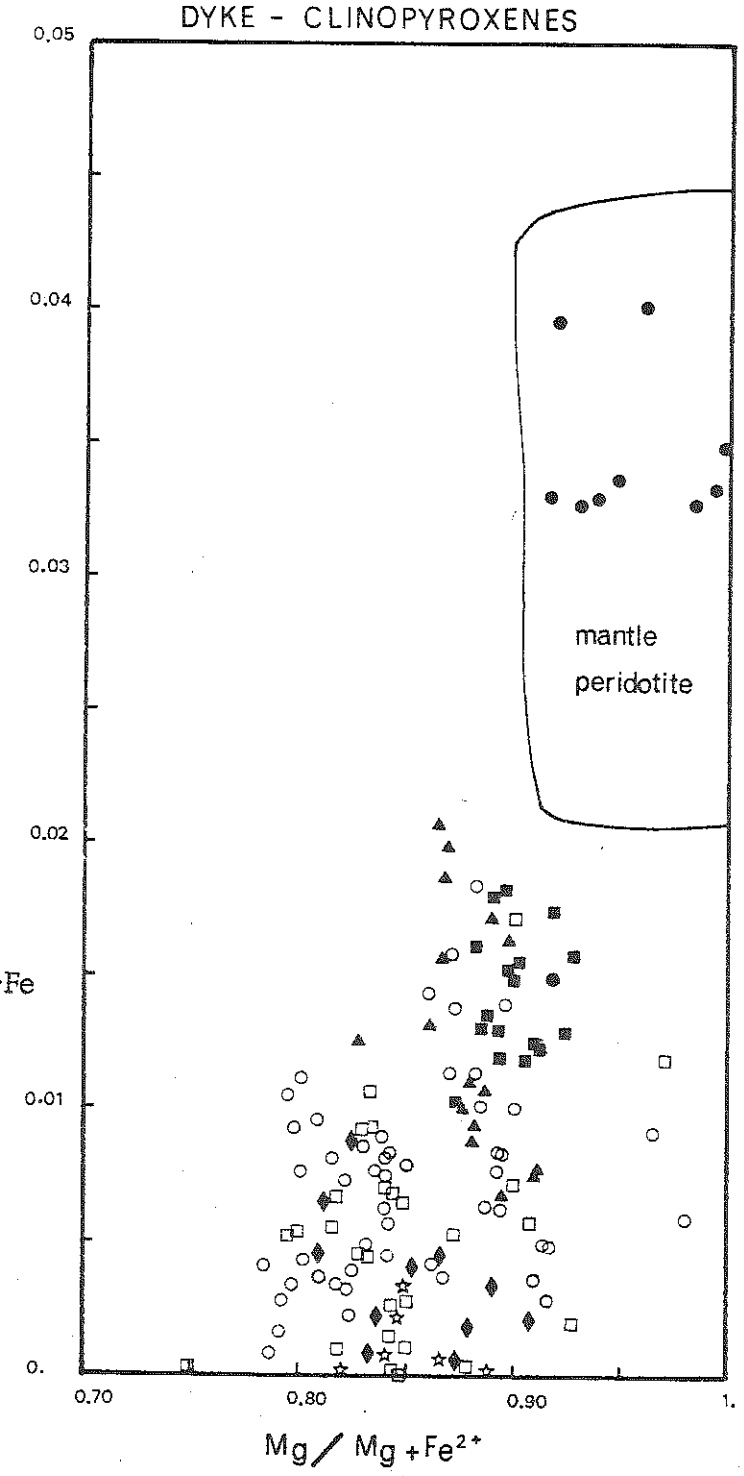


Fig. 22 Cr vs Mg/Mg+Fe in clinopyroxenes.

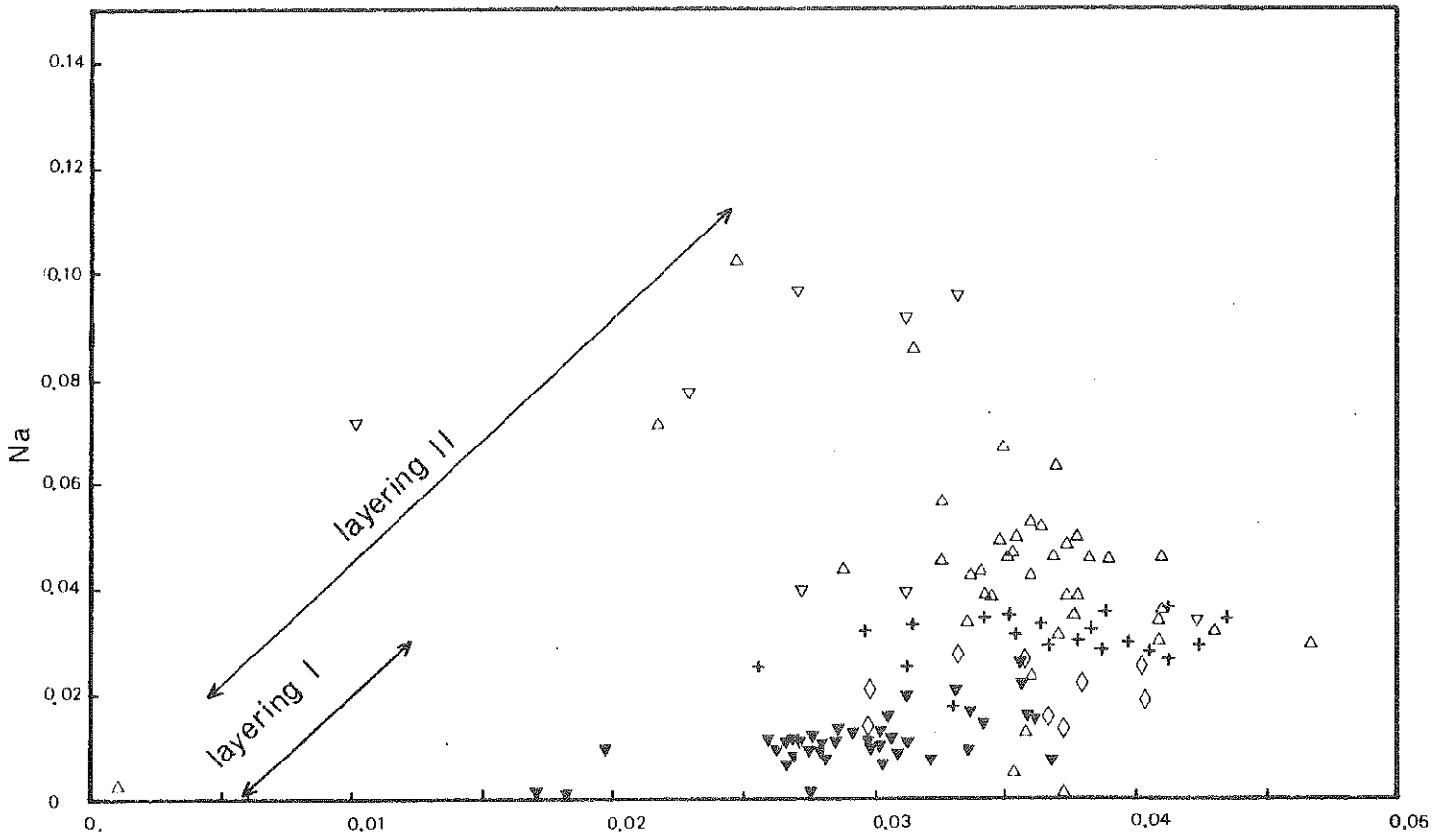


Fig. 23 Na vs Cr diagram for clinopyroxenes from the peridotites.

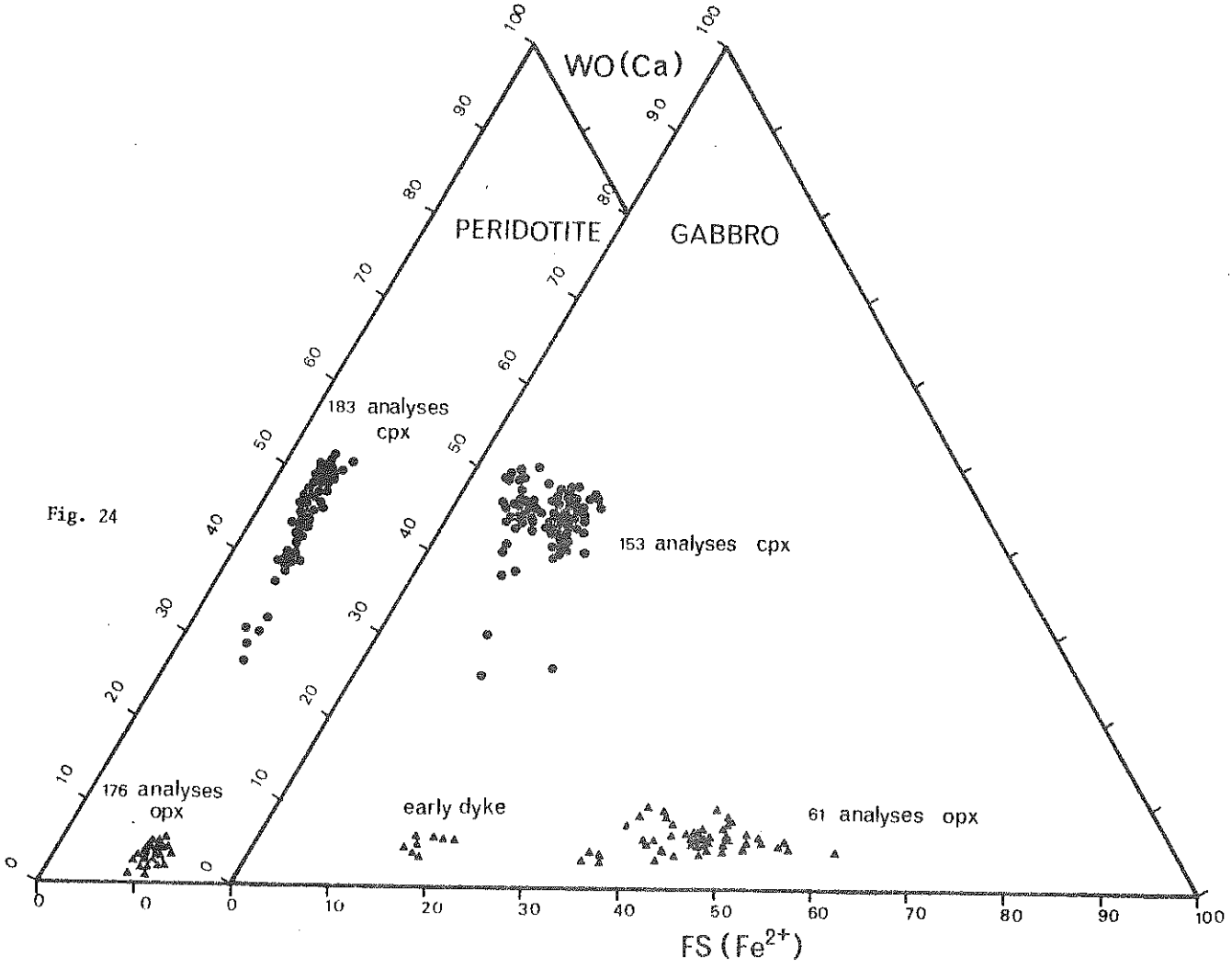


Fig. 24

Fig. 24 En-Wo-Fs diagram for clinopyroxenes and orthopyroxenes from the peridotites and the gabbroic dykes.

and shear zones than in the layered peridotites and Pl-rich veins. Tschermak's molecule decrease with differentiation. Pyroxene compositions (fig. 24) from the Monte Maggiore peridotites are similar to those reported from the Erro-Tobio area (ERNST and PICCARDO, 1979) and from the Bracco zone, western Liguria (COTTIN, 1978), as well as from the Lanzo massif in the Piedmont Alps (BOUDIER, 1976, ERNST, 1979). Compared with the dykes of the Monte Maggiore massif, Cpx from intrusive gabbro dykes at Lanzo shows less alumina and smaller amounts of chromium for a lower mg^* ratio. The compositions of pyroxenes from oceanic peridotite samples (HODGES and PAPIKE, 1976, SINTON, 1978) are similar to those reported here. Ti-rich Cpx from dredged peridotites are interpreted as cumulates crystallizing in the mantle (NICHOLLS et al., 1981).

D - SPINEL

Most of the chemical variation shown in peridotites are recorded by spinels. Mg and Al decrease from Layering I to Layering II and contaminated peridotites. The Al/Cr ratio (fig. 25) varies strongly in Layering I and more regularly with Fe/Mg ratio in Layering II. The Al decrease is accompanied by a Ti increase, especially in contaminated samples. Partitioning of Fe, Mg, Al and Cr between the coexisting phases varies with the Ti content (fig. 26, 27). In peridotites from ocean, a decrease of the Cr/Al ratio from core or rim has been noted (SINTON, 1978). A coupled Mg and Al increase exists in some spinels from the Indian ocean (BONATTI and HAMLYN, 1978).

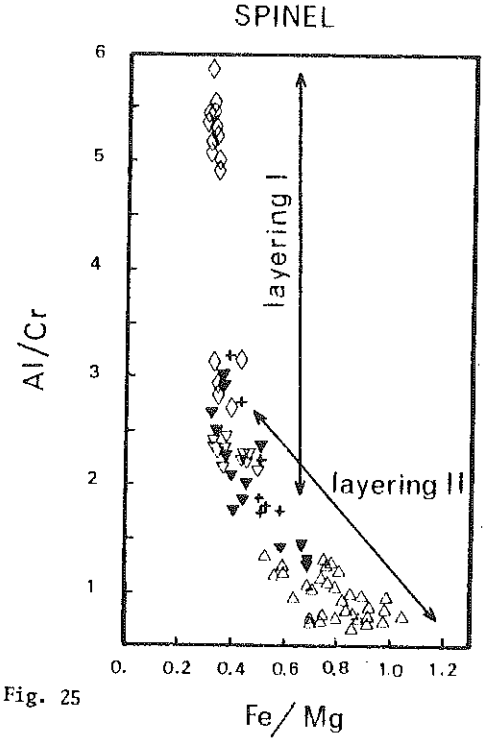


Fig. 25

Fig. 25 Al/Cr vs Fe/Mg in spinel from the peridotites.

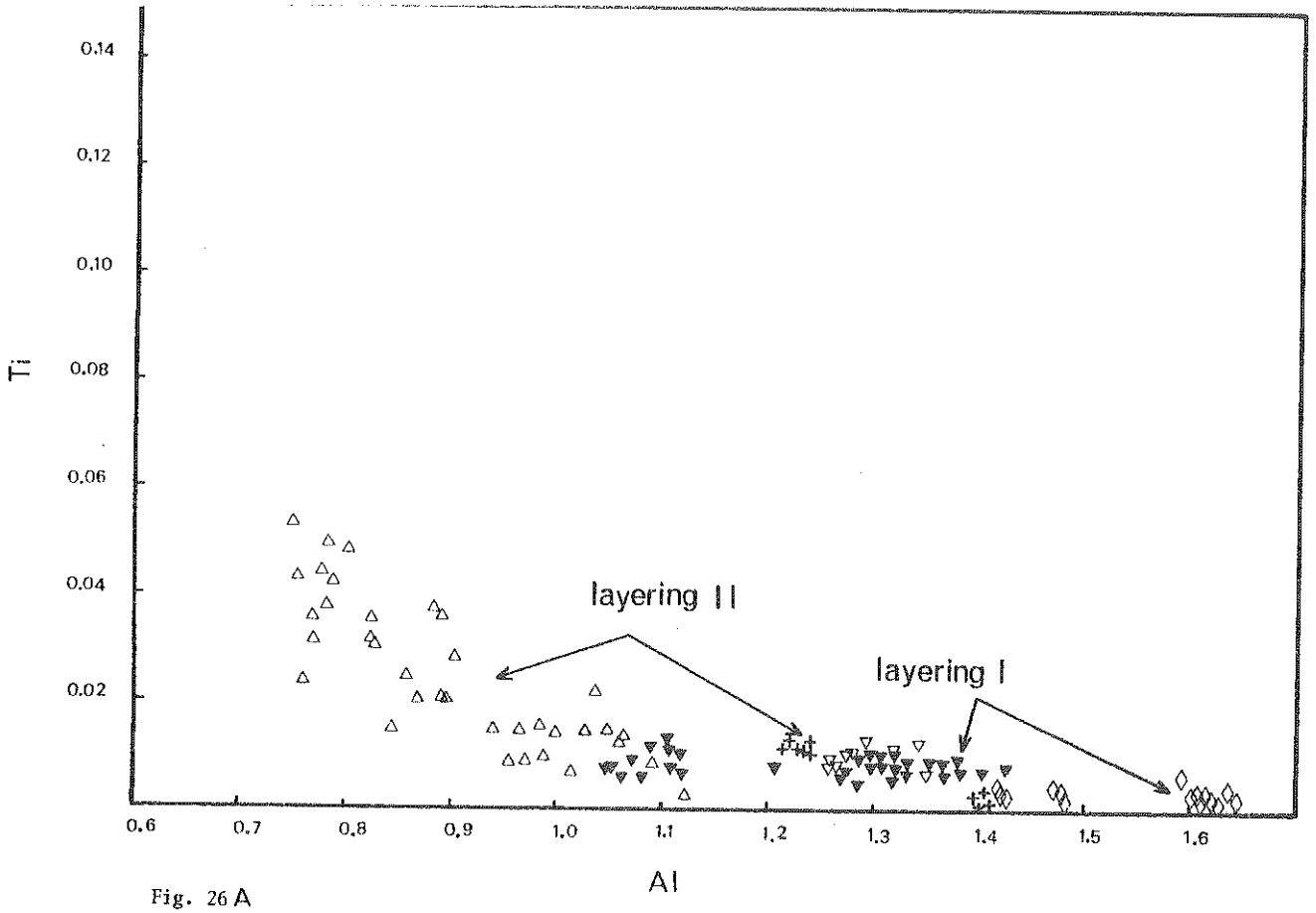


Fig. 26 A

Fig. 26 A Ti vs Al in peridotitic spinels.

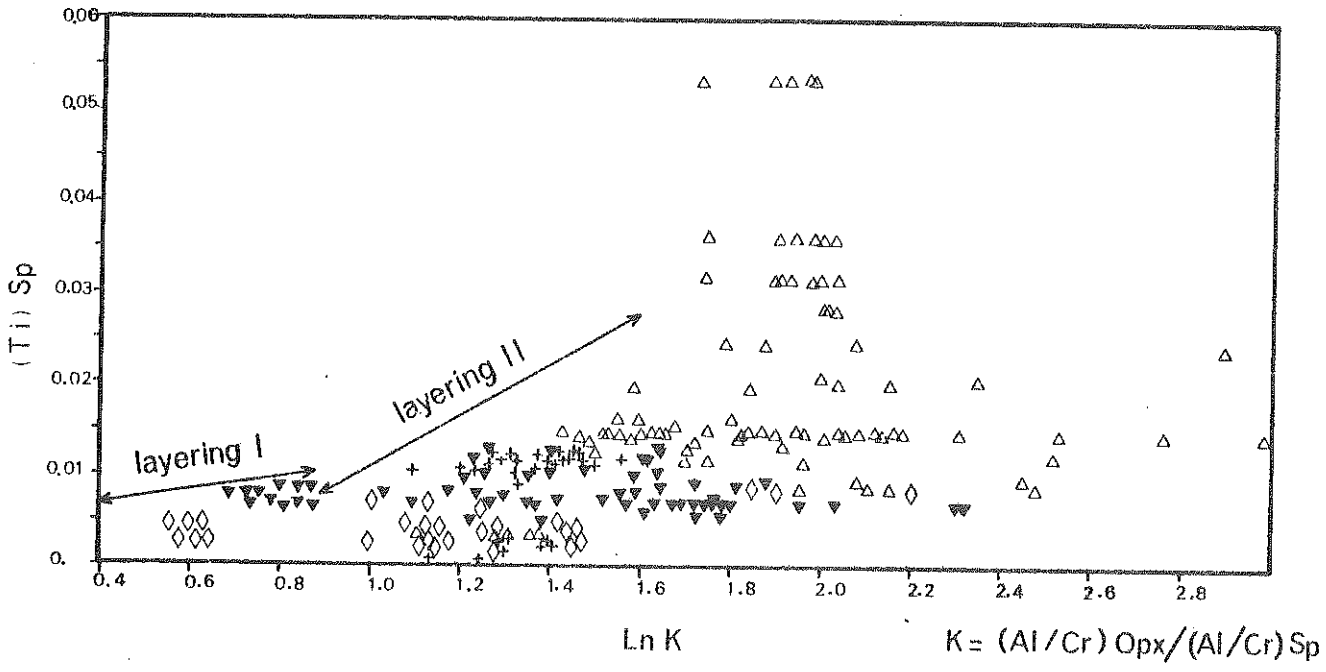


Fig. 26 B

Fig. 26 B Ti in spinel vs $K = (Al/Cr)_{Opx} / (Al/Cr)_{Sp}$

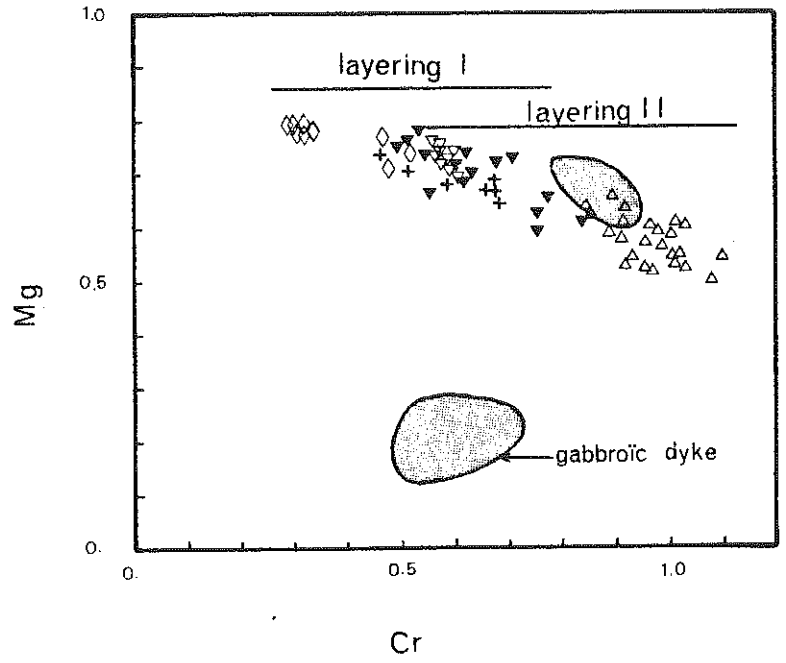


Fig. 27 Mg vs Cr in spinels.

E - PLAGIOCLASE

Pl occurs in the interstices in Pl-bearing peridotites. It is poecilitic in Layering II and some veins, crystallizing after Ol, Opx, Cpx and Sp. In early dykes, Ol, Sp and Pl crystallized together. Pl is earlier in troctolite and related gabbroic rocks than in the veins. Interstitial Pl from Layering II, vein and dunitic walls is near the bytownite-anorthite limit (fig. 28). In comparison

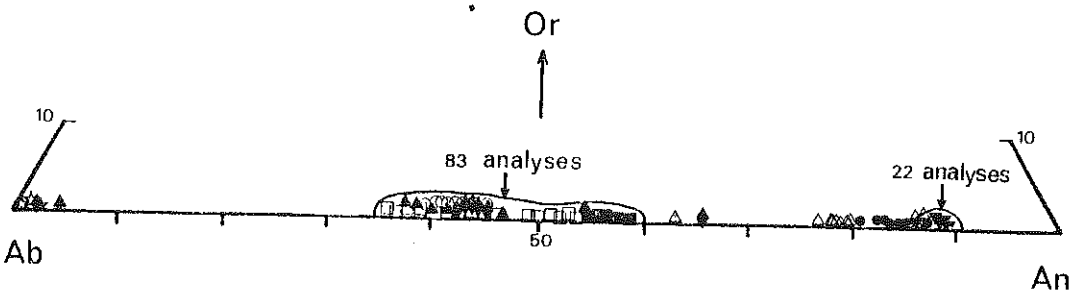


Fig. 28 Plagioclase composition from peridotites and gabbros in the Ab-An-Or triangle.

peridotites from the DSDP site 334 have relic Pl of An 89 composition (HODGES and PAPIKE, 1976). These Pl are more sodic than Pl from peridotites and lower cumulates from eastern-type ophiolites (An 95). In the Monte Maggiore massif, early dyke are slightly more sodic (An 84) (fig. 29), whereas shear-zone and dunitic walls have Pl of andesine composition (An 44), like Pl from some ocea-

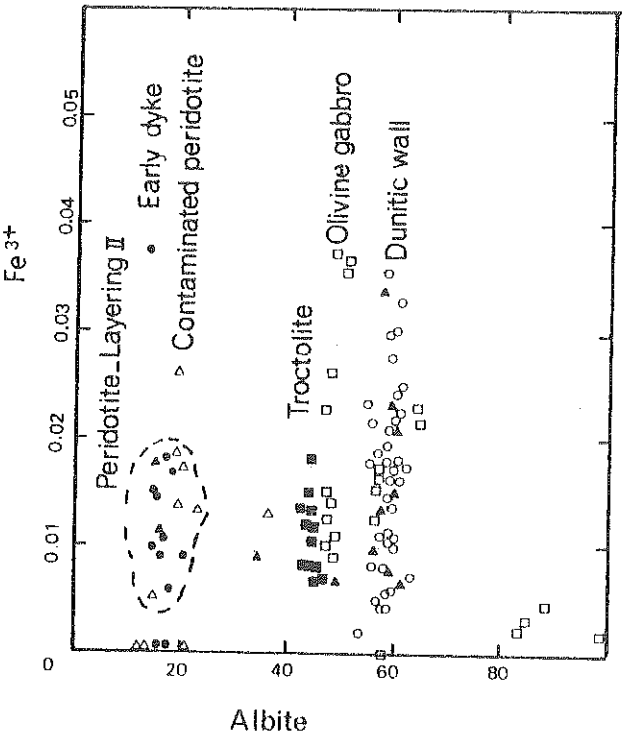


Fig. 29 Fe^{3+} content in plagioclase vs Ab content.

nic peridotites (MELSON et al., 1972, SINTON, 1979). Dykes of type II from the Lanzo massif have also such low values (BOUDIER, 1976). The K content of sodic Pl also increases in the contaminated samples, shear zones and dunitic walls. Troctolite and Ol-gabbro have more calcic Pl falling in the labradorite range (An 55).

F - AMPHIBOLE

Some brown hornblendes seem to be primary : they occur in some microscopic veins in contaminated peridotites together with apatite, Ti-rich Sp and Cpx. They exist also in some inclusions enclosed in pyroxenes related to some micro-fractures which destabilized the previous phases. In the dunitic wall, they are included in Sp or Ol. Amph. is also associated with sodic Pl of the shear zone in peridotites. According to LEAKE's classification, Amph. is pargasite, pargasitic hornblende, kaersutite and hastingsite (fig. 30). These phases were found in the mylonitic samples from St Paul's rocks (MELSON et al., 1972). Other amph. from the Monte Maggiore massif, such as magnesio-hastingsitic hornblende, magnesic-hornblende and actinolite, are related to metamorphic processes.

The mineralogical data indicate that Layering I may be distinguished from Layering II, as well as from contaminated samples. Shear zones and dunitic walls are related to the circulation of a magma rich in alkalis, iron and

incompatible elements. Chronological relationships indicate that this magma migrated later than the formation of Layering II and before some Ol-gabbros and

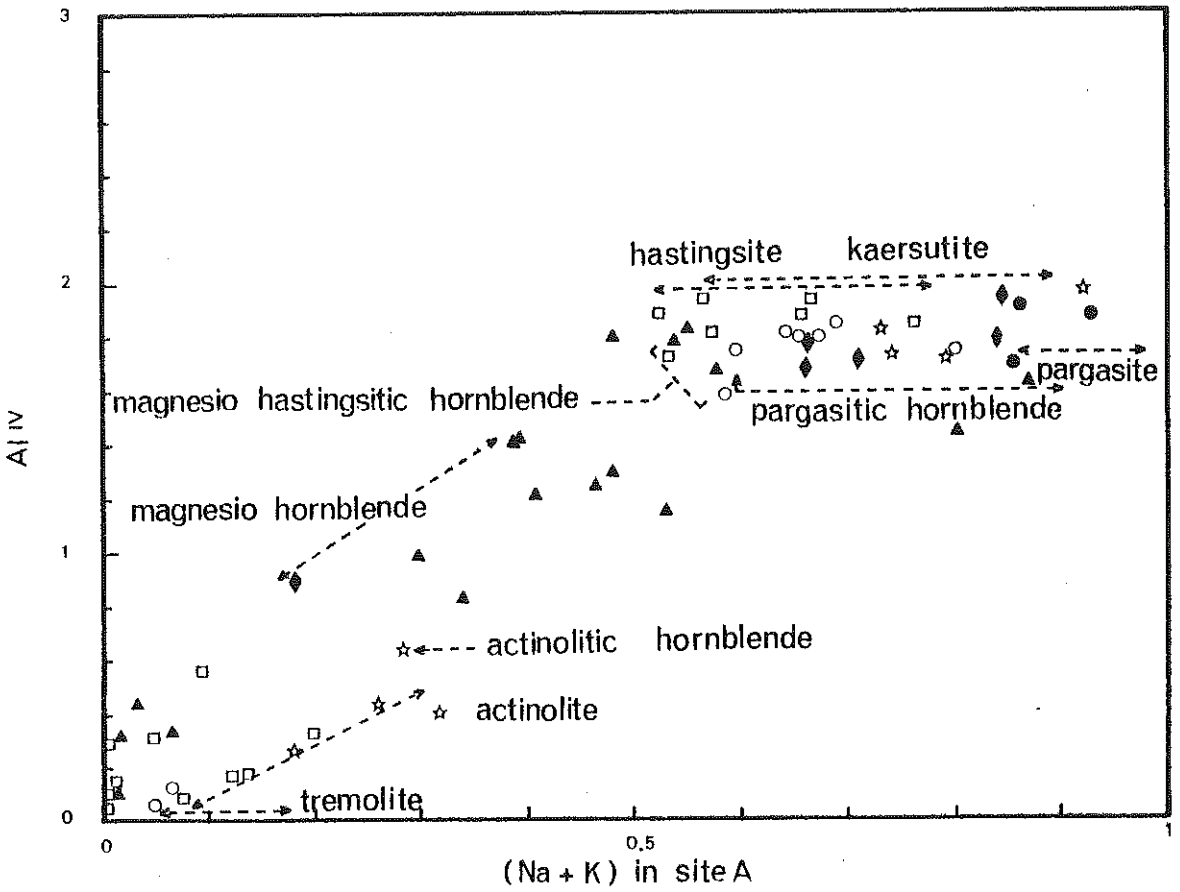


Fig. 30 Amphibole composition in gabbroic and peridotitic rocks.

troctolites. Layering II, Pl-rich veins and some early dykes crystallized at lower temperatures than samples from Layering I. Liquids (magnesian basalts) which equilibrated with peridotites from Layering II was more TiO_2-rich than those associated with Layering I (picritic basalt). Rocks from Layering II and Pl-rich veins may be related to fractionation from an ascending magma. At high pressure some Sp-peridotites with Al_2O_3-rich Sp may have a similar origin. Samples from the Layering I were equilibrated during partial fusion which occurred before fractionation, that is in the spinel or aluminous-pyroxene facies.

V - TOWARDS A POSSIBLE INTERPRETATION

The structural and petrographical maps, the chronology of the dykes and the mineralogy of rocks from the Monte Maggiore massif indicate that numerous magmatic events occurred under tectonic control during the uprise of a mantle diapir.

The petrographical map reveals that some areas along the coast and towards the southern edge of the massif consist of well folded and layered Cpx-rich peridotites, which were affected by partial fusion. Evidence for this comes from a decrease in size and proportion of Cpx towards the central part of the massif, the destruction of previous structural features like folds and layering More depleted peridotites form elliptical pods cut by the late products (Layering II). The variation of fusible and incompatible elements against the FeO-MgO content of the samples leads to the conclusion that the residual peridotites have an enriched character, which is due to late crystallization of Pl-Sp-Opx from circulating magma. The higher Ti-content of the major phases in the Layering II compared to Layering I is also in agreement with equilibration from a TiO_2-richer magma. Rocks related to Layering I are residual (pyroxenite to lherzolite and Cpx-poor lherzolite).

The chronology, the structural positions of the dykes and their mineralogical assemblages indicate that several magma pulses were injected at various depths. They are generally related to fractionation processes even if the dykes occupy similar planes as during partial fusion.

In the oldest assemblage, liquids migrated along grain boundaries in numerous directions-parallel to layering I and the foliation plane,in sub-horizontal planes and in conjugate and tensile area around an E-W trending zone. With time circulating liquids occupied better defined planes in tensile zones and along shear planes. The injection of gabbroic dykes led to the re-opening of all previous fractures. Generally the gabbroic dykes are strongly deformed, a fact compatible with an increase of deviatoric stress towards the surface as in other peridotitic massifs. Temperatures of equilibration and compositions of coexisting phases indicate that Layering I was in equilibrium with a picritic magma. Later asemblages equilibrated with a magnesian basalt. Some dunites along dykes and shear zones in peridotites crystallized from a magma rich in alkali, iron and incompatible elements, which migrated before the injection of tholeiitic magma. Circulation of liquids rich in incompatible elements is also shown by the contaminated peridotites having apatite, Ti-hornblende, Ti-Spinel and Ti-Cpx besides the earlier asemblage.

The petrographical and structural features will now be put in a chronological schema, which will then be discussed. All the fractionated samples occupy distension and shear zones related to a N 110º maximum compression di-

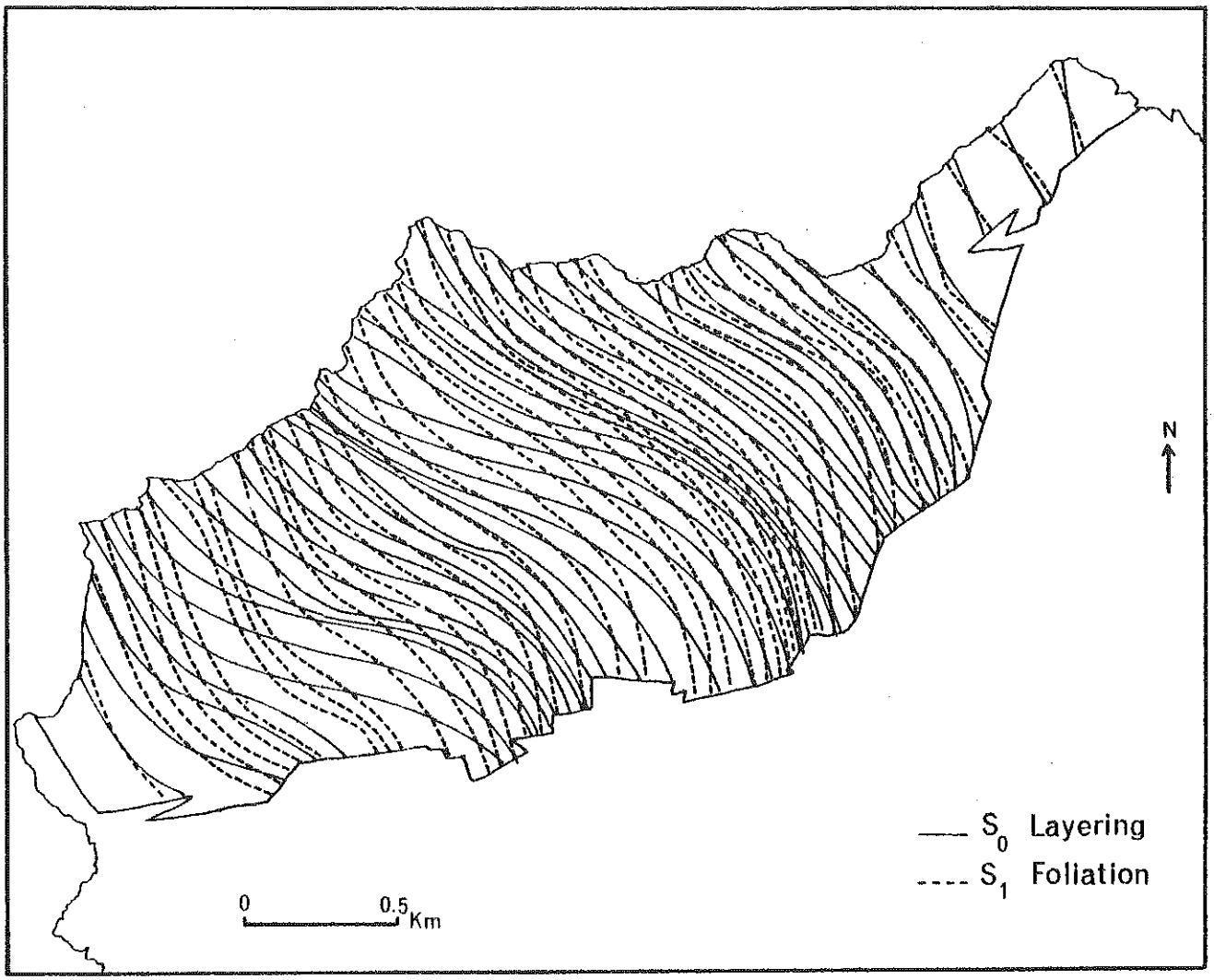


FiG Fig. 31 Intersection between foliation and layering. Layering I is nearly N-S; layering II nearly E-W.

rection. Lineation and the fabric of the peridotite is compatible with horizontal flowage. The fold axes in Layering I, which are at a large angle to the mineral lineation (fig. 31), may indicate an earlier least compressive direction with a steeper direction. Sub-vertical layering and fold axes are compatible with the uprise of a diapir. This mechanism may facilitate the opening of slightly inclined fracture zones which are later occupied by lower P products. In a mid-oceanic ridge environment, a diapir may come from the low velocity zone at about 25 Kb. It probably equilibrated in the Al-pyroxene facies, because of the high geothermal gradient in this magmatic zones (BOTTINGA, 1974). Although the peridotites were later reequilibrated in the Sp or Pl facies, no relic forms of garnet have been observed. Partial fusion occurred during the adiabatic uprise before fractionation possibly in the Sp facies. Incongruent melting of an Al-rich pyroxene may have occurred. The thermal cusp at a facies change may facilitate partial fusion. As the earlier recognizable fracture zones occupy numerous directions, primary magmas are thought to separate from the residual minerals. At least 10 % of melt may exist with peridotite

MAALOE, 1981) without changing the fabric of the unmelted peridotite, as indicated by experiment. In the field the position of the more depleted zone shows that the degrees of partial fusion varied. The setting of the residual peridotites depends on the position of the applied stress. Shear zones may preferentially channel heat flow (MC BIRNEY, 1963). In this zone a "thermal trap" may persist (YODER, 1976) as long as is necessary for partial fusion. Circulation of magma in extension zones may propagate partial fusion. The migration of alkali and incompatible-element-rich liquids occurred later. We may suppose that some alkali-rich veins existed and initiated partial fusion along some specific planes by lowering the temperature of the liquidus of the peridotites. These fluids may also reduce the normal stress which favours the propagation of cracks (YODER, 1976) and the extent of partial fusion. Collection and departure of the magma towards the surface is controlled by the degree of permeability of the peridotite (MAALOE, 1981).

As the residual peridotites rise, its temperature progressively decreases and the mantle is always invaded by magma coming from deeper zones. Magma migrates along planes where the previously extracted melt circulated. At first, hot magma whose temperature is closed to those of the enclosing peridotite may circulate in numerous veins, or along grains boundaries. Minerals with the highest melting temperature will crystallize first : Ol-Opx and some Sp at the higher pressure (fig. 32, 33). As crystallization occurs the volume of the rock increases and the smaller joints progressively vanish. Magmatic veins then occur in better defined planes corresponding to tensile fracture zones and locally to en-echelon fractures near the flow plane of the peridotite. This process is also facilited by an increase in the deviatoric stress. Simultaneously, as the temperature decreases, more phases crystallize sometimes because magma is trapped. Ol-Opx-Sp-Cpx and Pl crystallized in veins parallel and slightly oblique to the Layering II. Cooling of the mass and higher viscosities progressively decrease the number of migration planes. Fractionation occurred from the Sp facies, through the Seiland facies to the Opx-Pl-lherzolite facies (fig. 34). The special occurrence of alkali-rich fluids in the Seiland and especially in the Opx-Pl lherzolite facies may also result from a particularly advanded fractionation at high pressure. In the 18-10 Kb range, results from experimental petrology show that alkali-basalt may derive from a tholeiitic basalt through the fractionation of Al-rich enstatite between 13.5 and 18 Kb (GREEN and RINGWOOD,

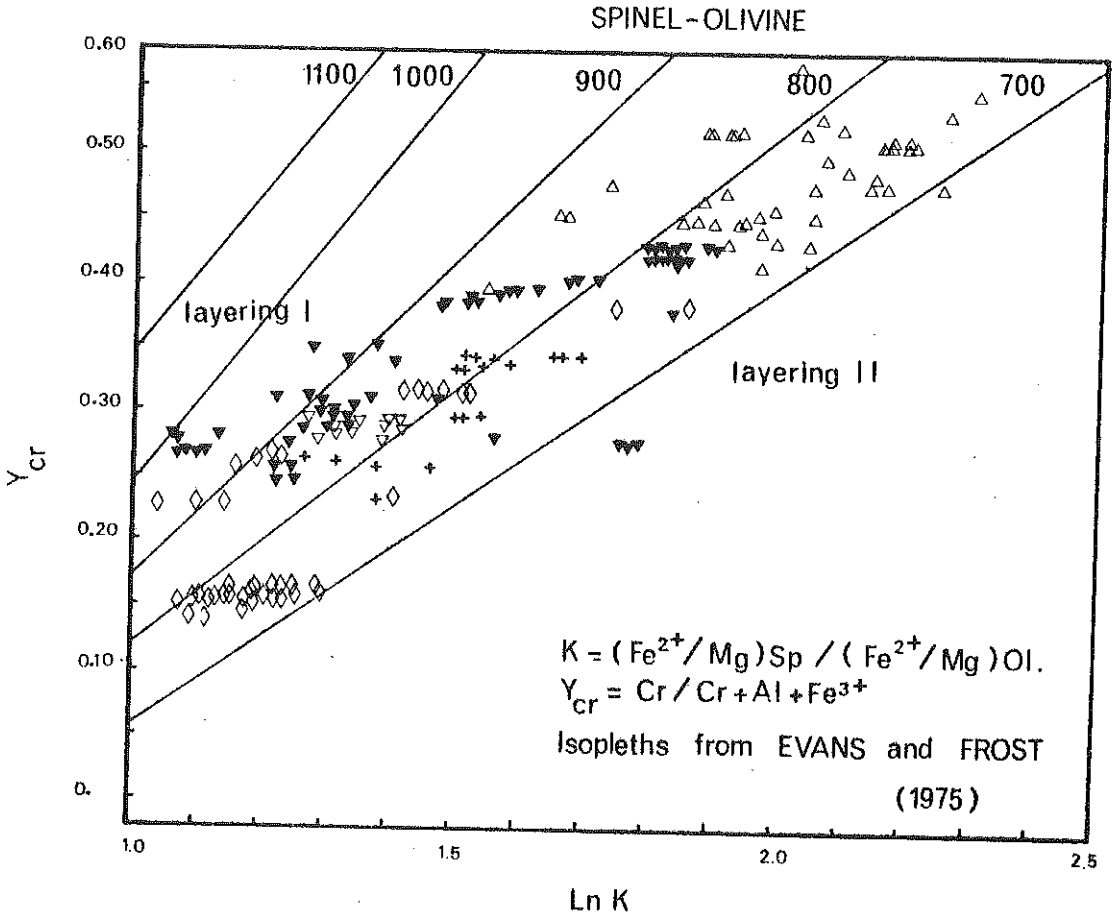


Fig. 32 Y_{Cr} vs Ln K

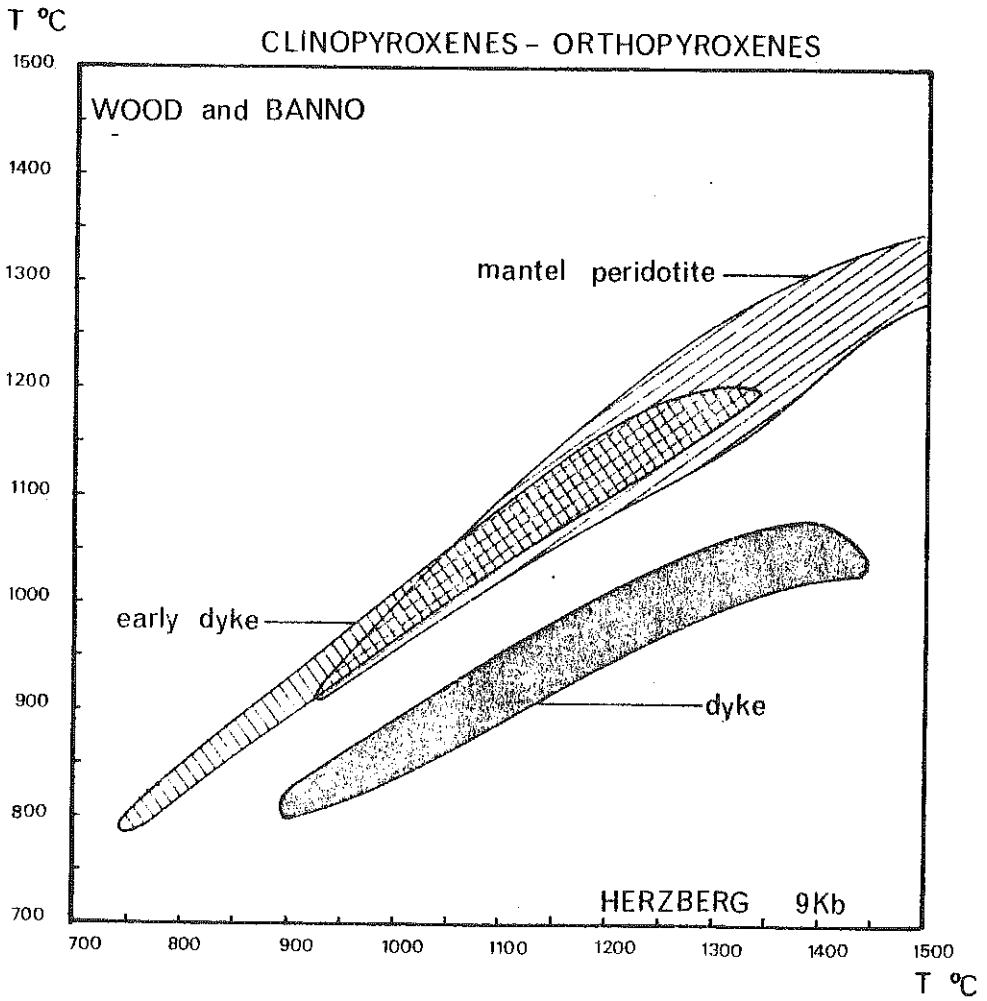


Fig. 33 T) Fig. 33 T°C WOOD and BANNO vs T°C HERZBERG at 9 Kb.

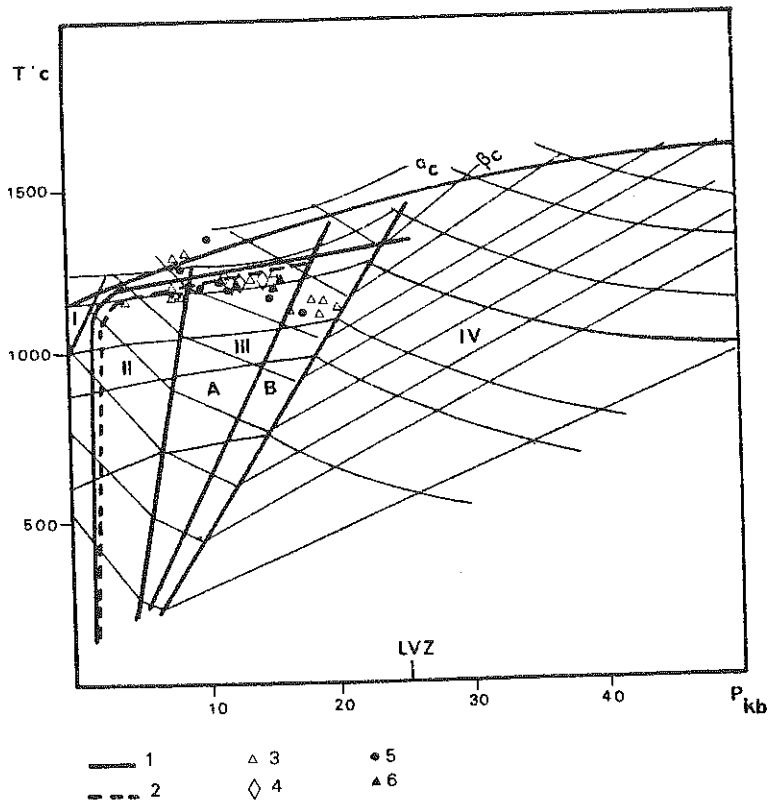


Fig. 34 Petrogenetic gird after O'HARA (1967).

I) protohypersthene-plagioclase-lherzolite field

II) orthopyroxene-plagioclase- lherzolite facies

III) spinel lherzolite facies

A) Seiland subfacies (diopside-enstatite-anorthite-spinel)

B) Ariegite subfacies (diopside-enstatite-garnet-spinel)

IV) Garnet lherzolite facies

$$\alpha_c = \text{wt \% } CaSiO_3 \times 100 / \text{ wt \% } (CaSiO_3 + MgSiO_3)_{Cpx}$$

$$\beta_c = \text{wt \% } Al_2O_3 \times 100 / \text{ wt \% } (CaSiO_3 + MgSiO_3 + Al_2O_3$$

1) oceanic geotherm after BOTTINGA (1974) at 8cm/y spreading

2) " " " " " 2cm/Y "

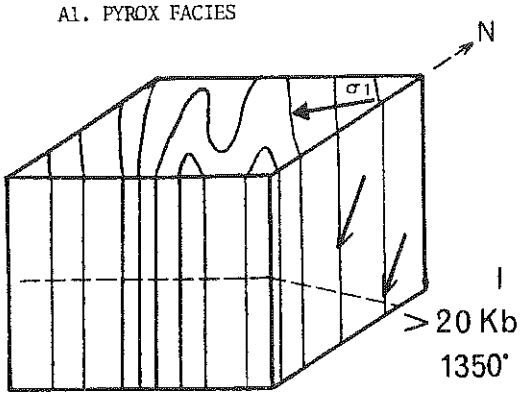
3) contaminated lherzolite

4) spinel lherzolite

5) layering I

6) layering II, plagiferous lherzolite

A1. PYROX FACIES



EARLIER RECOGNISIBLE STAGE

COARSE GRAINED AND CPX RICH PERIDOTITES

PURE SHEAR

HIGH HYDROSTATIC PRESSURE

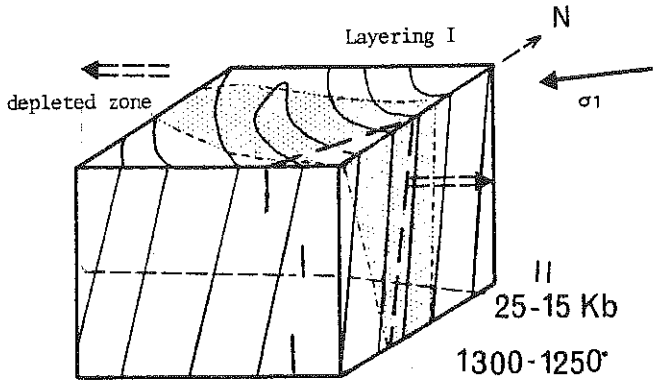
MECANISME OF PRESSURE SOLUTION CREEP

LOW STRESS AXIS SUB // TO THE FOLD AXES

POTENTIAL PURE EXTENSIONAL FRACTURE SLIGHLY INCLINED

DEPARTURE FROM LV z (25 Kb) OF SOLID DIAPIR

HIGH - PRESSURE ZONE OF THE SPINEL FACIES



RAPID ADIABATIC UPRISE

ROTATION OF THE DEVIATORIC STESSES

ACTION OF FLUID PRESSURE

INITIATION OF THE FUSION AT THE INFLECTION SURFACE OF

OF S_o IN A HEAT FLOW CHANNEL

PRODUCTION OF PICRITE

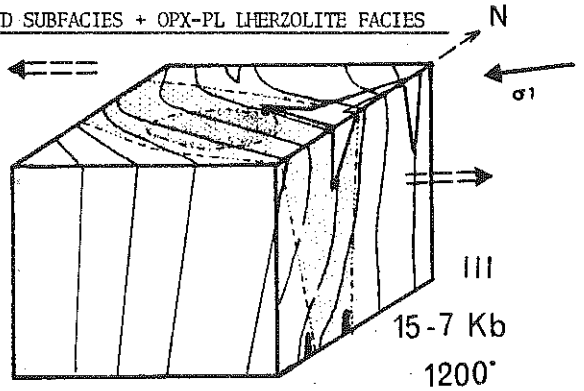
RESIDUE OF CPX-POOR LHERZOLITE

HP FRACTIONATION: OPX+OL+SP 15-20 Kb. 1250°.

MIGRATION: $//S_o$ IN TENSITE ZONE (N 70).

PARTIAL FUSION EVENT . BEGINNING OF HP FRACTIONATION

SEILAND SUBFACIES + OPX-PL LHERZOLITE FACIES



Layering CPX rich PERIDOTITE

SLOW UPRISE OF THE DIAPiR

MAGMA MIGRATION:

LAYERING II

VEINS IN TENSITE AND CONJUGATE PLANES.

FRACTIONATION/: OL.POX.CPX.SP.PL

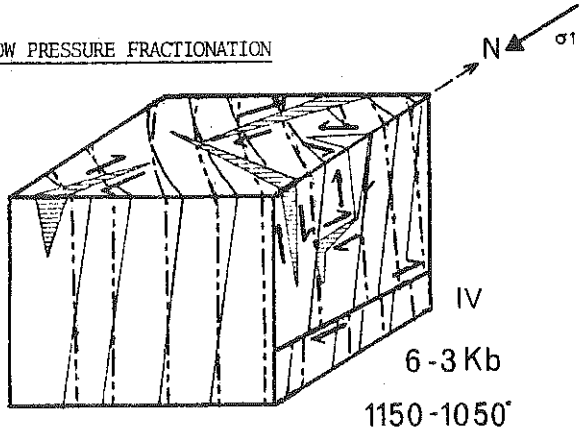
MAGNESIAN BASALT BASALT OF OPX AND SP FRACTIONATION TOWARDS SURFACE.

CONTAMINATION : ALKALI AND INCOMPATIBLE ELEMENTS IN SMALL CRACKS AND SHEAR ZONES

FORMATION OF SOME INDIGENOUS DYKES (DUNITES).

FRACTIONATION - CONTAMINATION

LOW PRESSURE FRACTIONATION



INTRUSION OF BASALT UNDER DEVIATORIC STRESSES.

SP.OL.CPX.PL. FRACTIONATION
- TROCTOLITE
- OLIVINE GABBRO
- FERROGABBRO

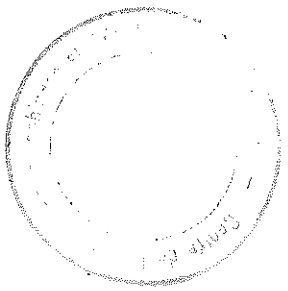
BLOC MOVEMENTS:

NORTHERN BLOC: DOWN AND TOWARDS EAST.

UPPER BLOC: TOWARDS NORTH.

DYKE INTRUSION

1967). Fractionation of Ol+Opx from an oceanite magma (Ol-rich liquid) may lead to an alkali basalt (PRESNALL et al., 1978). Fractionation of Opx-Cpx and Pl at 8-10 Kb leads also to alkali-rich liquids (O'HARA, 1965, 1968). Some veins rich in incompatible elements seem to occur early. Liquids have been equilibrated with Mg-rich phases. Alkali-rich veins from shear zones and some dunitic dykes seem to be later. They cut the Layering II. The composition of olivine in these dykes testifies to fractionation before emplacement. Shear zones correspond also to the first appearance of a porphyroclastic texture in the massif. They are linked to an increase in deviatoric stress (fig. 35). Wall rock reactions are also related to the intrusion of the dunitic dykes, when the temperature of the enclosing peridotite is still high and its liquidus lowered by the composition of the melt. In more superficial zones, no wall rock reaction occurs. The gabbroic dyke have crystallized from several magmas mainly tholeiitic, emplaced in previous fracture zones which have been re-opened by more or less alkali-rich magmas and fluids. They facilitated the propagation of the tholeiitic melt during its ascent after the peridotic fractionation. Ol-gabbro may occur towards the centre of a dyke bordered or not by dunite. Gabbroic dykes are deformed because the whole fracture zone may not correspond exactly to an expansion plane at higher levels. The rheological state of the magma may change rapidly because of the low temperature of the enclosing peridotite.



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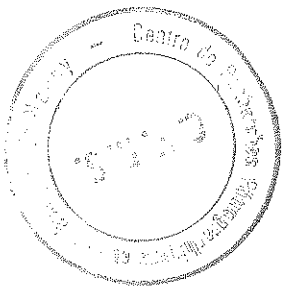


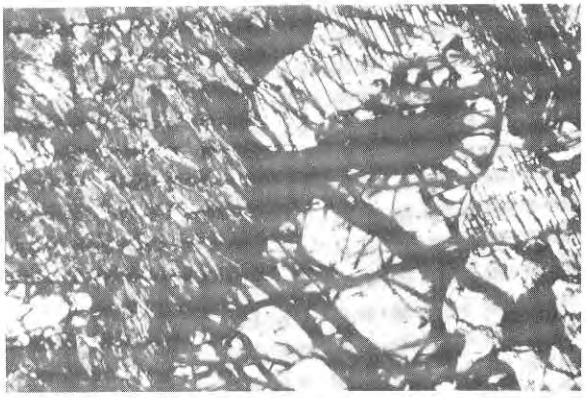
P L A T E I

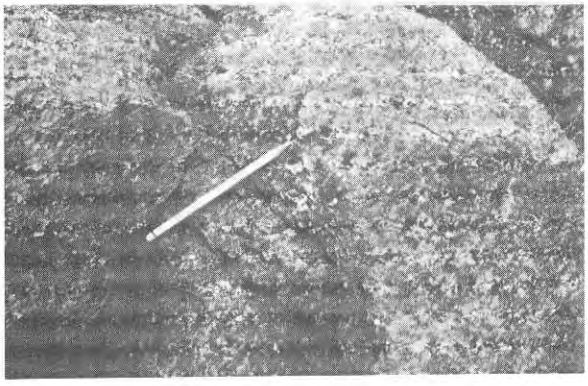
Fig.1 - Cpx-poor lherzolite with fine grained Cpx generally enclosed in Ol or Opx.

Fig.2 - A detail of the replacement of Cpx by Opx and Ol in pyroxenites. 2,5mm; cross polar.

Fig.3 - Disappearance of layering: pyroxenites are replaced by an assemblage of Ol + Sp $\pm$ Opx during partial fusion event.

Fig.4 - Layering I: folded pyroxenite, Cpx-poor lherzolite and harzburgite alternate.





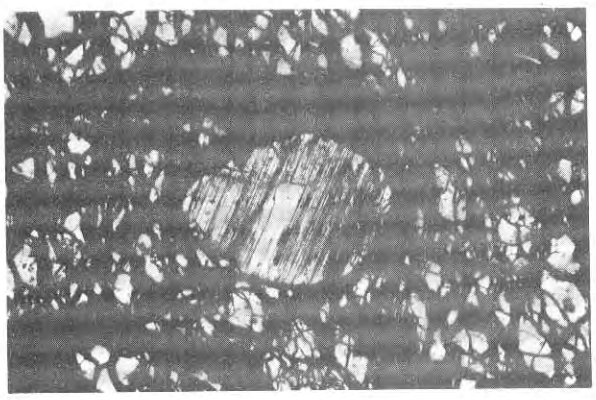
P L A T E II

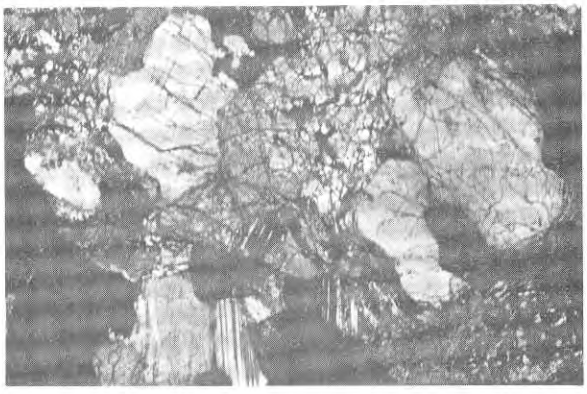
Fig. 1 - Rounded Cpx enclosed in poikiloblastic Ol. 2,5mm; cross polar.

Fig.2 - Coarse grained harzburgite with knobs of Opx developed during partial fusion and high pressure fractionation.

Fig.3 - Pl-lherzolite showing Pl concentration along minerals.

Fig.4 - Plagiferous lherzolite related to the layering II. 2,5mm; cross polar.





P L A T E III

Fig.1 - Contaminated peridotite. Destabilization of Opx. Cristallization of Ti-rich Cpx, pargasite, Ti-rich Sp along micro veins cutting the rocks. 2,5mm; cross polar.

Fig.2 - Development of apatite, pargasite, Ti-rich Sp and Pl along fractures in Cpx-poor lherzolite. 1mm; parallel polar.

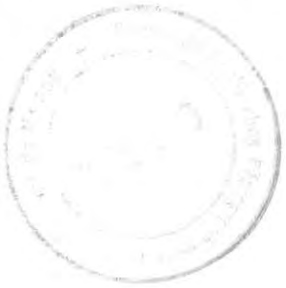
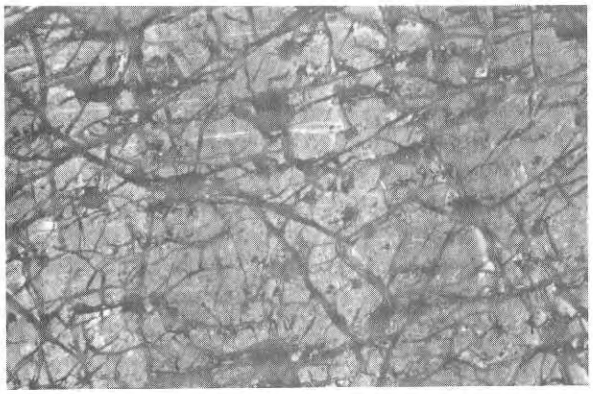
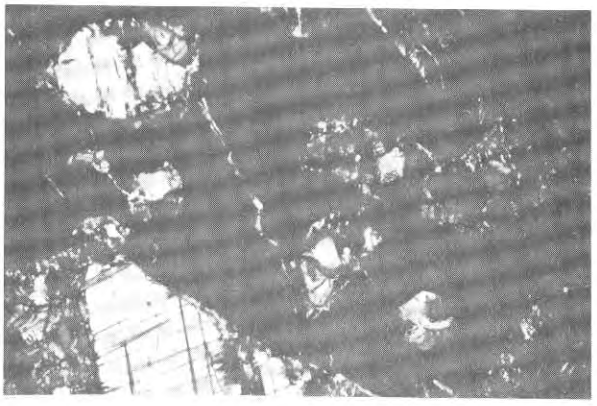


Fig.3 - Shear zone along which peridotite was deformed and contaminated. Deformed olivine gabbro has crystallized later from a tholeiitic magma.

Fig.4 - Early dyke showing the fractionation of Ol - Sp and Pl. Inclusions in Sp may contain pargasite, Opx and Cpx. 2,5; cross polar.





RÉSUMÉ - CONCLUSION

"C'est pourquoi nous avons maintenu entre notre entreprise et certains corps de doctrine, en particulier la "Tectonique globale", une distance, certes respectueuse. En effet, une construction soumise à un cadre théorique, si bien fondé qu'apparemment il soit, risque de vieillir très vite, et en tout cas beaucoup plus qu'une synthèse descriptive, hiérarchisée et critique".

P. ROUTHIER (1980)

"En fait, tout esprit véritablement scientifique procède, plus ou moins consciemment, par synthèses étagées, c'est-à-dire par rapprochements et généralisations de plus en plus larges. Face à cette notion les oppositions un peu trop classiques entre analyse et synthèse, entre induction et déduction, perdent quelque peu de leur valeur. A vrai dire, elles deviennent assez évanescentes pour qui a pratiqué la recherche ; ce sont de simples mouvements aller et retour de la pensée, plus ou moins embrassants, quasi simultanés et inconscients".

P. ROUTHIER (1969)

RÉSUMÉ-CONCLUSION

Les ophiolites, dont le terme a été défini depuis fort longtemps sont constituées d'un assemblage de péridotites mantellaires, cumulats et laves mis en place à différentes profondeurs. Elles présentent un caractère unitaire qui résulte de processus liés à la création des magmas ; une fusion laisserait en résidu les péridotites tandis que des liquides s'échappant vers la surface cristalliseraient les cumulats. Ce magmatisme peut avoir lieu dans des zones distensives issues de systèmes en extension ou en décrochement (mise en place primaire). Lorsqu'il cesse, ses témoins sont progressivement incorporés au continent lors de collision (mise en place secondaire). On obtient alors une ou plusieurs sutures dont certaines ne correspondent pas forcément à la patrie d'origine. La comparaison des sutures entre elles permet de mettre en évidence certaines discontinuités qui, si elles sont fondamentales, peuvent être liées à des sites géotectoniques particuliers. Soulignons, entre autre, les variations dans la composition des liquides qui, par analogie avec les équivalents effusifs actuels, indiqueraient globalement un environnement de ride médio-océanique ou d'arc insulaire. Dès lors, la genèse des ophiolites peut être extrêmement variée et un seul modèle pétrogénétique et structural apparaît comme insuffisant pour expliquer la nature des différents termes des assemblages ophiolitiques.

Notre travail, dont la formulation finale est présentée en quatre parties, a consisté à mettre en évidence les discontinuités géochimiques et structurales entre les assemblages, compte tenu aussi de leur situation géologique, et à expliquer, au sein de chacun d'entre eux, les mécanismes de fonctionnement, depuis la fusion jusqu'au fractionnement dans les chambres infra-crustales en passant par les systèmes de migration des magmas.

Dans la première partie, nous soulignons les discontinuités dont les plus évidentes sont géochimiques. Si deux grands types d'ophiolites peuvent être distingués en se basant, soit sur la composition des laves ou du résidu, soit sur la nature et l'ordre de cristallisation des cumulats, des subdivisions peuvent être apportées suivant que le magma est plus ou moins sursaturé en silice et en alumine ou riche en éléments incompatibles. En général, la diminution de ces derniers s'accompagne d'une sursaturation croissante en silice ; dans les cumulats

associés, le plagioclase devient de plus en plus tardif, et l'orthopyroxène précoce, alors que le manteau s'appauvrit progressivement. Ces caractères sont compatibles avec une origine à proximité d'une zone de subduction.

A l'opposé, les séries riches en éléments incompatibles, à plagioclase précoce et orthopyroxène tardif, se rapprochent de celles des rides médio-océaniques ou de bassins marginaux loin d'une zone de subduction. De la même façon, d'un point de vue structural, les ophiolites peuvent être classées en deux groupes. Dans le premier, les plans lithologiques et structuraux du manteau et des cumulats sont sub-horizontaux tandis que les filons bien réglés sont parallèles à l'axe de la ride. Dans le second groupe, le rubanement et la foliation des péridotites, fortement redressés, forment un angle quelquefois important avec les cumulats associés à des filons aux orientations diverses.

Les séquences ophiolitiques de ce dernier groupe varient latéralement. D'une manière générale, au sein d'une même suture, les discontinuités stratigraphiques observées parallèlement à celle-ci résultent de systèmes d'accrétion particuliers liés soit à des rides, soit à des transformantes : cas de la série de Rospigliani en Corse.

En dépit de ces divergences entre les sutures, les assemblages ophiolitiques et les terrains associés présentent quelques points communs qui constituent autant de contraintes quant à leur genèse(2ème partie). Une compilation des datations des membres de l'assemblage, des formations sédimentaires supra-ophiolitiques et des séries infra-péridotitiques impliquent une durée d'accrétion brève. Elle serait localement de 40 MA au maximum pour les ophiolites de type I (selon la classification pétrologique) et de 10 à 20 MA pour les autres. Cette constatation a conduit à envisager des mécanismes d'ouverture et de fermeture des bassins ophiolitiques autre que ceux résultant respectivement d'un système en expansion suivi d'une collision frontale. Dans cette optique, trois système d'ouverture et deux pour la fermeture ont été définis; ils peuvent s'appliquer à un grand nombre de sites géotectoniques. A côté de systèmes en extension couramment décrits, il peut exister des ouvertures obliques issues de la superposition du jeu

des rides et failles transverses et des ouvertures dans des bassins rhombiques parallèles à des décrochements. Enfin, si la mise en place secondaire des ophiolites peut être expliquée par une subduction puis une obduction lors d'une collision frontale, elle peut être aussi provoquée par une collision oblique liée au coulissage de plaques à bordures irrégulières. Il a été interprété que les ophiolites aux variations lithologiques sub-horizontales, en général largement charriées sur la plate-forme était issues de système en extension, parfois dans des bassins rhombiques, tandis que les autres, à structures verticalisées, proches de leur patrie d'origine, étaient plutôt compatibles avec un système en décrochement ou des ouvertures obliques. Soulignons au passage que les ophiolites les plus pauvres en TiO_2 ont souvent des structures verticalisées, qui seraient ainsi formées près des marges actives.

Dans la troisième partie, des séquences ophiolitiques caractéristiques des types pétrologiques I et II ont été étudiées d'un point de vue minéralogique et géochimique afin de rechercher l'origine des liquides observés en calculant, entre autre, la composition des magmas parentaux en équilibre avec le manteau. Les types Corse et Vourinos appartenant respectivement aux types I et II ont été analysés. En Corse, après une fusion (20 %) les magmas picritiques subissent un fractionnement de hautes pressions engageant orthopyroxène, spinelle, olivine et plus ou moins plagioclase avant de donner des cumulats de basses pressions (Ol-Pl-Cpx) et les liquides émis en surface. L'évolution de la chambre magmatique est compatible avec des cristallisations à l'équilibre qui favorisent la concentration des éléments incompatibles en fin de différenciation. Les derniers termes seraient en plus issus d'un processus d'immiscibilité. La cristallisation rapide des zircons, dans certaines roches acides pauvres en potassium, est responsable d'inclusions dont la composition témoigne du fractionnement du potassium et donc de son départ probable dans une phase vapeur. Dans le massif du Vourinos, deux séquences ont été dégagées. La première provient du fonctionnement de larges chambres où des cristallisations à l'équilibre ont eu lieu lors de remplissages successifs avant qu'une cristallisation fractionnée plus parfaite ne s'installe. Les liquides (basaltes à dacites) ne viendraient en surface qu'après 40 % de fractionnement (Ol+Cpx principalement) dans ces chambres alimentées par des basaltes parentaux magnésiens. La seconde série qui lui est postérieure, résul-

terait de magmas parentaux plus sursaturés en silice. Dans les chambres limitées, la cristallisation fractionnée plus parfaite conduirait rapidement à d'abondantes dacites et rhyolites. Ce type de série est souvent connu en fin d'évolution du magmatisme ophiolitique. Il indique que la quantité de magma arrivant vers la surface diminue tandis que la composition change graduellement en accord avec un appauvrissement continu du manteau.

Une étude des variations du comportement de Na et Cr dans les clinopyroxènes a permis de dégager les principaux processus magmatiques se déroulant dans les manteaux océaniques et continentaux à partir d'analyses tirées de massifs péridotitiques et de nodules. A côté de modifications liées aux fusions partielles et de rééquilibrations subsolidus lors de la diminution de pression et de température, d'autres tendances mettraient en évidence des fractionnements ou des processus métasomatiques. En particulier, les péridotites du Monte Maggiore, Corse témoigneraient d'épisodes de fusion partielle suivi de fractionnement et contamination par des liquides plus ou moins riches en alcalins et éléments incompatibles à différentes pressions. Par ailleurs, l'étude structurale et pétrologique entreprise dans ce massif a également permis de préciser le régime d'écoulement des magmas. Les premiers liquides enrobent les minéraux résiduels qu'ils séparent. En allant vers la surface, les contraintes déviatoriques qui augmentent, sont responsables d'ouverture dans des plans privilégiés parallèles à $\sigma 1$ ou en position conjuguée tandis que le volume disponible pour la circulation des magmas décroît. Les produits les plus récents sont aussi les plus déformés. La rotation plus ou moins prononcée des minéraux ayant poussé perpendiculairement aux fractures a permis de suivre les coulissages horizontaux et verticaux des blocs de péridotites.

Dans les ophiolites de type I le manteau originel serait une lherzolite, le résidu, une lherzolite appauvrie en Cpx ; dans celles du type II, une lherzolite appauvrie donnerait un résidu harzburgitique voire dunitique avec un taux de fusion croissant.

La fusion a lieu dans un diapir mantellaire effectuant une ascension adiabatique vers la surface (4ème partie). En fonction de la vitesse de remontée et du refroidissement, des fractionnements peuvent s'y dérouler comme en Corse.

La faible durée de l'accrétion ophiolitique laisse suggérer qu'au jeu distension-collision frontale, il est possible de lui substituer des coulissements qui peuvent intervenir dans n'importe quel environnement géotectonique. D'un autre côté, l'analyse des principaux assemblages ophiolitiques, la quantification des fractionnements a permis de juger de la validité de la classification des ophiolites en deux types pétrologiques principaux. La formation du premier groupe, voisin du magmatisme des dorsales médio-océaniques, résulte de l'établissement d'une zone d'accrétion qui engage un renouvellement plus ou moins continu du manteau sous-jacent. Ce mécanisme peut jouer dans les océans et les bassins marginaux en domaine continental ou océanique. En fonction de ces sites, l'allochtonie des massifs sera plus ou moins prononcée. Au contraire pour le second groupe, l'appauvrissement progressif du manteau peut être expliqué par son rôle passif non convectif. Comme l'apparition de ces ophiolites intervient tou - jours à un certain stade de maturité d'un bassin subsident rempli de produits sédimentaires et/ou volcano-sédimentaires, leur histoire est inéluctablement liée au développement d'une orogène. Celle-ci intervient en bordure de plaque passive puis active (un magmatisme calco-alcalin est quelquefois postérieur aux ophiolites) lors de leur coulissage en réponse à des convergences jouant en d'autres points du globe. La structure interne de ces ophiolites que l'on a pu classer en deux groupes dépend du degré d'extension. Des coulissages permettent d'expliquer la juxtaposition de ceintures ophiolitiques et formations associées d'âge décroissant vers la bordure de la marge continentale (cas de la Cordillère Nord Américaine) et l'ouverture de bassins décalée dans le temps et l'espace (cas de la Méditerranée orientale et de l'Indus).

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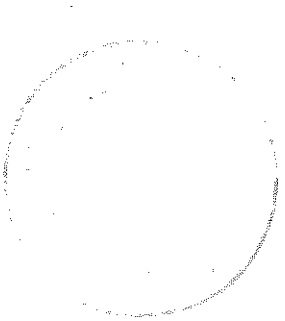

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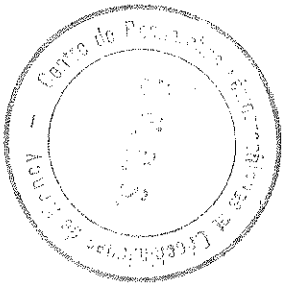

Résumé

Le caractère unitaire des ophiolites résulte de processus inéluctables à tout magmatisme qui engendre résidu, cumulat et liquide. Une comparaison des ceintures ophiolitiques entre elles a permis de souligner d'importantes discontinuités pétrologiques, structurales et stratigraphiques liées à des systèmes géotectoniques différents mais aussi à des variations longitudinales le long de la ceinture considérée.

La classification pétrologique des ophiolites en deux types proposée précédemment a été affinée tandis que la quantification des fractionnements (majeurs et traces) à partir des exemples Corse et Vourinos a permis de juger de sa validité. En particulier dans les ophiolites de type II (Vourinos) on retiendra que les liquides observés ont pu subir 40 % d'extraction réalisée principalement dans des chambres magmatiques sous-crustales. La composition des magmas qui évolue avec le temps -diminution des incompatibles et augmentation de la sursaturation en silice- est en accord avec un appauvrissement progressif du manteau.

Dans les ophiolites à affinité de ride médio-océanique, (type I), le manteau est renouvelé lors de l'accrétion ; à partir des hétérogénéïtés pétrologiques et structurales observées dans un massif de péridotites de Corse, l'importance des processus de fusion, fractionnement et métasomatisme dans un diapir mantellaire mis en place dans une zone d'accrétion a pu être appréciée et la chronologie des mécanismes précisée.

Les ophiolites de type II ont servi de base à une classification structurale des ophiolites en deux groupes dont la structure interne pourrait être liée aux conditions d'ouverture. En particulier, dans les ophiolites à structures verticalisées qui sont souvent plus pauvres en éléments incompatibles que celles à variations lithologiques et structurales sub-horizontales, la morphologie des chambres a été estimée à partir de l'étude d'un massif en Iran.

Dans toutes les ophiolites examinées, la durée de l'accrétion est brève dépassant rarement 40 Ma même pour celles qui montrent le plus d'affinité géochimique avec les échantillons océaniques. Ces différents aspects, de même que la nature des formations géologiques associées, ont conduit à substituer au jeu distension-collision frontale, des coulissements induits par la dérive des plaques majeures. Ces derniers sont responsables de la succession d'ouverture et fermeture plus ou moins obliques par rapport aux ceintures bordant les continents. Si certaines ophiolites correspondent à des tentatives d'océanisation, d'autres sont intégrées dans le développement d'un orogène. Dans ce cas, les ophiolites apparaissent après la formation de bassins à marges passives devenant quelquefois actives avant d'être ensemble rapidement tectonisées.

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