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Modélisation de la source des séismes par inversion des données sismologiques et géodésiques : application aux séismes du Nord de l'Algérie

Hamoud Beldjoudi

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Unité de recherche : **Laboratoire Géoazur**

Thèse de doctorat

Présentée en vue de l'obtention du grade de

Docteur en Sciences de la Planète et de l'Univers

de

UNIVERSITE COTE D'AZUR

Par

Hamoud Beldjoudi

Modélisation de la source des séismes par inversion des données sismologiques et géodésiques : application aux séismes du Nord de l'Algérie

Dirigée par **Bertrand Delouis**

Soutenue le 11 juillet 2017

Devant le jury composé de :

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Résumé :

La caractérisation de la source d'un séisme se fait à partir de l'analyse des mesures des déplacements transitoires et statiques du sol, et dépend de la quantité et de la qualité de ces mesures. Nous avons travaillé sur la détermination des mécanismes au foyer des séismes modérés de Tadjena (Mw 5.0, 2006), Béni-Ilmane (Mw 5.5, 2010), Hammam Melouane (Mw 5.0, 2013), Bordj-Ménaïel (Mw 4.1, 2014), Alger (Mw 5.7, 2014) et M'ziraa (Mw 5.1, 2016). Pour cela, nous avons inversés les formes d'ondes en champ proche et régional des stations large bandes et accélérométriques du réseau sismologique algérien (ADSN). Nous avons déterminé la distribution spatio-temporelle du glissement cosismique du séisme de Boumerdes-Zemmouri (Mw 6.8, 2003) en inversant conjointement l'ensemble des données disponibles, sismologiques (télésismiques, accélérometriques) et géodésiques (GPS, InSAR, soulèvement côtier). Nous avons travaillé sur la relation qui peut exister entre le séisme de Boumerdes-Zemmouri et les séismes de Hammam Melouane, Bordj Ménaïel et Alger (Mw 5.7, 2014) en termes de transfert de contrainte (CFF). Nous avons calculé le champ de contrainte dans différentes régions de l'Algérie par inversion des mécanismes aux foyers disponibles dans chaque région.

Mots Clés : sismologie, séismes, source, contrainte, inversion, mécanisme au foyer, accéléromètres, large-bande, GPS, InSAR, soulèvement côtier, CFF, Algérie, Boumerdes-Zemmouri

Abstract:

Studies of the earthquake source are based on observations of seismic and static ground motions, which depend on the quality and the density of measurements. In this thesis, we present studies of the determination of the focal mechanism of the Tadjena (Mw 5.0, 2006), Béni-Ilmane (Mw 5.5, 2010), Hammam Melouane (Mw 5.0, 2013), Bordj-Ménaïel (Mw 4.1, 2014), Alger (Mw 5.7, 2014) and M'ziraa (Mw 5.1, 2016) earthquakes, by modeling waveforms in the near and regional field with broadband and strong motion stations of the Algerian Digital Seismic Netwok (ADSN). In addition, we determined the coseismic slip distribution of the Boumerdes-Zemmouri earthquake (Mw 6.8, 2003) by inverting a most comprehensive set of data (teleseismic, strong motion, GPS, InSAR, coastal uplift). We calculated the Coulomb Failure Function between the Boumerdes-Zemmouri earthquake (source fault) and the Hammam Melouane, Bordj Ménaïel and Algiers (Mw 5.7, 2014) events (receiver faults). We computed the stress tensor in different regions of Algeria obtained from the inversion of the available focal mechanisms.

Keywords: seismology, earthquakes, source, stress tensor, inversion, focal mechanism, strong motion, broadband, GPS, InSAR, coastal uplift, CFF, Algeria, Boumerdes-Zemmouri

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Introduction

Les objectifs de la thèse concernent l'amélioration de la connaissance de la sismicité du nord de l'Algérie,

Une des motivations a été de tirer parti du développement du réseau large bande et accélérométrique numérique du CRAAG depuis 2007. Depuis que ce réseau fonctionne, des séismes modérés se sont produits que nous avons analysés avec une approche de modélisation des formes d'ondes à distance locale à régionale, qui n'avait pas été employée auparavant.

Bien que le séisme de Boumerdes ait fait l'objet de nombreuses études, il reste des aspects de la source qui n'ont pas été complètement abordés. Les travaux antérieurs ne sont basés que sur une partie de ces données disponibles. Nous proposons la construction d'un modèle de rupture cosismique tenant compte de l'ensemble des données disponibles, sismologiques (télésismiques et accélérométriques) et géodésiques (GPS, InSAR, et soulèvement côtier). Par ailleurs, certaines complexités concernant les structures activées par le choc principal n'ont jamais été testées dans les modèles cosismiques, notamment les escarpements et les failles en pallier- rampe décrits par Deverchère et al. (2005) au large de la région épicentrale.

On sait que les perturbations de contraintes produites par un choc principal sont fortement influencées par la distribution du glissement cosismique. Il est donc intéressant de préciser cela en utilisant le modèle de rupture que nous obtenons pour le séisme de Boumerdes. Nous choisirons quelques failles cibles associées à des séismes récents et localisées au voisinage du séisme de Boumerdes, afin d'y calculer les changements de contraintes de Coulomb. Ceci afin de déterminer si ces séismes modérés ont pu être déclenchés par le séisme de Boumerdes.

Nous profitons également de la disponibilité de nouveaux mécanismes au foyer pour effectuer des redéterminations du tenseur des contraintes au niveau de différents secteurs du nord de l'Algérie, en prenant en compte également les mécanismes plus anciens.

L'ensemble de ces travaux vont s'organiser en cinq chapitres, dont voici un bref descriptif.

Chapitre 1 : consacré à l'état des connaissances sur le contexte sismotectonique du nord de l'Algérie. Nous décrivons la sismicité dans le Nord algérien, en relation avec le développement des réseaux de surveillance sismologique en Algérie, avec des cartes de sismicité relatives à chaque période, historique et instrumentale. Nous présentons également brièvement les principaux séismes qui ont secoué le Nord algérien et qui ont fait l'objet d'études spécifiques.

Chapitre 2 : nous présentons les notions générales et théoriques concernant l'approche par inversion des formes d'ondes pour le mécanisme au foyer dans une première partie, puis les notions concernant l'inversion multi-données pour la distribution du glissement cosismique.

Chapitre 3 : nous présentons les résultats de l'approche en point source reliant les sismogrammes aux mécanismes aux foyers en analysant cinq séismes modérés récents en Algérie**,** deux de ces séismes ayant fait l'objet de publications dont je suis le premier auteur et qui sont incluses intégralement dans le chapitre: le séisme de Tadjena, dans la région du Chélif (Ouest algérien) survenu en 2006 (Beldjoudi et al., 2011), et la séquence sismique de Béni-Ilmane dans la région des Biban (Est algérien) survenue en 2010 (Beldjoudi et al., 2016). Les résultats que j'ai obtenus pour l'inversion de la source d'autres séismes sont également présentés dans ce chapitre : le séisme de Bordj Ménaïel, localisé au SE de la faille de Boumerdes, dont l'article complet de Semmane et al. (2015) est présentéen annexe 1 ; le séisme de Hammam Melouane, survenu sur la bordure sud de la Mitidja en 2013, dont l'article complet de Yelles-Chaouche et al. (2017) constitue l'annexe 2. Un dernier séisme, localisé dans le Sud algérien (transition Atlas saharien-plateforme) est présenté aussi dans ce chapitre.

Chapitre 4 : nous présentons l'inversion du glissement du séisme de Boumerdes en intégrant un jeu de données sismologiques et géodésiques complet, un nouveau modèle de faille est proposé. Cette partie est présentée sous forme d'un article en préparation. Nous avons ensuite essayé de comprendre dans la partie qui suit, qu'elle a été l'influence du séisme de Boumerdes sur la sismicité de la région en termes de perturbation de la fonction de rupture de

Coulomb (Coulomb Failure Function, ΔCFF) où nous avons ciblé quatre séismes modérés, survenus entre 2013 et 2017.

Chapitre 5 : nous présentons tous les résultats de l'inversion du tenseur des contraintes en Algérie que j'ai entrepris depuis 2009, avec un premier tenseur calculé dans la région des Babors (Est algérien) à la suite du séisme de Lâalam dans la localité de Kherrata (Beldjoudi et al. (2009). Un autre tenseur est calculé dans la région du Chélif, à la suite du séisme de Tadjena (Beldjoudi et al., 2011). Un troisième tenseur est calculé dans la région des Biban à la suite de la séquence sismique de Lâalam (Beldjoudi et al. 2016). Un quatrième et cinquième tenseurs sont calculés respectivement dans la région de Mila (Est algérien) intégré dans l'article de Semmane et al. (2012) sur la sismicité induite et dans la région de Boumerdes où le papier vient d'être accepté (Kherroubi et al., 2017).

Une conclusion avec quelques perspectives vient clore mes travaux de thèse

Trois papiers en co-auteur sont présentés en annexe, le premier sur le séisme de Bordj Ménaïel de 2014 (Annexe 1), le second sur la crise séismique de Hammam Mélouane de 2013 (Annexe 2), le troisième sur la sismicité induite dans la région de Mila en 2007 (Annexe 3).

Chapitre 1 : État des connaissances sur le contexte sismotectonique du Nord de l'Algérie

Introduction

L'Algérie occupe la partie nord du continent Africain. Cette partie nord de l'Algérie, représente le front de collision entre les plaques Afrique et Eurasie. La sismicité de l'Algérie est localisée sur cette partie nord formée par quatre domaines morpho-structuraux: l'Atlas Tellien et partie off-shore, les hauts plateaux, l'Atlas Saharien et la partie nord de la plateforme saharienne (**Figure 1-1**). La fréquence et la magnitude de la sismicité est importante sur l'Atlas Tellien. Cette sismicité s'atténue si on se dirige vers le sud. Des séismes modérés sont enregistrés dans l'Atlas Saharien et la partie nord de la plateforme saharienne. Le catalogue de la sismicité disponible au niveau du Centre de Recherche en Astronomie Astrophysique et Géophysique (CRAAG) a répertorié un premier séisme qui s'est produit le 02 janvier 1365 à Alger sans donner plus de détail.

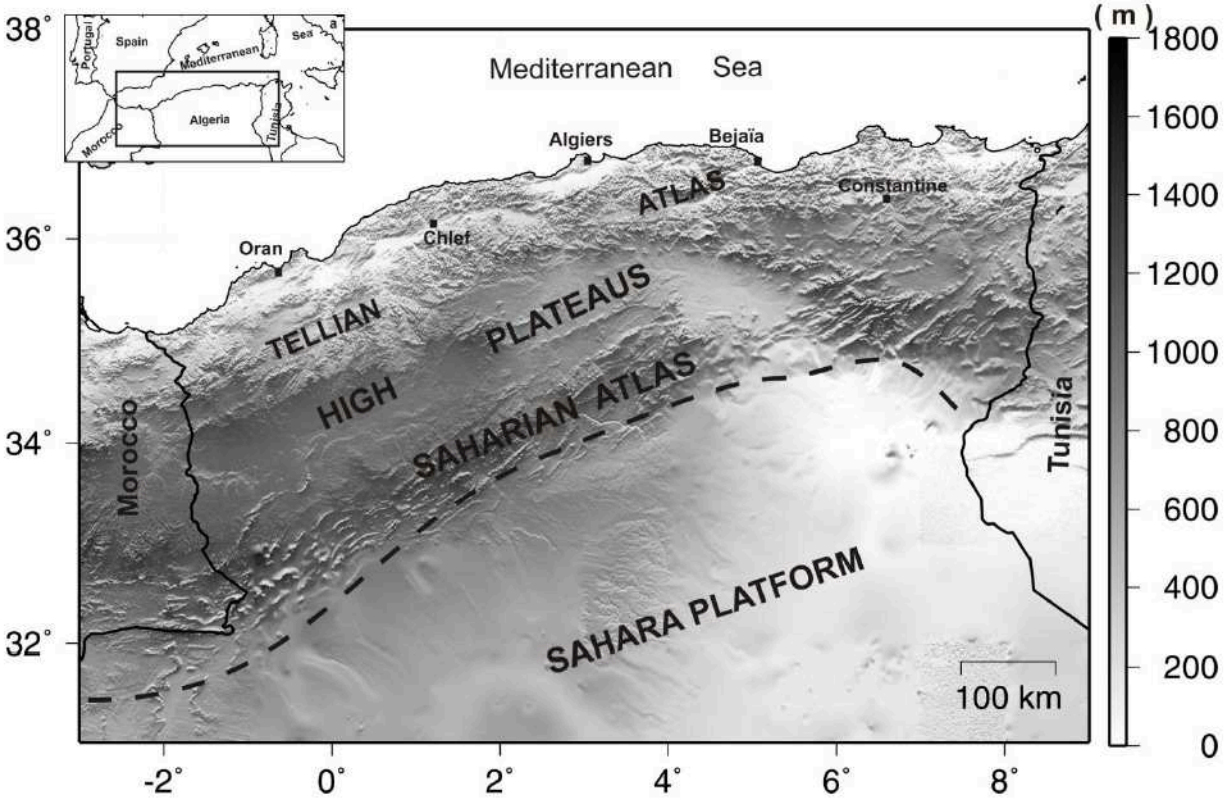

Figure 1-1 : *Domaines morpho-structuraux où se localise la sismicité algérienne : l'Atlas tellien, les hauts plateaux, l'Atlas saharien, la partie extrême nord de la plateforme saharienne.*

1.1 Contexte Cinématique

Les vitesses de convergence entre les plaques Afrique et Eurasie ont été calculées principalement par des mesures GPS (Nocquet et Calais, 2004 ; Serpelloni et al., 2007) et des modèles géologiques et géophysiques (Argus et Gordon, 1991 : *Nuvel-1* ; De mets et al., 1994 : Nuvel-1A).

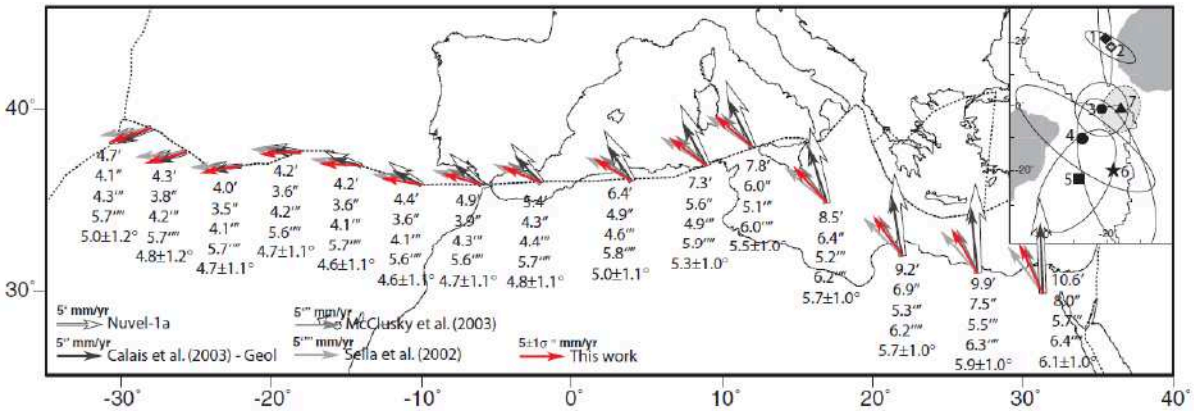

Figure 1-2 : *carte de synthèse des vitesses de convergence aux frontières de plaques tectoniques Afrique-Eurasie obtenues par compilation de modèles géologiques et géodésiques (d'après Serpelloni et al., 2007).*

Les résultats de mesures GPS conduites entre 1991 et 2004 (Sella et al., 2000 ; Mc Clusky et al., 2003 ; Serpelloni et al., 2007) sont présentés sur la **Figure 1-2**. Les vecteurs de déplacement montrent une direction de rapprochement NNW-SSE à NW-SE avec des vitesses de convergences qui diminuent d'Est en Ouest. A la longitude d'Annaba (7.30°) il a été mesuré, 5.3 mm/an, à la longitude d'Alger (~3°) une vitesse de 5.0 mm/an et à l'ouest à la longitude (-1°), une vitesse de 4.8 mm/an.

1.2. Sismicité du nord de l'Algérie

La sismicité historique de l'Algérie est décrite par plusieurs catalogues sismiques selon deux principales périodes : (1) la période de la colonisation française de l'Algérie avec les travaux de Rothé (1950) qui a répertorié un premier séisme qui s'est produit le 3 février 1716 à Alger (**Figure1-3**), avec des dégâts considérables, où *beaucoup de maisons ont été écroulées et un grand nombre d'habitants périrent écrasés.* Une intensité entre IX-X lui a été attribuée. Une forte réplique presque aussi forte que le choc principal a été enregistrée le 6 février de la même année. (2) La période post-indépendance avec les travaux de plusieurs auteurs

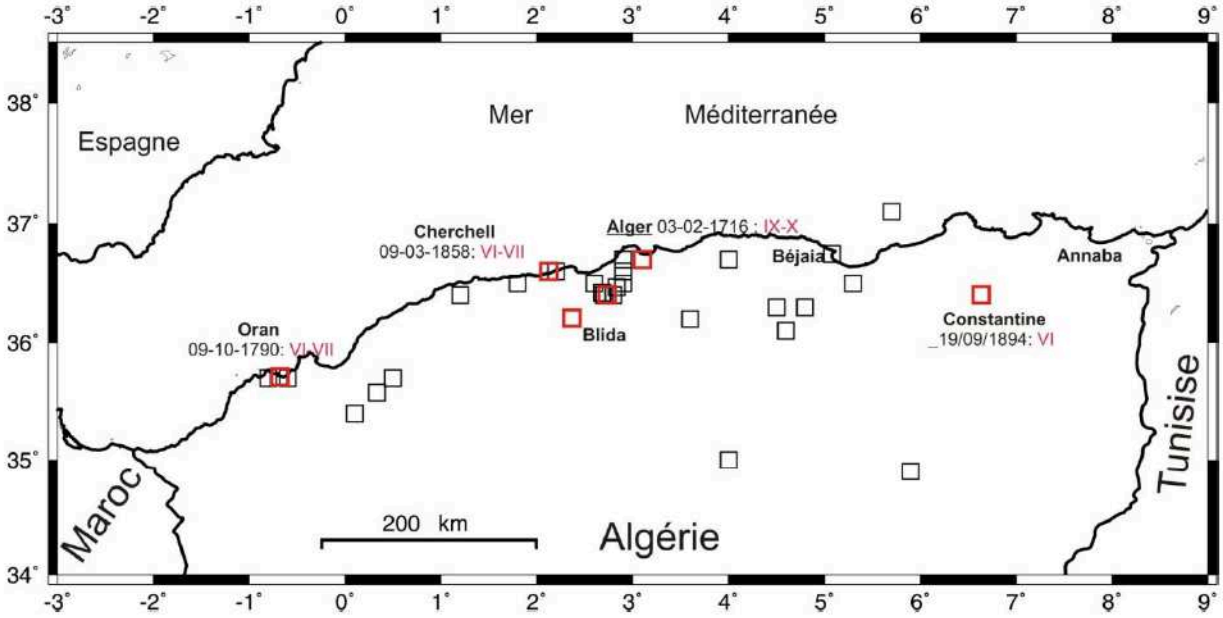

Figure 1-3 : *Carte des séismes historiques en Algérie entre 1365 et 1900. La localisation est donnée par rapport au nom de la ville où le maximum de dégâts a été observé.*

(Benhalou, 1985, Meziane et al. 1994) qui ont construit un catalogue de la sismicité de l'Algérie en reprenant et en explorant les précédents travaux. Le catalogue de Meziane et al., (1994), couvre la période 1365-1992 avec un premier séisme dans la région d'Alger le 02 janvier 1365. Harbi et Maouche (2009) ont revu la sismicité historique de l'Est de l'Algérie et ont publié un catalogue de la sismicité historique de la région.

Les premiers instruments de mesure des séismes (sismomètres) ont été inventés à la fin du 19e siècle, mais en Algérie la première station sismologique a été installée en 1910 au site Alger-Bouzaréah (Mokrane et al., 1994). La deuxième station a été installée en 1935 sur le barrage d'Oued Fodda dans la région du Chélif (région d'El Asnam, où s'est produit le séisme du 10/10/1980) et une dernière station a été installée dans la région de Sétif (région Est) en 1958 (**Figure 1-4**), et de fait nous commençons à voir les traits de la sismicité de l'Algérie au fur et à mesure que ces réseaux se développaient et se densifiaient. Une carte de sismicité compilée de 1365 à 1979 (**Figure 1-5**) montre que tout le nord de l'Algérie est concerné par cet aléa naturel à savoir la partie off-shore, le Tell Atlas, les hauts plateaux ainsi que l'Atlas saharien.

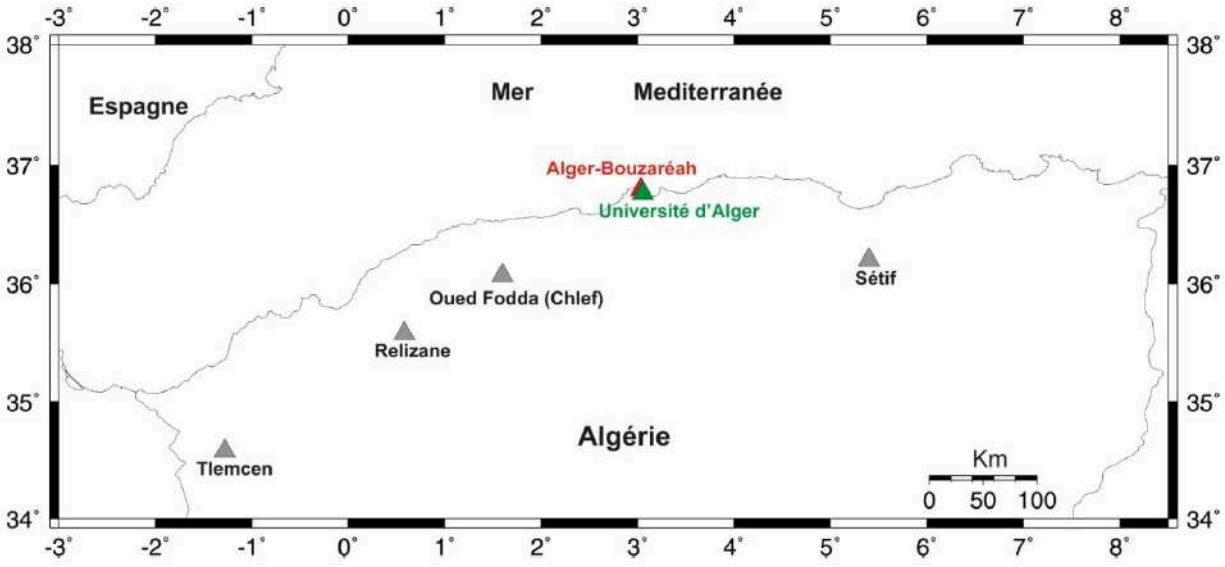

Figure 1-4 : *Réseau sismologique installé entre 1910 (station : Bouzaréah-Alger) et 1958 (station: Sétif).*

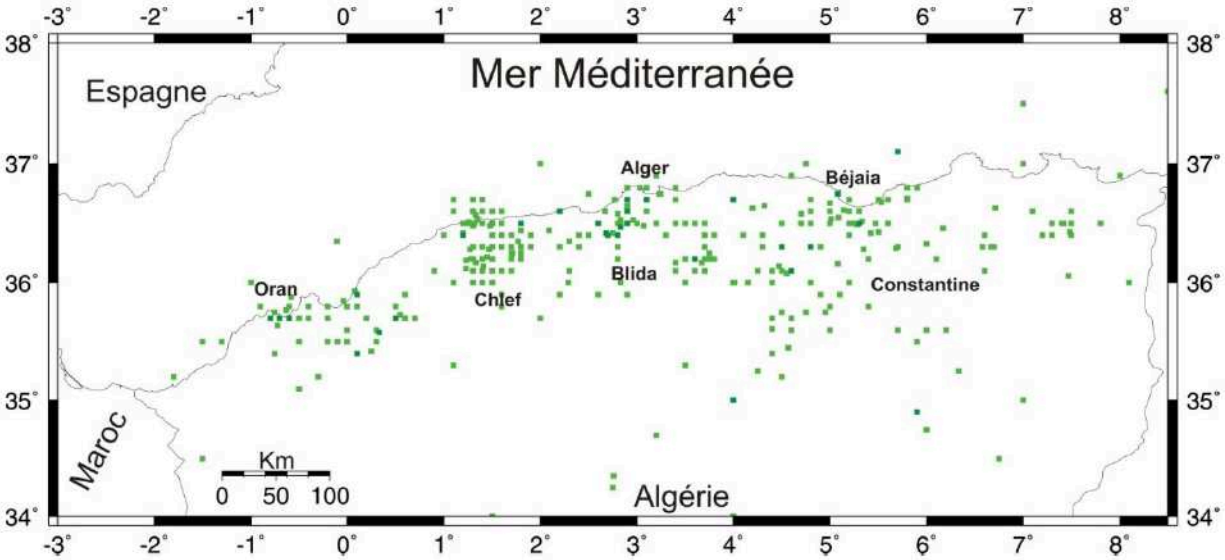

Figure 1-5 : *Carte de la sismicité de l'Algérie de 1365-1979 où épicentres macrosismiques et instrumentaux sont localisés.*

Suite au séisme d'El Asnam (Ms=7.3) du 10 octobre 1980, le réseau de surveillance sismologique qui existait déjà en Algérie a été densifié et on a vu le nombre de stations augmenter jusqu'à 32 stations courte période avec une transmission de la donnée par ondes radio réparties suivant quatre régionales : Alger, El Asnam, Oran et Constantine (**Figure 1-6**).

Chaque régionale était constituée d'une station courte période et d'une réception hertzienne pour renvoi sur Alger. Ce réseau est devenu opérationnel depuis le mois de mai 1990 à savoir dix ans après le séisme d'El-Asnam (1980). La sismicité enregistrée durant cette période est bien sûr localisée sur la partie Nord et confirme les traits de la sismicité déjà localisée en Algérie (**Figure 1-7**).

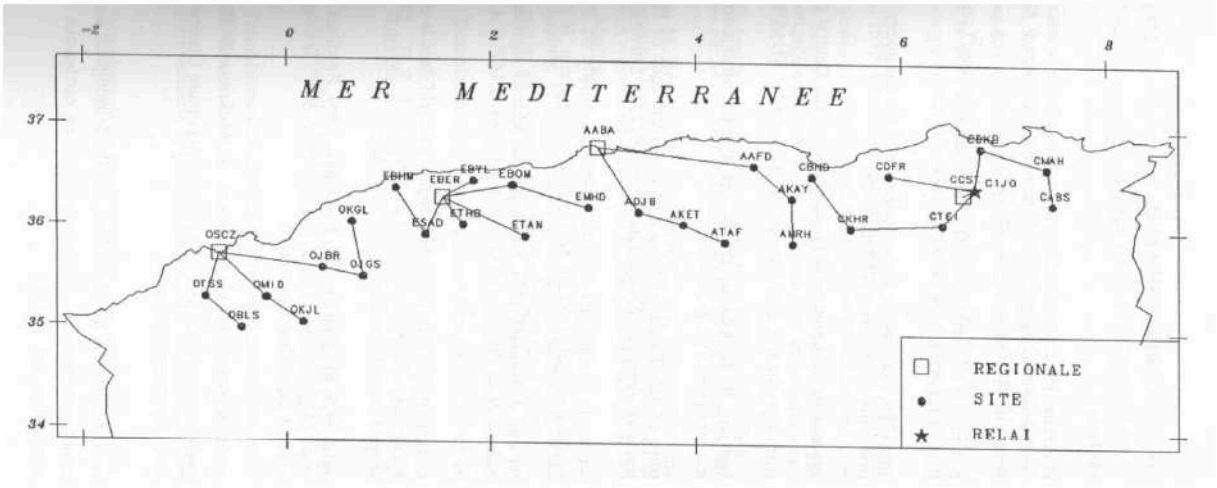

Figure 1-6 : *Réseau de surveillance sismologique télémétré Algérien (RSTA), installé depuis mai 1990 (Mokrane et al., 1994).*

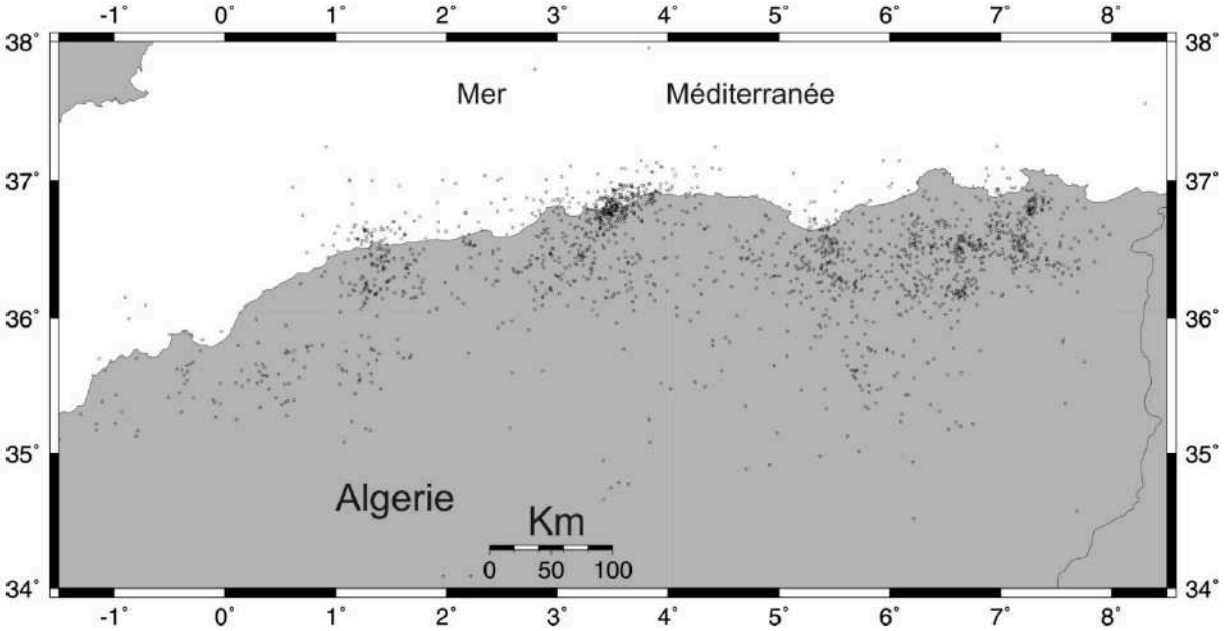

Figure 1-7 : *Carte de la sismicité de l'Algérie localisée entre 1992 et 2006 avec le premier réseau courte période télémétré en Algérie mis en service depuis mai 1990.Ce réseau a bien contribué à la localisation des séismes Algérie et même méditerranéens.*

Une partie du réseau a été détruite suite aux événements qu'a connus l'Algérie entre 1992 et 1999. Depuis 1999 le CRAAG a commencé la réhabilitation des sites endommagés et détruits. Le séisme de Boumerdes du 21 mai 2003 a été l'événement qui a poussé les pouvoirs publics à donner plus de moyens pour l'acquisition du matériel sismologique. Et en décembre 2006, un premier séisme a été localisé et un premier papier publié par Beldjoudi et al. (2011) grâce au réseau sismologique numérique Algérien (ADSN : Algerian Digital Seismic Network) nouvellement installé composé pour un premier temps par une dizaine de stations large bandes de type GEODEVICE (**Figure 1-8**).

La sismicité de l'Algérie est caractérisée par des séismes faibles à modérés, parfois forts. Le séisme d'El-Asnam du 10 octobre 1980 (Ms=7.3) reste jusqu'à présent le plus fort séisme enregistré dans la Méditerranée occidentale. Cette sismicité est concentrée dans la partie nord de l'Algérie qui représente le front de collision de la plaque Afrique avec l'Eurasie (**Figure 1-9**). Elle est présente d'Est en Ouest, sur le Tell Atlas et la partie off-shore, les hauts plateaux ainsi que la plateforme saharienne (**Figure 1-9**). Cette sismicité est superficielle, elle est localisée dans les 20 premiers kilomètres de la croûte terrestre. Le régime de contrainte est compressif orienté NNW-SSE à NW-SE, et entraine des structures en compression orientées NE-SW à EW ainsi que des structures en décrochement (Ouyed et al., 1983 ; Meghraoui, 1988 & 1996 ; Bounif et al., 2004, Delouis et al., 2004 ; Yelles-Chaouche et al., 2004, Beldjoudi et al., 2009, 2011 et 2016).

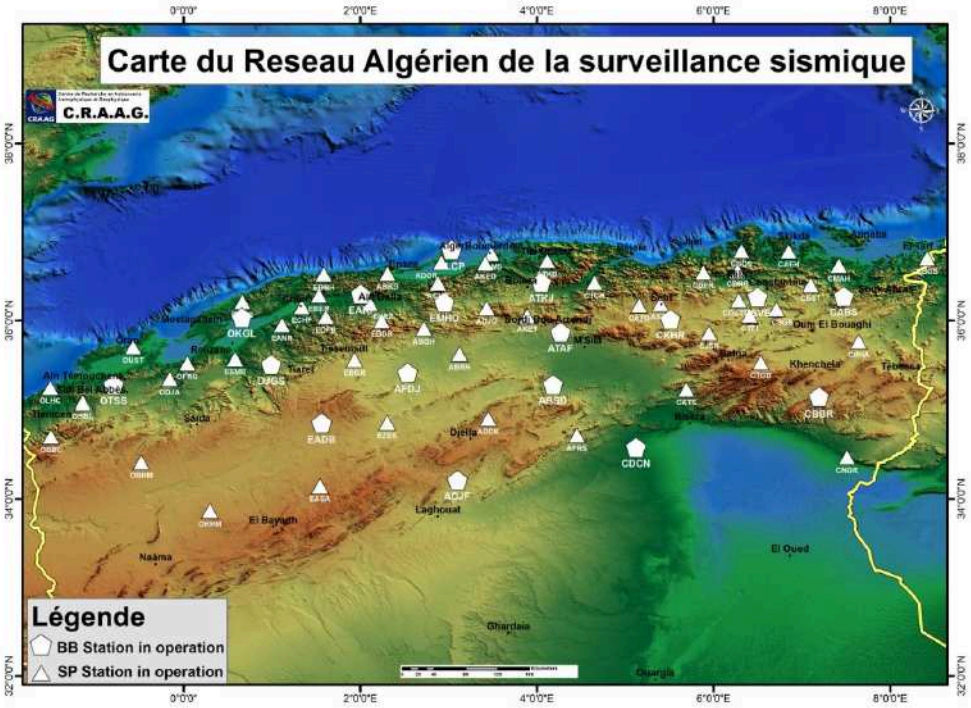

Figure 1-8 : *Carte du réseau de surveillance sismologique (ADSN : Algerian Digital Seismic Network) opérationnel depuis janvier 2007.*

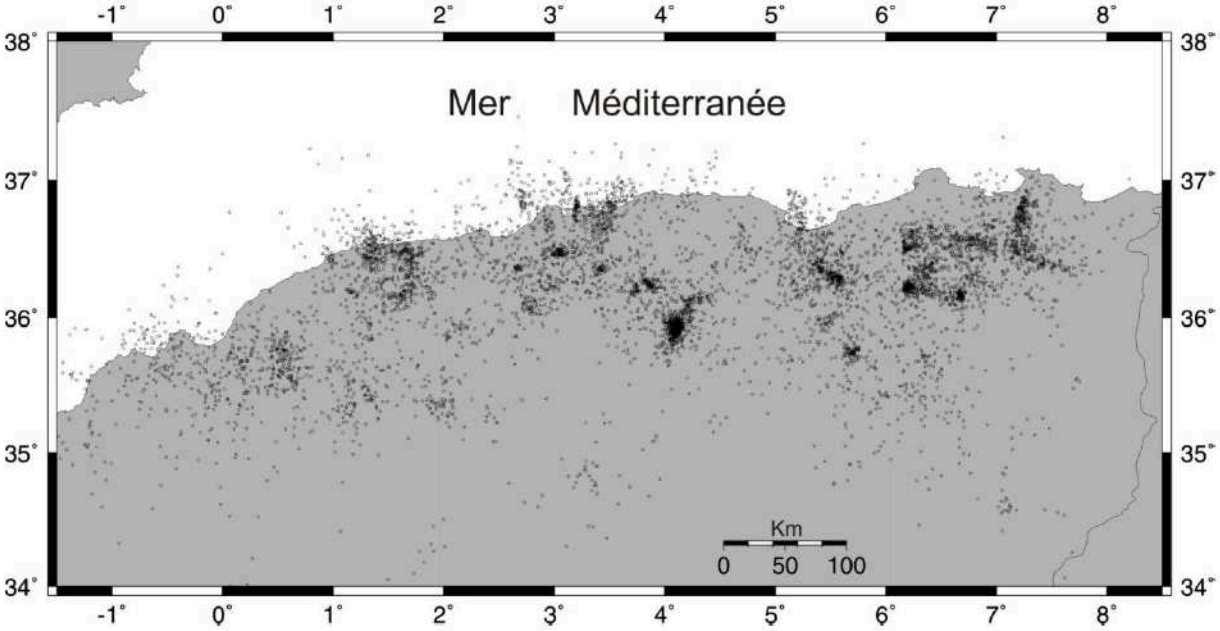

Figure 1-9 : *Carte de la sismicité du Nord d'Algérie (2007-2016) compilée depuis le catalogue du CRAAG. Tous les séismes sont localisés grâce au réseau ADSN.*

1.3. Les principaux séismes instrumentaux depuis 1980

1.3.1. Le séisme d'El Asnam du 10 octobre 1980

Ce séisme superficiel (36.16° N, 1.39° E, profondeur = 10 km) reste le plus violent survenu dans l'Atlas Tellien occidental (Ms=7.3) depuis l'ère instrumentale (**Figure 1-10**). Il a été étudié par de nombreux auteurs (Ouyed et al., 1981, 1981 & 1983 ; Deschamps et al., 1982 ; Ruegg et al., 1982 ; Yielding et al., 1989 ; Mokrane et al., 1994 ; Lammali et al., 1997 ; Meghraoui, 1988 & 1996 ; Beldjoudi, 2003). La faille (d'El Asnam ou de Oued Fodda) est située sur le pli-faille du bassin de Cheliff. La source sismique est de type inverse orientée NE-SW et plonge vers le NW (strike = 255, dip = 54, rake = 83). Ce séisme a laissé des traces de surface visibles sur 40 km le long de la faille avec un rejet vertical maximum observé de 6 mètres (**Figure 1-10**). Il a été ressenti sur un rayon de 250 km et son intensité maximale Io=IX (échelle MSK). Des milliers de pertes en vies humaines (2633 morts) ont été enregistrées et la ville d'El Asnam et ses villages limitrophes ont été détruits.

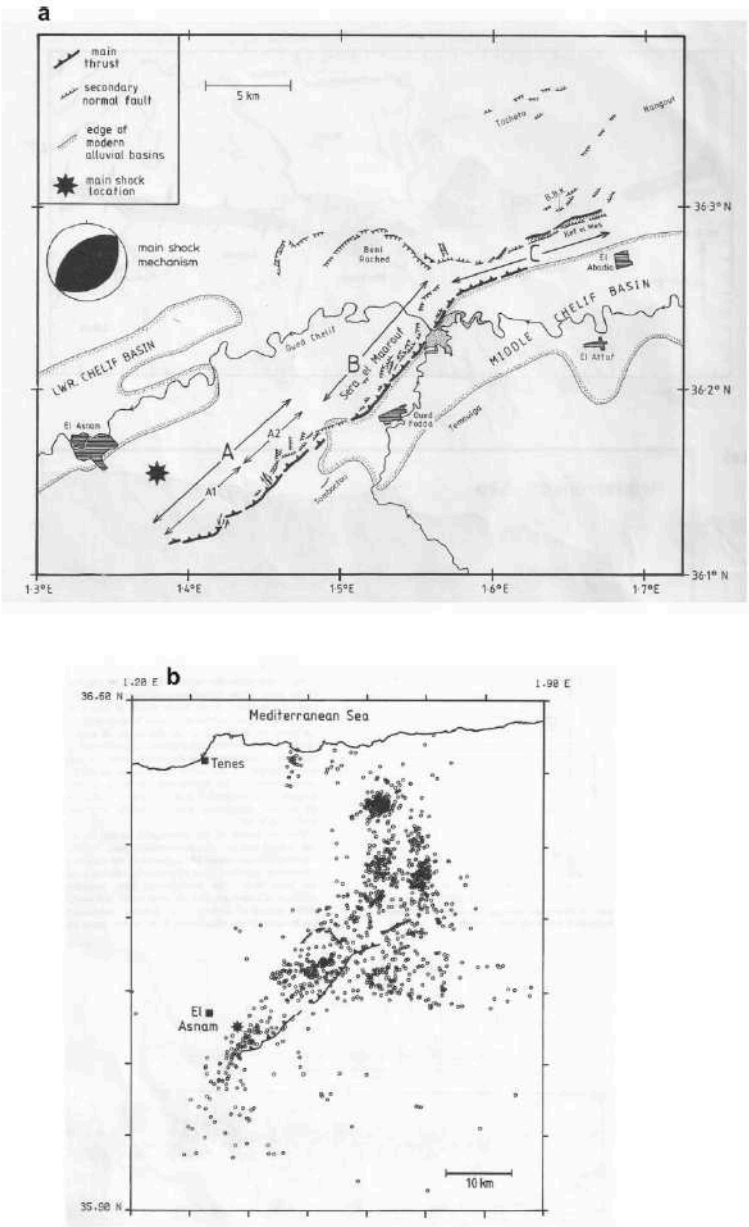

Figure 1-10 : *a) Carte montrant le mécanisme au foyer du séisme d'El Asnam du 10/10/1980 ainsi que sa localisation épicentrale (l'étoile). A, B et C sont les segments de la faille. b) La répartition des 1279 répliques bien localisées. Les traces de surfaces sont mentionnées par des lignes (Yielding et al., 1989).*

1.3.2. Le séisme de Constantine du 27 octobre 1985

Cet événement s'est produit dans l'Atlas tellien oriental, sa magnitude était Ms=6.0. L'intensité maximale observée a été estimée à VIII (échelle MSK). Ce séisme s'est produit dans la région de Constantine (36.46°N, 6.76° E, profondeur = 10 Km). Le mécanisme au foyer montre une faille verticale de mouvement senestre orientée NE-SW (strike = 217, dip = 84, rake = 19).

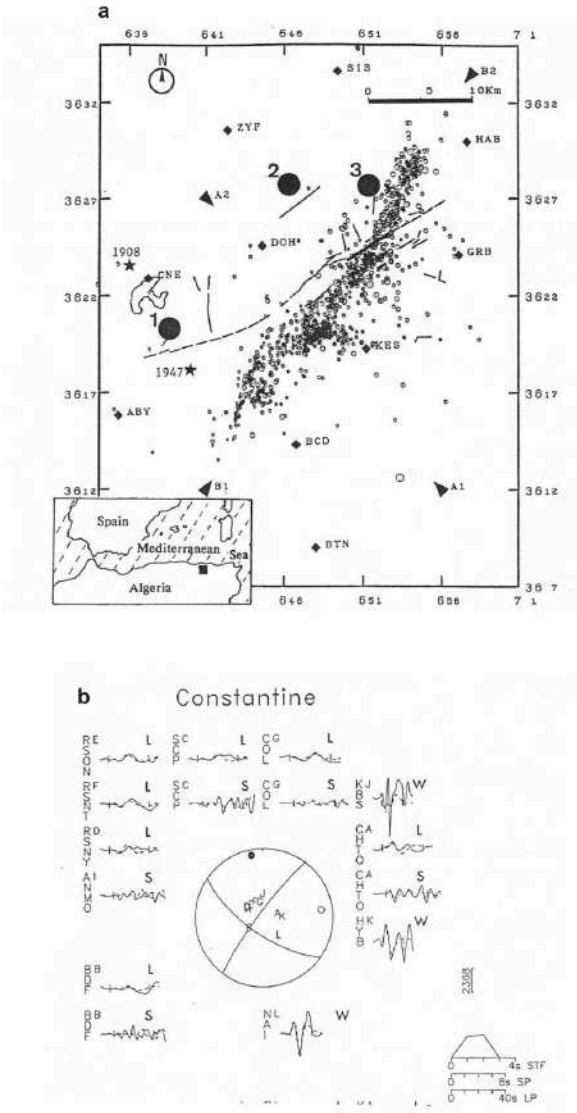

Figure 1-11 : *a) distribution des répliques du séisme du 27/10/1985. Les deux étoiles noires indiquent les localisations des séismes de 1908 et 1947. Les grands cercles noirs indiquent les différentes localisations obtenues par différents centre sismologiques, (1 : CSEM ; 2 : NEIC ; 3 : CMT Harvard). Les traits montrent la faille de Ain Smara et une partie de la faille active .b) Mécanisme au foyer du séisme de Constantine obtenu par modélisation des formes d'onde à une distance télésismique (Bounif et al., 2003) .*

Le mouvement vertical observé ne dépasse pas les 4 cm (Bounif et al., 2003). Uniquement dans les trois semaines suivantes, 1500 répliques ont été enregistrées (Bounif et al., 1987 ; Deschamp et al., 1991 ; Sebaï, 1997 ; Bounif & Dorbath, 1998) (**Figure 1-11**). La source sismique a été obtenue par la modélisation des formes des ondes P enregistrées par des stations situées à des distances télésismiques.

1.3.3 Le séisme de Tipaza (Mont Chenoua) du 29 octobre 1989

L'Atlas Tellien centre est secoué par un séisme de magnitude Mw = 6.0. Cet événement s'est produit à 70 km à l'ouest de la capitale Alger. L'épicentre a été localisé en mer au nord du Mont-Chenoua (Lat. = 36.66° ; Long. = 2.48° ; profondeur = 10 km) par les organismes internationaux, le CSEM (Lat. = 36,79° ;Lon. = 2.45°) et NEIC (Lat. = 36,73° ;Lon. = 2.44°). Il a été localisé par le CRAAG, dans le Mont-Chenoua. Les répliques enregistrées et localisées par Bounif et al. (2003) indiquent une direction NE-SW de la faille compatible avec le résultat de l'inversion des données large bande télésismiques qui indiquaient une source sismique inverse avec une direction NE-SW (**Figure 1-12**). Des coupes perpendiculaires à la direction d'allongement principale de la distribution des répliques ont montré un plongement du plan de la faille vers le NW (strike = 246 ; dip = 56 ; rake = 86). A Alger le séisme a été bien ressenti et la population a été prise de panique et plusieurs familles passèrent la nuit dehors (Sebaï, 1997 ; Bounif et al., 2003).

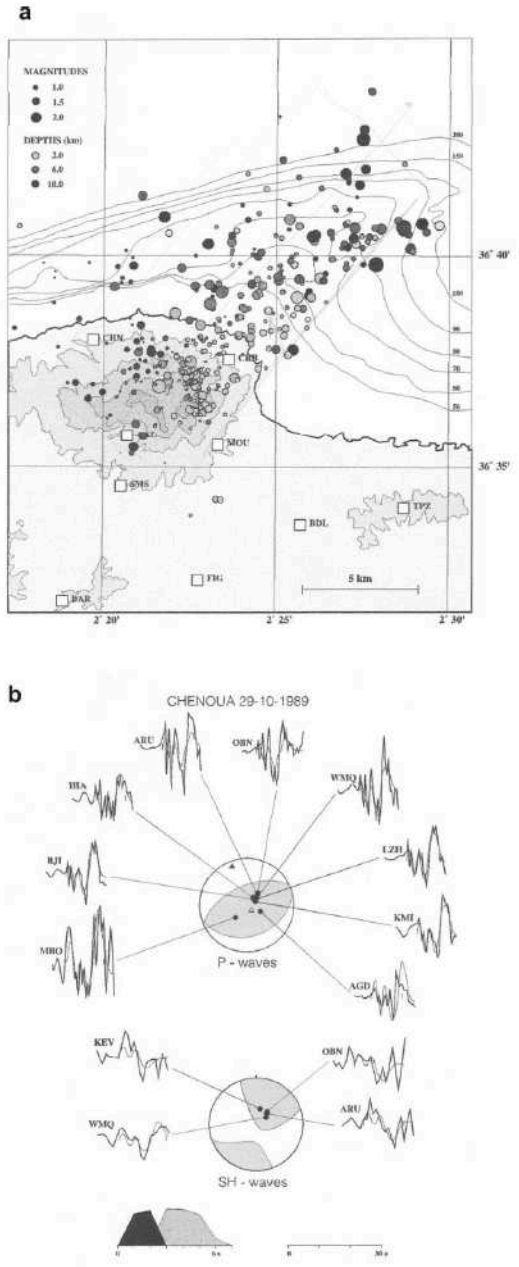

Figure 1-12 : *a) localisation des répliques du séisme de Mont-Chenoua du 29/10/1989 (d'après Bounif et al. 2003). b) Source sismique obtenue par la modélisation des ondes P et des ondes SH. (Haut) : mécanisme au foyer obtenu par inversion des ondes P à des distances télésismiques. (Bas): Fonction Temporelle de la Source. (d'après Deschamps et al., 1991).*

1.3.4. Le séisme d'Ain Temouchent du 22 décembre 1999

Ce séisme de magnitude (Mw = 5.7) s'est produit à Ain Temouchent située 70 km à l'ouest de la ville d'Oran. C'est un séisme superficiel localisé dans les 10 premiers km de la croûte terrestre. L'intensité maximale a été estimée à VII (Echelle MKS). Le séisme a provoqué de

sérieux dommages dans la ville d'Ain Temouchent et ses villages limitrophes (**Figure 1-13**). Le mécanisme au foyer (**Figure 1-14**) de ce séisme présenté par Yelles-Chaouche et al. (2004) montre une faille inverse orientée NE-SW. Belabbès et al., (2008) en utilisant les données d'interférométrie radar (InSAR) ont confirmé la direction NE-SW de la faille et ont pu montrer que la faille plonge vers le NW.

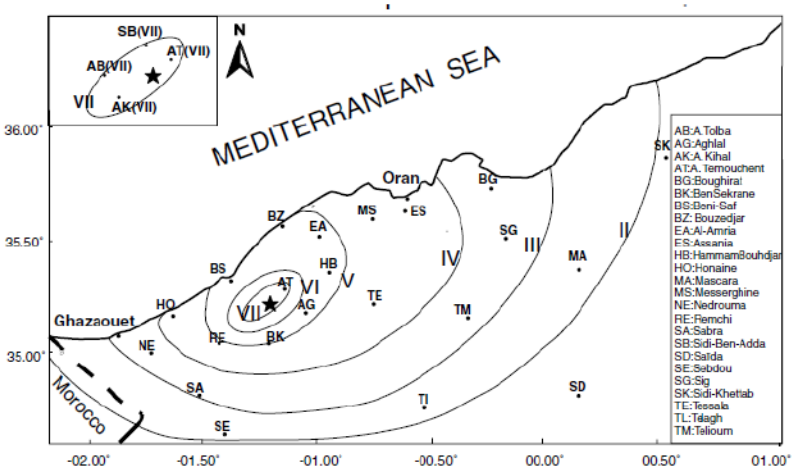

Figure 1-13 : *Carte des intensités macrosismiques du séisme de Ain Temouchent (Yelles-Chaouche, et al., 2004)*

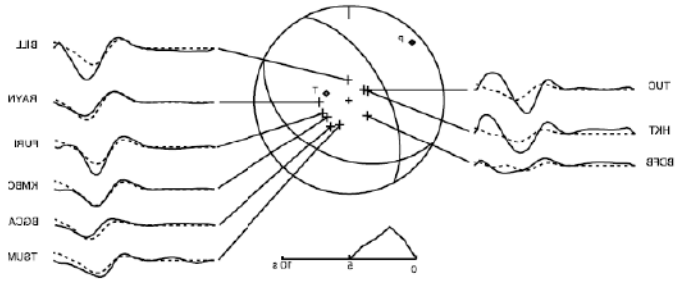

Figure 1-14 : *modélisation de la source sismique du séisme de Ain Temouchent par inversion des formes d'ondes télésismiques (Yelles-Chaouche et al., 2004)*

1.3.5. Le séisme de Boumerdes-Zemmouri du 21 mai 2003

Ce séisme a secoué la région de Boumerdes (50 km à l'est d'Alger) le 21 mai 2003 à 19h44mn (heure locale). De magnitude Mw = 6.8, il est le séisme le plus violent qui s'est produit au nord de l'Algérie après le séisme d'El Asnam de 1980. Il a été ressenti dans un rayon de 250 km. Il a fait plus de 2000 victimes sans compter les dégâts et les dommages dans les wilayas de Boumerdes, Alger, Tizi-Ouzou et Bouira. Ce séisme a fait l'objet de plusieurs

travaux scientifiques (Delouis et al., 2004 ; Bounif et al., 2004 ; Meghraoui et al., 2004 ; Deverchère et al., 2005 ; Semmane et al., 2005 ; Ayadi et al., 2008 ; Belabbès et al., 2009 ; Yelles et al., 2004). L'étude de la source sismique a été faite par divers auteurs en utilisant des approches différentes. Delouis et al. (2004) ont utilisé l'inversion jointe des données sismologiques (téléséisme) et des données géodésiques (GPS + soulèvement côtier). Semmane et al. (2005) ont utilisé les enregistrements des mouvements forts (accéléromètres) et les observations géodésiques (GPS + soulèvement côtier). Meghraoui et al., (2004) ont tenté de modéliser la source seulement en utilisant les données géodésiques (GPS + soulèvement côtier). Belabbès et al., (2009) ont modélisé la source sismique en utilisant les données géodésiques (GPS + soulèvement côtier) mais cette fois en ajoutant la donnée interférométrie radar (InSAR). Toutes ces études citées ci-dessus ont montré que la faille du séisme du 21/05/2003 est orientée NE-SW, et que son pendage plonge vers le SE. En allant dans les détails, ces modèles obtenus par différentes approches présentent des différences dans les valeurs qui quantifient la source à savoir l'azimut, le pendage et les dimensions de la faille. La **Figure 1-15** présente les projections en surface des limites des modèles de rupture utilisés par ces différents auteurs. Delouis et al. (2004), Semmane et al. (2005), Meghraoui et al., (2004) ont pu modéliser la quantité de glissement sur la faille. Deux aspérités sont observées sur tous les modèles présentés, l'une localisée au Sud-Ouest de l'hypocentre, l'autre au Nord-Est.

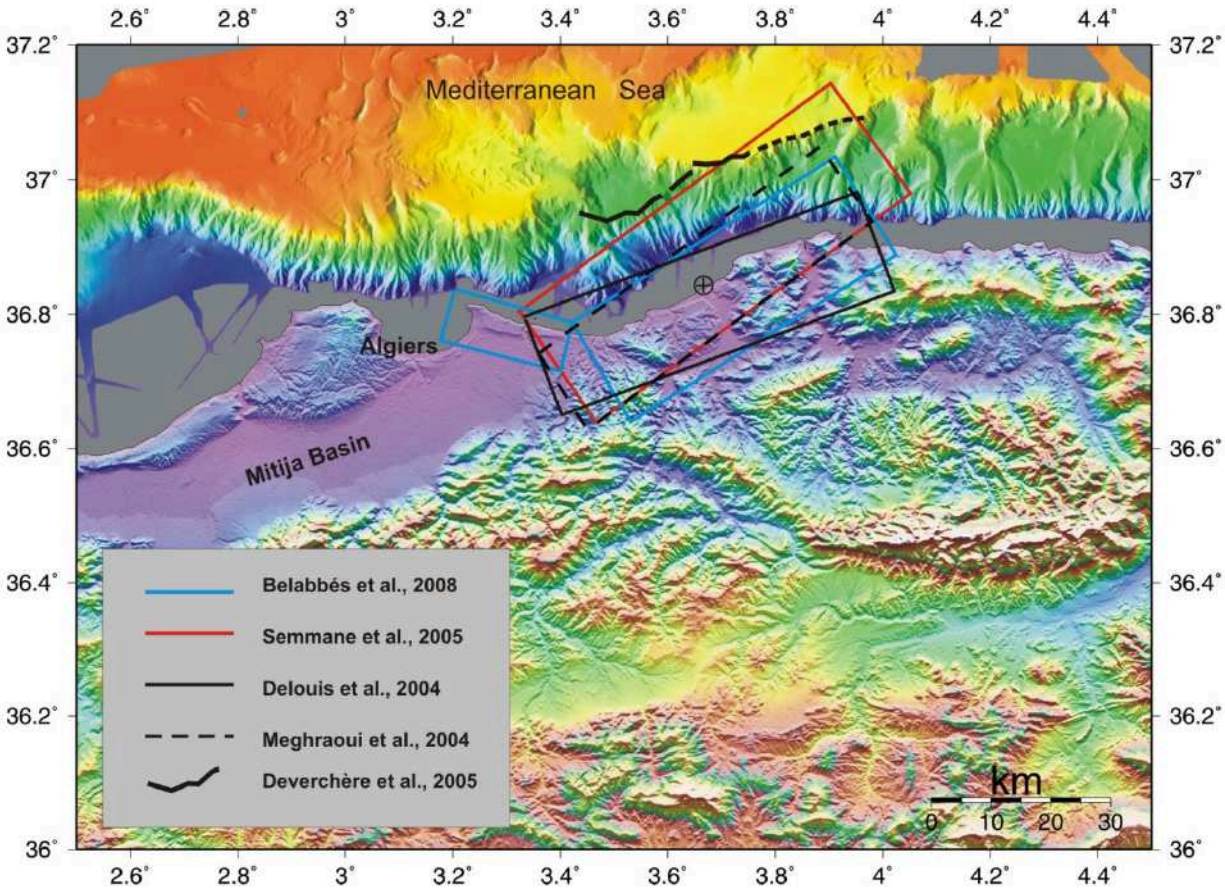

Figure 1-15 *Les rectangles de couleurs différentes montrent les projections en surface des limites des différents modèles de failles utilisés par différentes études (d'après Belabbès et al, 2009). Les modèles en bleu et en noir sont ceux proposés par Belabbès et al., (2009) et Delouis et al., (2004) respectivement. Le cercle avec une croix montre la localisation de l'épicentre du choc principal du séisme de Boumerdes-Zemmouri.*

1.4. Structures Actives au Nord de l'Algérie

1.4.1. L'Algérie dans le contexte Méditerranéen

Située à la frontière entre les plaques Afriques et Eurasie, l'évolution de la Méditerranée est liée au déplacement relatif des deux plaques tectoniques. Après un déplacement senestre de l'Afrique par rapport à l'Europe, les deux plaques convergent de la fin du Crétacé (**Figure 1-16**) à l'actuel et le bassin Méditerranéen évolue dans cet espace géologique qui ne cesse de se réduire depuis le Crétacé (Bouyahiaoui, 2014). Bien que la Méditerranée forme un bassin marin unique de 2.500.000 km^2, elle est en réalité composée d'une mosaïque de sous-bassins d'âges différents qui jalonnent la frontière des plaques.

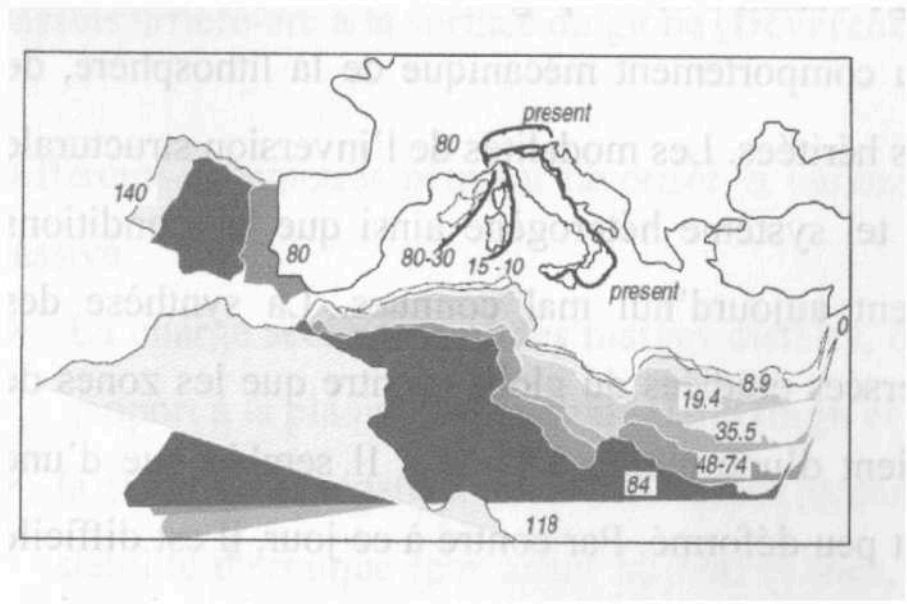

Figure 1-16 : *Déplacement relatif de l'Afrique et de l'Ibérie par rapport à l'Eurasie fixe, depuis le Crétacé (Bouyahiaoui, 2014). Les chiffrent sont en millions d'années.*

1.4.2. Evolution de la Méditerranée occidentale et du domaine nord-algérien

Il est admis maintenant que la configuration actuelle du domaine nord-algérien résulte de trois phases :

1- L'ouverture de l'océan Téthys, formé d'Est en Ouest à travers la Pangée entre le Permien supérieur et le Jurassique moyen, séparant les continents Gondwana au sud, et Laurasia au Nord, la terminaison actuelle de cet océan était alors l'actuelle Europe du Sud et l'actuelle Afrique du Nord (Bouyahiaoui, 2014).

2- La fermeture de l'océan Téthys maghrébin le long d'une zone de subduction située sous la bordure de la plaque Eurasienne, au nord du bassin, active depuis le Cénozoïque jusqu'à vers 18-15 Ma. Le résultat de cette fermeture est la formation de la chaîne alpine maghrébine qui s'étend du Maroc à la frontière Tunisienne (Bouyahiaoui, 2014).

3- L'ouverture à l'arrière de la subduction, de l'actuel bassin Algéro-provençal, à l'Oligo-Miocène, qui accompagne la fermeture progressive de l'océan Téthysien (**Figure 1-17**).

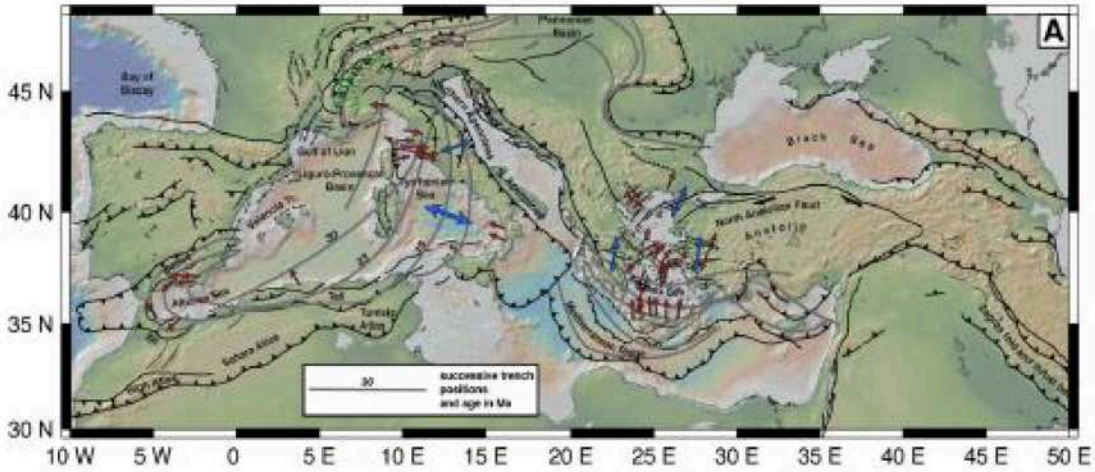

Figure 1-17 : *principales structures dans le bassin Méditerranéen, fronts de chevauchement, zones de subduction, les principales failles normales et en décrochement. En gris les positions successives des différentes fosses. Les flèches rouges montrent les directions des extensions ductiles et en bleu les directions en extension actives des arrières arc (Jolivet et al., 2009).*

1.4.3. Structures actives sur le continent

Le taux le plus élevé de la sismicité du Nord de l'Algérie est essentiellement localisé dans l'Atlas Tellien, cette sismicité est associée à des structures néotectoniques en plis et plis-faille de direction NE-SO (Belabbès, 2008) qui affectent les bassins néogènes post-nappes, nous citons les bassins néogènes de la Mitidja et du Chélif (**Figure 1-18**). Les bassins post-nappes correspondent à des bassins intra-montagneux, de direction E-O, et sont caractérisés par des structures actives avec une déformation en compression durant le Quaternaire affectant les dépôts du Quaternaire récent (Belabbès, 2008).

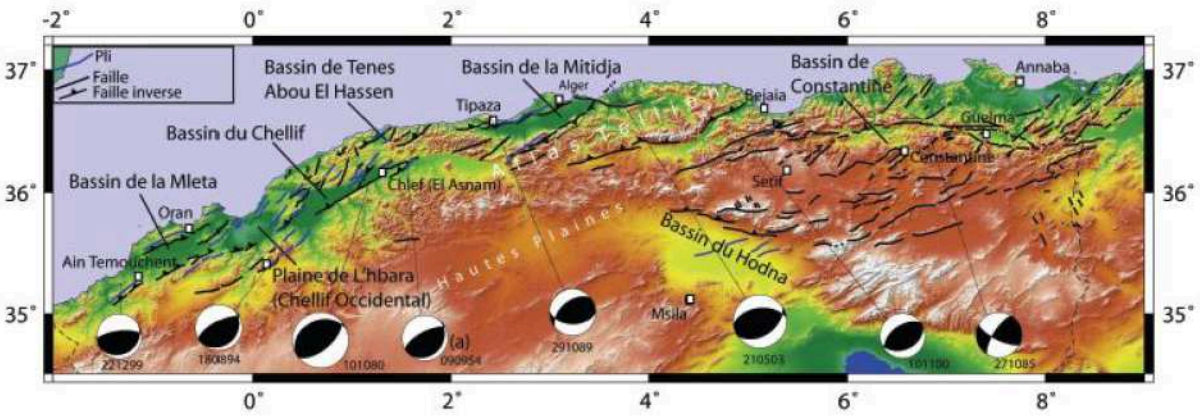

Figure 1-18 : *Carte morpho-tectonique du nord de l'Algérie (Meghraoui, 1988). Mécanismes au foyer (Global CMT) des principaux séismes depuis 1980. (a) Mécanisme au foyer du séisme d'El Asnam, 1954 (Espinoza et Lopez-Arroyo, 1984) (D'après Belabbès, 2008).*

1.4.4. Structures actives sur la partie offshore (marine)

Les données marines récoltées pendant les deux campagnes marines Maradja 1 (2003) et Maradja 2/Samra/(2005) ont permis de préciser les principaux traits morphologiques de la marge algérienne. De la côte vers le large cette marge peut être décomposée en différents domaines séparés par des ruptures de pente (**Figure 1-19a et 19b**).

On distingue ainsi du continent vers le bassin, le plateau continental, le talus continental, le glacis et enfin la plaine abyssale (Bouyahiaoui 2014).

Le plateau continental s'étend du trait de la côte jusqu'à une rupture de pente qui le sépare du talus continental. Des variations morphologiques importantes sont constatées à la limite du plateau continental (isobathe 200 m) : ce plateau est réduit (<10 km) en face des massifs élevés qui jalonnent la côte, et s'élargit en face des bassins ; il peut atteindre une largeur moyenne de l'ordre de 10-15 km dans les zones de Skikda, Béjaïa et à l'Ouest d'Alger, et de 25 km dans les zones d'Annaba et de l'Ouest d'Oran (Bouyahiaoui, 2014).

Le talus continental (pente continentale), relie le bord du plateau continental au bassin profond, il présente une forte pente dépassant souvent les 10° et pouvant atteindre localement 20°, sa largeur varie entre 8 et 20 km, à Annaba, il peut atteindre les 35 km (Bouyahiaoui, 2014, Kherroubi, 2011), cet escarpement est caractérisé par de nombreux canyons qui drainent les sédiments détritiques de l'arrière pays vers la plaine abyssale.

Le glacis assure la jonction entre talus continental et la plaine abyssale par une pente douce (2-3°), il constitue le lieu de transit et d'accumulation des sédiments et s'étend sur des distances variant de 0 à 30 km du pied de pente, et il est à une profondeur qui peut aller de 2400 à 2600 m (Bouyahiaoui, 2014).

La plaine abyssale correspond au vaste domaine, relativement plat qui relie les différentes marges continentales bordant un bassin. Une profondeur maximale de 2800 m est atteinte dans la partie abyssale du bassin algérien. La plaine abyssale est située à une distance moyenne de la côte de l'ordre de 50 km (Bouyahiaoui, 2014).

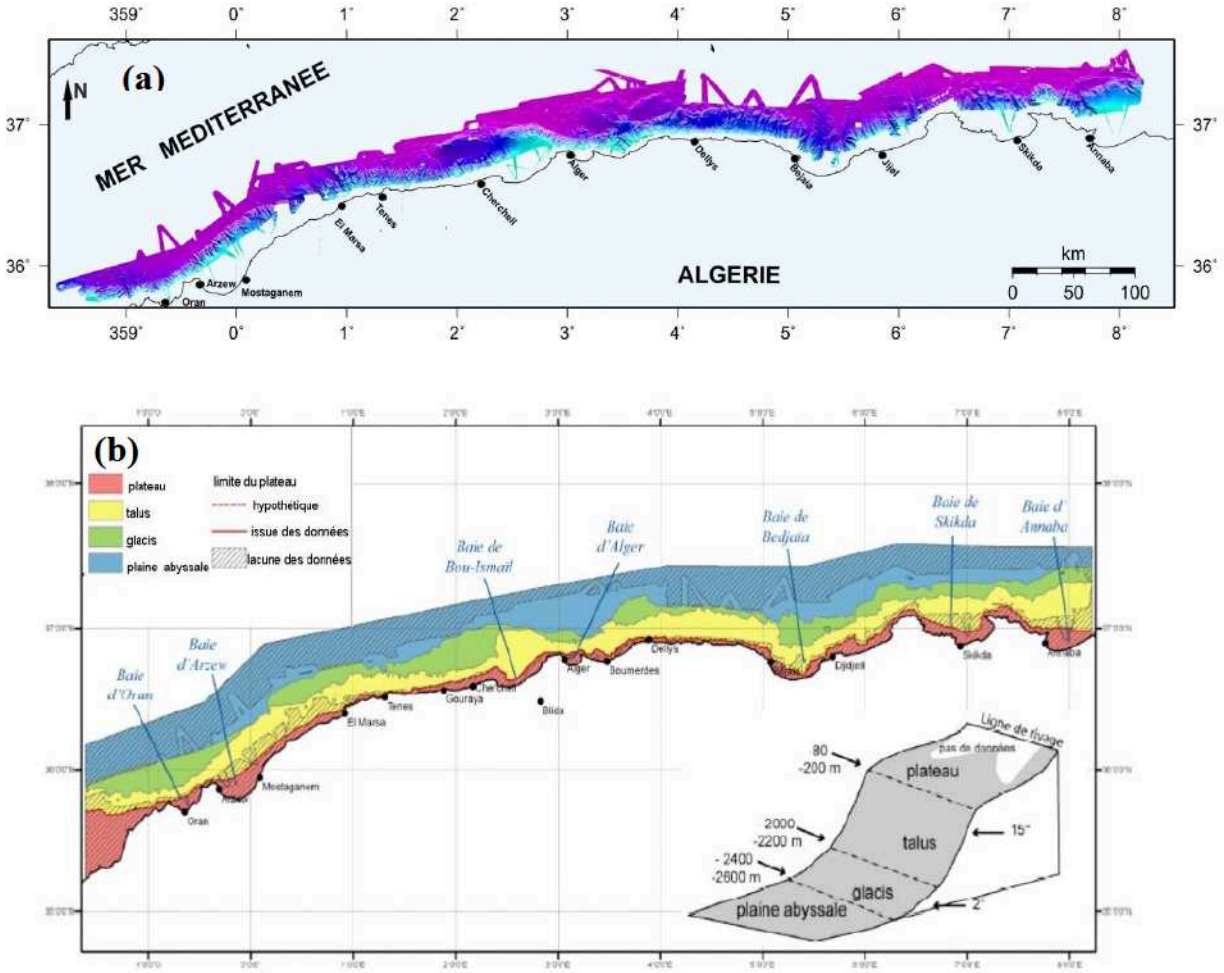

Figure 1-19 : a) *carte bathymetrique 3D de la marge algérienne, suite aux campagnes marine Maradja1 et Maradja2/Samra. b) en bas schéma de délimitations des éléments morphologiques. En haut, configuration schématique des domaines physiographiques de la marge algérienne (Bouyahiaoui, 2014).*

1.4.5. Déformation en mer

Les travaux de Deverchère et al., (2005), Domzig (2006), Kherroubi et al., (2009), Yelles-Chaouche et al. (2009a et 2009b), Leprêtre et al., (2013), Bouyahiaoui (2014) et (Badji et al., 2015), ont montré que la déformation plio-quaternaire n'est pas homogène tout au long de la marge algérienne. Dans le secteur de Tipaza (l'Ouest d'Alger), elle est caractérisée par la plus longue faille récente sur la marge algérienne, c'est la faille de Khyr-Al-Din de longueur ~80 km (**Figure 1-20**). Dans le secteur de Boumerdes (l'Est d'Alger), la déformation est marquée par le développement de bassins syntectoniques qui montrent un basculement des séries et des dépôts en éventail, il s'agit (1) des bassins en piggy-back, en bordure du bassin profond et (2) des bassins en rollover, le long de la pente profonde, de taille plus modeste, qui se développent le long d'une faille à pendage nord fortement inclinée. Un long segment linéaire est présent au large de Mostaganem au Nord-Ouest de l'Algérie (**Figure 1-19**), ce segment ne montre pas d'indice de compression (Badji et al, 2015). Dans le secteur de Jijel à l'est de l'Algérie (**Figure 1-19**), des failles inverses de la marge se propagent vers le bassin selon un système en plats et rampes. Un rollover est observé en bordure de bassin profond, où émerge une faille inverse (Bouyahiaoui, 2014). Dans le secteur de Annaba **(Figure 1-19)**, Kherroubi et al., (2011) ont permis de cartographier un ensemble de plis et chevauchement en pied de marge.

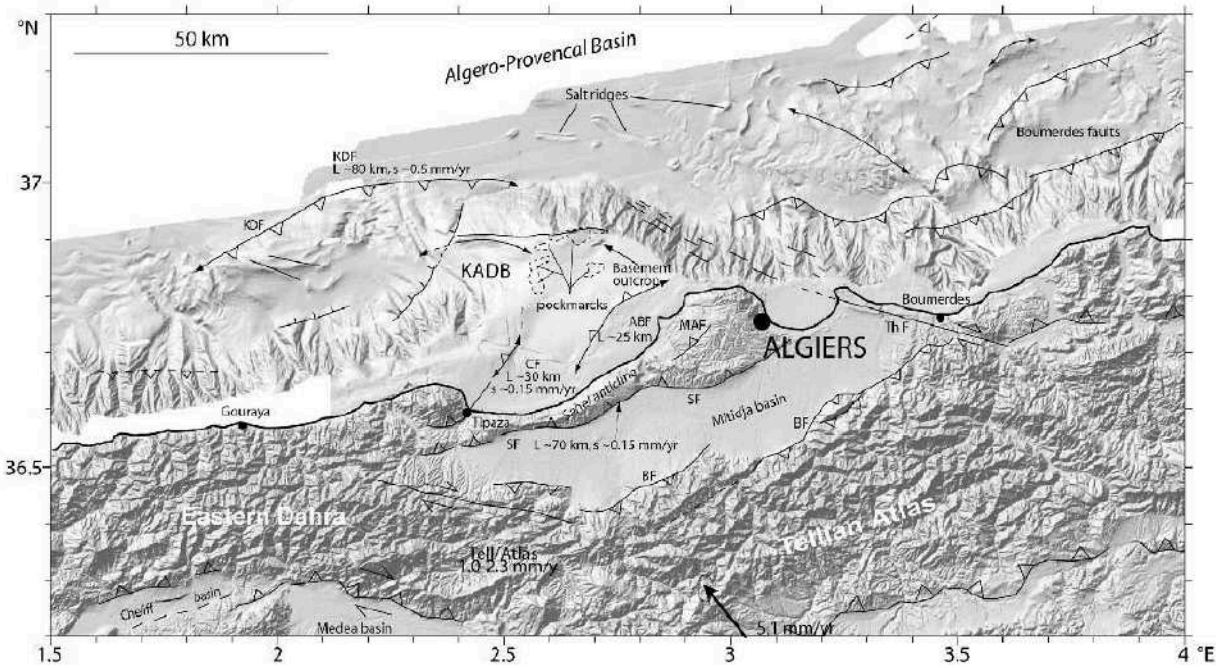

Figure 1-20 : *Principales structures actives de la partie offshore et continentale de la région d'Alger. SF :Faille du Sahel, BF : Faille de Blida, Th.F : Faille de Thénia,MAF : Faille de Mahelma, ABF : Faille de Ain Bénian,KDF :Faille de Kayr-Al-Din, CF : Faille de Chenoua. La flèche en bas indique la direction et le taux du vecteur déplacement (d'après Yelles-Chaouche et al., 2009b).*

Chapitre 2: Méthodologie d'inversion de la source

2.1. L'approche par inversion des formes d'ondes pour le mécanisme au foyer

Le mécanisme au foyer est généralement déterminé par l'une des deux approches suivantes : à partir des premiers mouvements au niveau d'un réseau de stations, ou bien par inversion des formes d'ondes. Par formes d'ondes on entend la variation d'amplitude du signal sismique au cours du temps. L'approche par formes d'ondes est plus complexe à mettre en œuvre mais elle présente un certain nombre d'avantages. En modélisant les sismogrammes complets, les différents types d'ondes qui apportent chacun leur contrainte propre sur les paramètres du mécanisme sont tous pris en compte, ce qui peut permettre de contraindre un mécanisme au foyer avec moins de stations. Par ailleurs, la modélisation des formes d'ondes s'effectuant à relativement basse fréquence, elle est moins sensible au modèle de vitesse que celle des premiers mouvements. En outre, la modélisation des formes d'ondes permet de déterminer le moment sismique, donc la magnitude de moment Mw.

Nous rappelons ci-dessous la définition des paramètres du mécanisme au foyer ainsi que les éléments de la théorie de la source sismique qui contrôlent les formes d'ondes. Finalement nous présenterons la méthodologie utilisée pour l'inversion des formes d'ondes. Cette description reprend des éléments de l'habilitation à diriger des recherches (HDR) de Bertrand Delouis (2009), tout en y apportant certains compléments.

2.2. Définition des paramètres du mécanisme au foyer

Le mécanisme au foyer peut être décrit par trois angles (**Figure 2-1**) :

- l'azimut (ou strike) Φs : angle (0 à 360° en tournant dans le sens horaire) entre la direction du Nord et la direction du plan de faille correspondant par convention à la direction de la plus grande pente du plan - 90° ;

- le pendage (ou dip) δ : angle entre le plan horizontal et la ligne de plus grande pente du plan de faille (de 0° pour un pendage horizontal à 90° pour un pendage vertical) ;

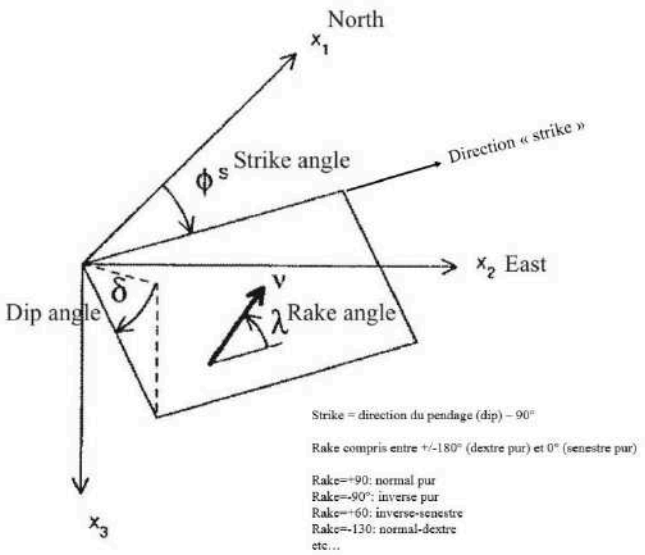

Figure 2-1 *: Définition des angles strike, dip, et rake d'un mécanisme au foyer (Delouis, 2009).*

- l'angle de glissement (ou rake) λ : angle entre l'horizontale sur le plan de faille et la direction du vecteur glissement $\vec{v}$ sur ce même plan de faille. Par convention (Aki and Richards, 1980), cet angle varie entre -180° et +180°. Les valeurs positives traduisent un mouvement de montée du bloc supérieur, c'est-à-dire une composante de jeu inverse. Les valeurs négatives traduisent un mouvement de descente du bloc supérieur, c'est-à-dire une composante de jeu normal. La valeur 0° correspond à un glissement purement senestre. Les valeurs -180° et +180° correspondent à un seul et même glissement purement dextre.

Notons que si un triplet (strike, dip, rake) décrit de manière évidente l'un des plans nodaux du mécanisme au foyer, avec le mouvement qui lui est associé, il définit de manière unique le second plan nodal. En effet, le second plan nodal est par définition le plan perpendiculaire au mouvement sur le premier plan nodal (et réciproquement). Les deux plans nodaux sont perpendiculaires.

Les plans nodaux sont tels que les ondes P émises ont une amplitude nulle si le rai sismique à la source est contenu dans l'un de ces plans. L'onde P change de polarité (compression à dilatation ou vice-versa) au passage des plans nodaux. Au total, les deux plans nodaux subdivisent l'espace en quatre cadrans, deux en compression diamétralement opposés, deux en dilatation également diamétralement opposés (**Figure 2-2** sous forme d'un bloc diagramme et **Figure 2-3** pour un exemple en projection stéréographique classique). La correspondance entre des jeux de faille et les mécanismes au foyer correspondants est illustrée sur la **Figure 2-4**.

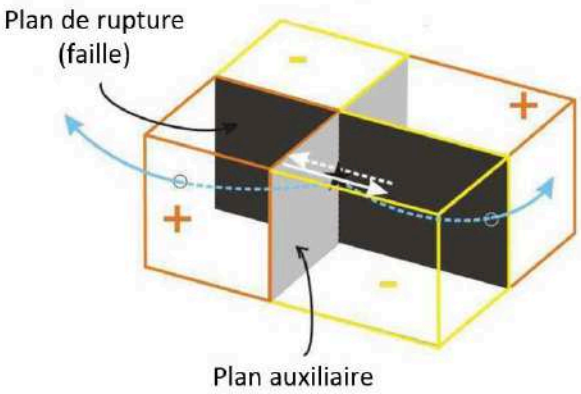

Figure 2-2. *Illustration montrant les quatre cadrans dans le cas d'une faille verticale décrochante sénestre. Le plan de faille est gris foncé, le plan auxiliaire gris clair, les cadrans en compression (+) en orange, et les cadrans en dilatation (-) en jaune. Les flèches bleues matérialisent des rais sismiques à la source. L'étoile noire représente le foyer (hypocentre). D'après J. Balestra (2017).*

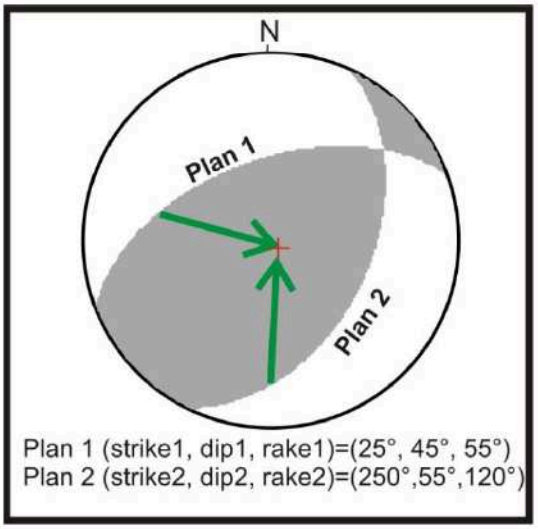

Figure 2-3 : *exemple de mécanisme au foyer avec les paramètres strike, dip, rake des deux plans nodaux. Le mécanisme est dessiné dans sa représentation la plus classique, en projection stéréographique de type Schmidt hémisphère inférieure. Les zones grisées correspondent aux cadrans en compression (premier mouvement de l'onde P en compression), les zones blanches sont les cadrans en dilatation (premier mouvement de l'onde P en dilatation). Les flèches vertes correspondent à la direction et au sens de mouvement du bloc supérieur pour chacun des plans nodaux.*

L'un des plans nodaux correspond au plan de rupture sur la faille, alors que l'autre plan nodal n'a pas de matérialisation géologique et est dénommé le plan auxiliaire. Si le mécanisme au foyer est déterminé à l'aide d'un modèle de source ponctuelle, ce qui est implicite pour la méthode des polarités et qui est le cas le plus fréquent en modélisation des formes d'ondes, il n'est pas possible d'établir quel est le plan nodal qui correspond au plan de rupture. Dans ce cas, des considérations autres, comme la distribution des répliques, la distribution des déformations de surface, ou bien la connaissance de la faille géologique, peut permettre de lever cette ambiguïté.

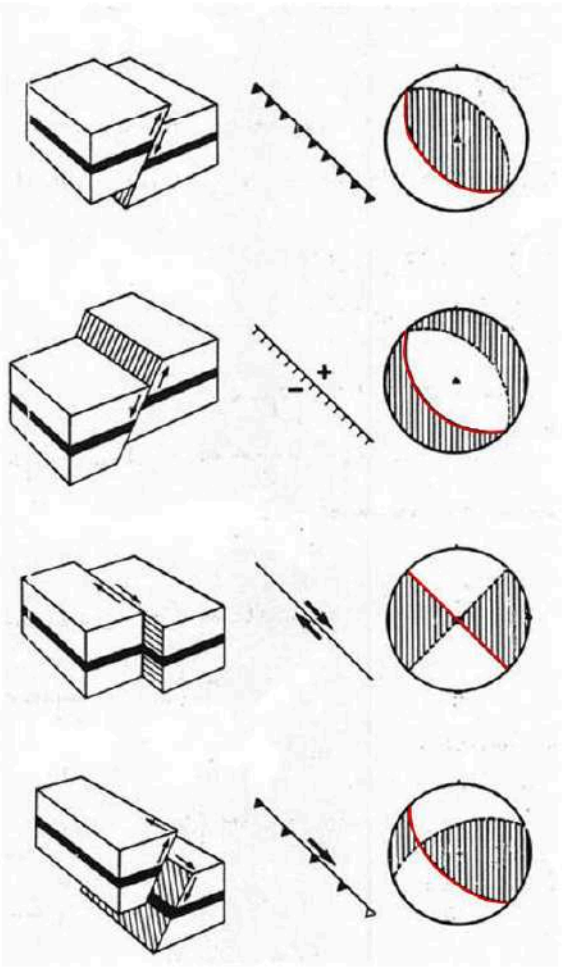

Figure 2-4. *Illustration des relations entre mécanismes au foyer et jeux de faille. Sur les mécanismes les cadrans en compression sont hachurés et le plan nodal correspondant au plan de faille est souligné en rouge. L'autre plan du mécanisme est le plan auxiliaire. Les jeux de faille sont, de haut en bas, inverse pur, normal pur, décrochant pur (dextre sur la faille), et décrochant-inverse (jeu oblique, dextre-inverse sur la faille).*

2.3. Le déplacement du sol produit par une source ponctuelle

Nous limitons notre description au mouvement du sol en déplacement généré par une source double couple de type dislocation : un mouvement cisaillant sur une surface de dislocation d'épaisseur infinitésimale (la faille). Ce modèle est adapté pour décrire la source de la plupart des séismes tectoniques.

Dans le cas d'une source ponctuelle cisaillante dans un milieu élastique homogène, isotrope, et infini, le déplacement du sol $\vec{u}(t,r)$ à une distance « r » et à un instant « t » autour de la source est donné par la relation suivante :

$$\vec{u}(t,r) = \tag{1}$$

$$\frac{1}{4\pi\rho r^4} \cdot \overrightarrow{R_p} \cdot \int_{r/\alpha}^{r/\beta} \tau \cdot M_0(t-\tau)d\tau \tag{1.1}$$

$$+\frac{1}{4\pi\rho\alpha^2} \cdot \frac{1}{r^2} \cdot \overrightarrow{R_l^P} \cdot M_0\left(t-\frac{r}{\alpha}\right) + \frac{1}{4\pi\rho\beta^2} \cdot \frac{1}{r^2} \cdot \overrightarrow{R_l^S} \cdot M_0\left(t-\frac{r}{\beta}\right) \tag{1.2}$$

$$+\frac{1}{4\pi\rho\alpha^3} \cdot \frac{1}{r} \cdot \overrightarrow{R_l^P} \cdot \dot{M}_0\left(t-\frac{r}{\alpha}\right) + \frac{1}{4\pi\rho\beta^3} \cdot \frac{1}{r} \cdot \overrightarrow{R_l^S} \cdot \dot{M}_0\left(t-\frac{r}{\beta}\right) \tag{1.3}$$

« α » et « β » correspondent à la vitesse des ondes P et des ondes S respectivement, « ρ » à la densité du milieu. Le symbole « . » signifie la dérivée temporelle ($^{d}/_{dt}$).

Le terme « M_0 » correspond au moment sismique décrit par :

$$M_0 = \mu \cdot S \cdot \Delta u \tag{2}$$

où « μ » est la rigidité du milieu (module élastique de cisaillement), « S » la surface de la rupture, et « Δu » le glissement moyen. Ce terme intervient dans chaque champ pour décrire le déplacement $u(t,r)$. Notons aussi que le terme M_0 apparait sous la formulation « $\dot{M}_0$ » (taux de moment sismique) dans la relation (1 .3).

Ce déplacement $u(t,r)$est décrit par trois termes que l'on peut définir par :

- un terme dit de champ proche (indice p) (1.1) ;
- un terme dit de champ intermédiaire (indice i) (1.2) ;
- un terme dit de champ lointain (indice l) (1.3).

Les facteurs « $\frac{1}{4\pi\rho}$ », « $\frac{1}{4\pi\rho\alpha^2}$ », « $\frac{1}{4\pi\rho\beta^2}$ », « $\frac{1}{4\pi\rho\alpha^3}$ » et « $\frac{1}{4\pi\rho\beta^3}$ » sont des constantes qui dépendent des propriétés du milieu de propagation.

Les facteurs « $\frac{1}{r^4}$ », « $\frac{1}{r^2}$ », et « $\frac{1}{r}$ » décrivent l'atténuation géométrique pour les termes en champs proche « p », intermédiaire « i », et lointain « l », respectivement. A noter que pour le champ proche, la distance r apparait aussi dans les bornes de l'intégrale, ce qui va modifier l'atténuation géométrique qui au final sera en $1/r^2$ comme pour le champ intermédiaire.

Les termes « $\overrightarrow{R_p}$ », « $\overrightarrow{R_l^P}$ » et « $\overrightarrow{R_l^S}$ » ; « $\overrightarrow{R_l^P}$ » et « $\overrightarrow{R_l^S}$ » sont les facteurs de radiations des ondes correspondantes, avec les exposants P ou S décrivant des ondes qui voyagent à la vitesse des ondes P (α) ou des ondes S (β), qui modulent l'amplitude de ces ondes et

définisse sa polarité (son signe), pour chaque champ, en fonction de la position relative du rai sismique au foyer par rapport au mécanisme au foyer.

Les termes « $t - \frac{r}{\alpha}$ » et « $t - \frac{r}{\beta}$ » décrivent le décalage temporel dû au temps de propagation de l'onde P et de l'onde S.

Les termes $M_0\left(t - \frac{r}{\alpha}\right)$, $M_0\left(t - \frac{r}{\beta}\right)$, $\dot{M}_0\left(t - \frac{r}{\alpha}\right)$, et $\dot{M}_0\left(t - \frac{r}{\beta}\right)$ sont des fonctions du temps. L'ensemble des fonctions du temps dans l'équation (1) contrôlent la forme d'onde, donc la forme du sismogramme (pour un milieu de propagation très simple homogène).

2.4. Facteurs de radiation et amplitudes des ondes

Nous ne décrivons ici que les facteurs de radiation des ondes de volume P et S, donc en champ lointain.

L'amplitude des ondes générées va varier en fonction de la direction du rai sismique à la source par rapport à la direction du mouvement sur le plan de faille. L'ensemble des vecteurs $\overrightarrow{R^P}$ et $\overrightarrow{R^S}$ dont la norme module l'amplitude des ondes générées est regroupé sous une enveloppe en forme de lobe. Il y a quatre lobes au total, dont la distribution est en relation avec les quatre cadrans décrits précédemment (**Figure 2-2**). Ces lobes décrivent les diagrammes de radiation des ondes P et S (**Figure 2-5**). Ces deux diagrammes montrent donc que la radiation des ondes P et des ondes S est différente. On voit notamment que les lobes de l'onde P s'annulent au niveau du plan de faille et du plan auxiliaire (plans nodaux) mais montrent un maximum à 45° de ceux-ci. Les lobes de l'onde S montrent un maximum au niveau du plan de faille et du plan auxiliaire, mais s'annulent à 45° de ceux-ci. Un point commun, les ondes P et S sont toutes deux nulles au niveau de la direction intersection des deux plans nodaux. Les flèches rouges en pointillés sur les diagrammes de la **Figure 2-5** renseignent sur la polarisation (direction et sens du mouvement) des ondes P et S : la polarisation des ondes P est dans la direction de propagation du rai sismique (ce sont des ondes de compression - dilatation), la polarisation des ondes S est perpendiculaire à la direction de propagation du rai sismique (ce sont des ondes de cisaillement).

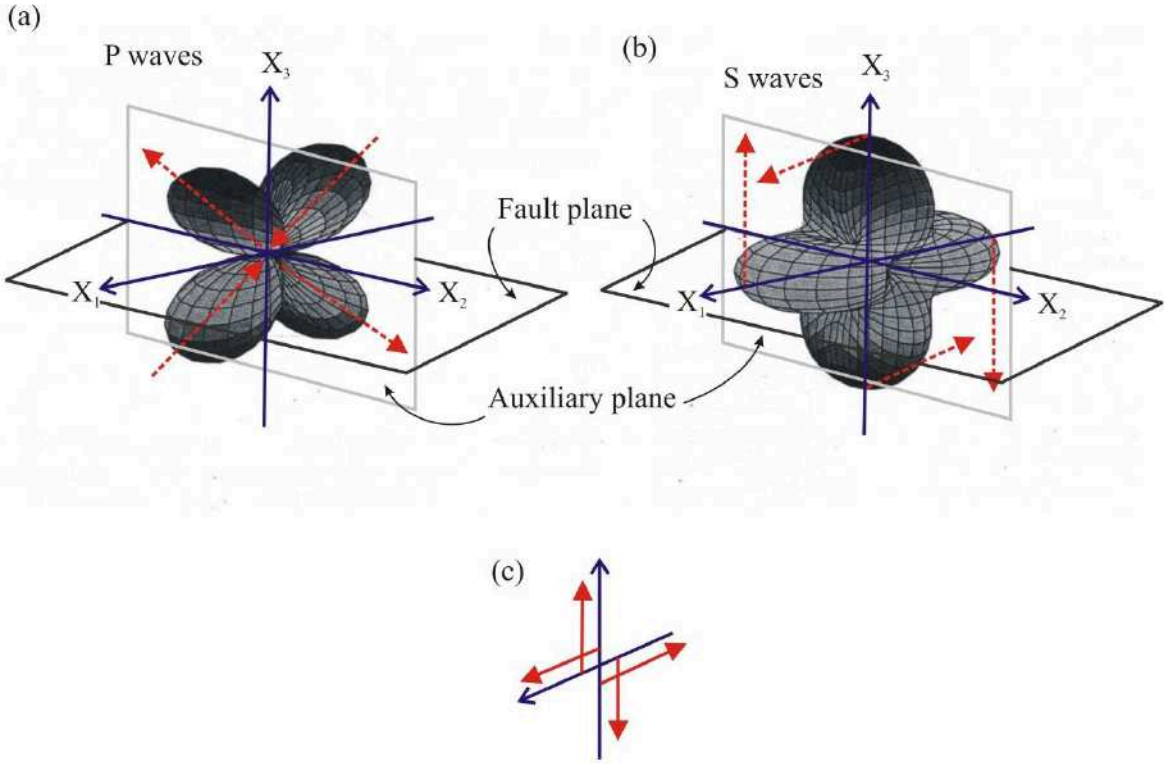

Figure 2-5: *Diagramme de radiation des ondes P (a) et S (b) pour un point source. Le plan de faille ici est horizontal, le plan auxiliaire vertical. Flèches rouges pointillées: polarisation de l'onde P (a) et de l'onde S (b). (c) Double couple de forces produisant le même effet sur le milieu qu'un mouvement cisaillant ponctuel sur le plan de faille (ou sur le plan auxiliaire).*

Le diagramme de radiation des ondes de volume est un paramètre essentiel pour la détermination des mécanismes au foyer : il est dépendant des paramètres du mécanisme : l'azimut « Φs » (strike), le pendage « δ » (dip), et l'angle de glissement « λ » (rake). Il en va de même pour les autres types d'onde, champs intermédiaire et proche, dont les diagrammes de radiation ne sont pas représentés ici.

2.5. La fonction temporelle de la source et la forme des ondes

Soit $M_0(t)$ la fonction du moment sismique au cours du temps (**Figure 2-6b**). La courbe de la fonction $M_0(t)$ décrit l'évolution du moment sismique qui croit puis se stabilise pour atteindre la valeur plateau M_0 (moment sismique final) en fin de rupture.

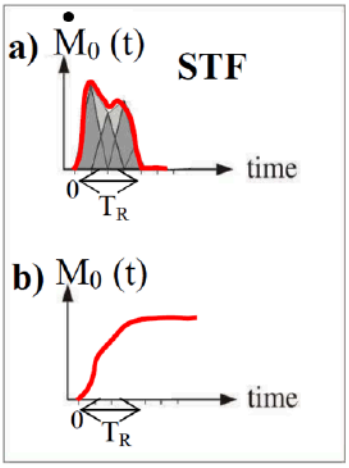

Figure 2-6: *(a) courbe rouge : fonction taux de moment sismique* $\frac{d}{dt}M_0(t)$. *Egalement appelée la fonction temporelle de la source (STF). Les triangles se chevauchant mutuellement sont une manière de discrétiser cette fonction, la somme des aires des triangles étant approximativement égale à l'aire sous la courbe rouge. (b) fonction moment sismique* $M_0(t)$. *"T_R": temps (durée) de la rupture (Delouis, 2009).*

La dérivée $\dot{M}_0\,(t)$de la fonction M_0(t) décrit l'évolution du taux de moment sismique au cours du temps. Cette fonction taux $\dot{M}_0(t)$ (**Figure 2-6a**) décrit ainsi la fonction temporelle de la source sismique (STF). La fonction taux $\dot{M}_0\,(t)$décrit également la forme et la variation de l'amplitude des ondes P et S en champ lointain en fonction du temps, ce que l'on simplifie souvent en l'appelant la « forme d'onde », ou « waveform » en anglais. Elle est définie par les termes « $\dot{M}_0\left(t-\frac{r}{\alpha}\right)$ » et « $\dot{M}_0\left(t-\frac{r}{\beta}\right)$ », les termes « $\frac{1}{4\pi\rho\alpha^3}\cdot\frac{1}{r}\cdot\overrightarrow{R_i^P}$ » et « $\frac{1}{4\pi\rho\beta^3}\cdot\frac{1}{r}\cdot\overrightarrow{R_i^S}$ » ne variant pas au cours du temps dans le cas d'une source ponctuelle.

La forme des ondes en champ intermédiaire est donnée par la fonction « $M_0\,(t-\frac{r}{c})$ », affectée d'un signe positif ou négatif selon le signe du facteur de radiation correspondant (équation 1.2), c'est-à-dire une fonction proche d'une fonction rampe.

La forme des ondes en champ proche est donnée par l'intégration de la fonction « $M_0\,(t-\tau)$» (**Figures 2-7 et 2-8**). L'intégrale et la double intégrale sont aussi des fonctions croissantes mais avec une tendance nettement plus marquée due à la forme parabolique de l'intégrale de $M_0\,(t-\frac{r}{c})$. Le champ proche démarre dès le début du sismogramme, c'est-à-dire en même temps que l'onde P, comme l'indiquent les termes fonction de (t-r/α).

Le terme champ proche génère un déplacement permanent (on dit aussi statique) qui décroit en $\frac{1}{r^2}$ (1.1). En effet le terme $1/r^4$ se combine avec la fonction intégrale et Wu and Ben-Menahem (1985) ont montré que globalement cela produisait une atténuation en $1/r^2$. Le déplacement $u(t,r)$ dépend du diagramme de radiation en champ proche « $\overrightarrow{R_p}$ », et de l'évolution de l'aire sous la courbe de la fonction M_0 (t), donc l'évolution du glissement et de la surface de rupture au cours du temps.

En champ intermédiaire, le déplacement décroit en $\frac{1}{r^2}$. Le déplacement $u(t,r)$ dépend de la fonction $M_0\left(t-\frac{r}{c}\right)$, où « c » est la vitesse des ondes. Il y a donc un terme qui démarre avec l'onde P et dure le même temps qu'elle, et un terme qui démarre avec l'onde S pour une durée identique.

Les champs proches et intermédiaires s'atténuent donc plus rapidement que le champ lointain. Le déplacement statique est donc généré par la somme des champs proches et intermédiaires, le déplacement décroissant très vite à mesure que l'on s'éloigne de la source.

La **Figure 2-7** illustre un sismogramme synthétique construit à partir de l'influence des différents champs, et la **Figure 2-8** un exemple réel.

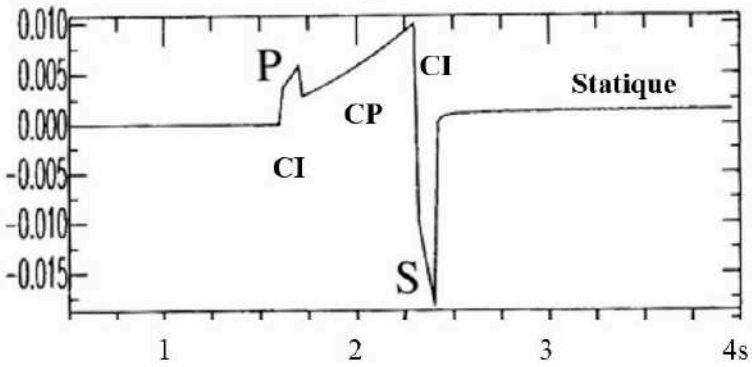

Figure 2-7: *Sismogramme théorique en déplacement pour un point source à quelques kilomètres de la source. La fonction source temporelle est un simple créneau de durée 0.2s. Le milieu de propagation est un demi-espace homogène. P : onde P ; S : onde S ; CP : champ proche ; CI : champ intermédiaire ; statique : déplacement statique résiduel. Composante verticale.*

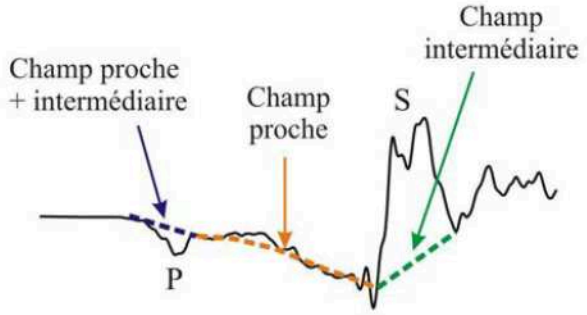

Figure 2-8 : *cas d'un séisme réel, séisme 30 km au large de Nice, Ml 4.5 Février 2001 (Delouis, 2009).*

2.6. Les conditions pour une modélisation en point source

Le calcul du champ d'onde complet en déplacement, à distance proche de la source, et à la surface libre de la terre peut être abordé par l'approche des nombres d'ondes discrets avec la prise en compte d'un milieu stratifié tabulaire (Bouchon, 1981). Les fonctions de Green sont calculées dans le domaine fréquentiel, le retour dans le domaine temporel se faisant avec une transformée de Fourier inverse.

Le modèle point source n'est utilisable que sous les deux conditions suivantes :

- la distance épicentrale doit être bien plus grande que la dimension de la rupture ;
- les longueurs d'ondes doivent aussi être bien plus grandes que la dimension de la rupture.

La première condition fait que les ondes émises par les différentes zones de la rupture arrivent à une station donnée avec des différences angulaires minimes. Les ondes produites par l'ensemble des points qui ont rompus parcourent quasiment la même distance, et les effets de ces décalages angulaires lors de la propagation sont négligeables. La deuxième condition signifie que les grandes longueurs d'ondes échantillonnent les traits dominants de la rupture, sans être perturbées par les détails de cette même rupture. Il est ainsi plus simple d'aborder la détermination du mécanisme au foyer de la rupture. L'approximation « point source » n'est valable que si ces deux conditions sont bien vérifiées.

Toutefois quelques conditions supplémentaires sont à vérifier :

- Il faut disposer de signaux large bande dont la réponse instrumentale (au moins le facteur de sensitivité permettant de reconstituer les amplitudes physiques) est bien connue.

- Il faut trouver la bande de fréquence adéquate pour modéliser les formes d'ondes.

La question du filtrage est peut-être le point crucial de l'approche par formes d'ondes. Puisque l'on ne dispose généralement pas d'un modèle de vitesse 3D précis pour la propagation des ondes, on cherche à modéliser les ondes de plus basse fréquence possible, afin de s'affranchir de l'inadéquation du modèle de vitesse, généralement un simple modèle 1D. Par ailleurs, il a été montré que les ondes basses fréquences, en particulier celles issues des termes champs proche en déplacement, ont un fort pouvoir de résolution du mécanisme au foyer (Delouis and Legrand, 1999). La difficulté est que les basses fréquences peuvent être contaminées par du bruit, lié aux conditions de site des instruments, ou bien liés à des problèmes instrumentaux (en particulier effet de « tilt »). Il faut donc trouver la fréquence de coupure des basses fréquences permettant d'éliminer les basses fréquences contaminées par du bruit tout en préservant au maximum celles produites par le séisme. La fréquence de coupure des hautes fréquences doit, elle, être choisie afin d'éliminer les hautes fréquences qui ne peuvent pas être reproduite par la combinaison d'un modèle de source simple et d'un modèle de propagation également simple, tout en préservant une bande de fréquence suffisante pour contraindre le mécanisme au foyer. Delouis (2016) a montré qu'une fréquence maximale égale à trois fois la fréquence minimale permettait de définir une bande de fréquence adéquate. L'expérience montre que ce ratio peut même être descendu à deux (fmax = 2 x fmin).

2.7. Traitement des données et Méthode d'inversion

Les données utilisées dans cette thèse pour le calcul des solutions focales sont obtenues à partir des enregistrements du réseau de surveillance sismologique du CRAAG (ADSN : Algerian Digital Seismic Network). C'est un réseau composé de stations large bandes, courtes période et d'accéléromètres. Dans le but d'obtenir les déplacements, les sismogrammes large bandes sont intégrés une fois et les accéléromètres deux fois. Les sismogrammes en déplacement sont filtrés entre 0.03 Hz et 0.15 Hz, mais chaque sismogramme aura sa propre bande de filtrage. Ceci afin d'enlever les hautes fréquences et ne laisser que les basses fréquences qui contiennent les informations nécessaires pour contraindre l'orientation de la

faille et la direction du glissement et qui sont moins sensible au modèle de propagation et à la complexité de la rupture.

Deux approches d'inversion ont été utilisées dans le cadre de la thèse :

- Celle présentée dans Delouis et Legrand (1999), modifiée dans Orgulu et al. (2005), et adaptée au point source. Nous la nommerons « Méthode 1 » dans la suite de cette section.

- La méthode FMNEAR (Delouis, 2014).

Les deux approches ont de nombreux points communs et quelques différences.

2.7.1. Points communs

Les paramètres recherchés sont ceux du mécanisme au foyer, strike, dip, rake **(Figure 2-1)**, ainsi que le moment sismique M_0 permettant le calcul de la magnitude de moment Mw.

L'exploration des paramètres strike, dip, rake est fait par une recherche sur grille (*grid search*), en deux étapes. La première étape est une exploration de l'ensemble des valeurs possibles mais avec un pas assez large. La deuxième une exploration plus fine autour des meilleures solutions trouvées lors de la première étape.

Dans les deux cas, le critère de sélection des solutions est la minimisation d'une fonction d'ajustement des formes d'ondes de type erreur quadratique moyenne RMS normalisée :

$$RMS = \frac{\sum_k rms_k}{nstat} \tag{3}$$

nsta : nombre de stations, $\sum_k$: somme de k=1 à nstat, avec :

$$rms_k = \sqrt{\frac{\sum_j (o_j - c_j)^2}{\sum_j (o_j)^2}} \tag{4}$$

$\sum_j$ est la somme de j=1 a N, N étant le nombre d'échantillons pour les trois composantes des sismogrammes. O_j et C_j sont respectivement les sismogrammes observés et calculés. Le signal

synthétique est calculé en utilisant la méthode du nombre d'onde discret de Bouchon (1981) pour un modèle de vitesse 1D.

La fonction temporelle de la source (STF) $\dot{M}_0(t)$ est représentée par une fonction triangulaire (Méthode 1) ou plusieurs (FMNEAR) se chevauchant mutuellement comme montré sur la **Figure 2-6a**.

2.7.2. Différences principales

La différence essentielle entre les deux approches est liée au nombre de solutions testées : plusieurs milliers dans le cas de la Méthode 1 et un peu moins d'une centaine pour FMNEAR. De fait, FMNEAR est une évolution de la Méthode 1 orientée vers le temps quasi réel, donc ayant vocation à être beaucoup plus rapide. La recherche sur grille a donc été optimisée en ce sens, tout en gardant l'efficacité nécessaire pour une détermination assez précise du mécanisme au foyer. Chaque solution testée faisant appel à un calcul des sismogrammes synthétiques à toutes les stations incorporées dans l'analyse, le temps de calcul augmente fortement en fonction du nombre de solutions testées. Ce temps de calcul est typiquement de 12h ou plus avec la Méthode 1 et de quelques minutes à dizaines de minutes avec FMNEAR (dépend du nombre de stations, de la distance des stations, du filtrage utilisé). La Méthode 1 offre l'avantage d'une exploration plus complète et systématique de l'espace des solutions, permettant de représenter graphiquement les solutions trouvées et juger ainsi de leur unicité ou non. Avec FMNEAR, l'unicité de la solution est évaluée au travers d'un indice de confiance, qui sera élevé s'il n'existe pas de solutions ayant une RMS proche de celle de la meilleure solution avec des mécanismes significativement différents, et qui sera faible dans le cas contraire. Nous renvoyons à Delouis (2014) pour plus de détail.

Avec les deux approches, un modèle de vitesse 1D propre à chaque station peut être utilisé, défini par l'utilisateur, mais la Méthode 1 incorpore une recherche automatique d'un modèle de vitesse 1D optimisé par station, bien que cela rallonge considérablement le temps de calcul.

Pour les deux approches, le filtrage, donc la bande passante, de chaque composante de chaque station peut être défini par l'utilisateur. Toutefois, toujours dans l'optique du temps quasi réel, et pour un calcul automatisé, l'approche FMNEAR incorpore une recherche automatique du filtrage optimal de chaque composante. Cette recherche est basée sur l'analyse du spectre du

signal sismique en accélération, comparé au spectre du bruit de fond. Nous renvoyons de nouveau à Delouis (2014) pour plus de précisions sur cet aspect.

2.8. L'inversion multi données pour la distribution du glissement cosismique

2.8.1. Modèle cinématique de source étendue

Nous considérons ici le cas de séismes de forte magnitude enregistrés par de multiples données, dont certaines à faible distance, impliquant que les conditions d'utilisation du modèle de source ponctuelle ne sont plus vérifiées. Il est alors nécessaire de basculer sur un modèle de « source étendue », dit aussi de « dimension finie ». Il faut prendre en considération la distribution spatio-temporelle du glissement cosismique sur le plan de faille, donc répartir le mouvement sur une surface. Ce type de modèle est également appelé un modèle cinématique de la source. L'approche utilisée consiste à discrétiser la surface de rupture en une somme de petites surfaces, appelées « sous-failles », suffisamment petites pour que les ondes émises par celles-ci vérifient en revanche les conditions de source ponctuelle.

La rupture est ainsi représentée par une somme de sous-ruptures décalées dans le temps et dans l'espace. Pour la modélisation des ondes, le plan de faille global est remplacé par un ensemble fini de sous-failles, au centre desquelles est placée une source ponctuelle dont le déclenchement correspond au passage du front de rupture qui s'est propagé depuis l'hypocentre. Pour la modélisation des données géodésiques (déplacement statique du sol), chaque sous-faille est représentée par une petite surface de dislocation cisaillante.

La **Figure 2-9** illustre la représentation de la discrétisation d'un plan de faille en une somme finie de points sources.

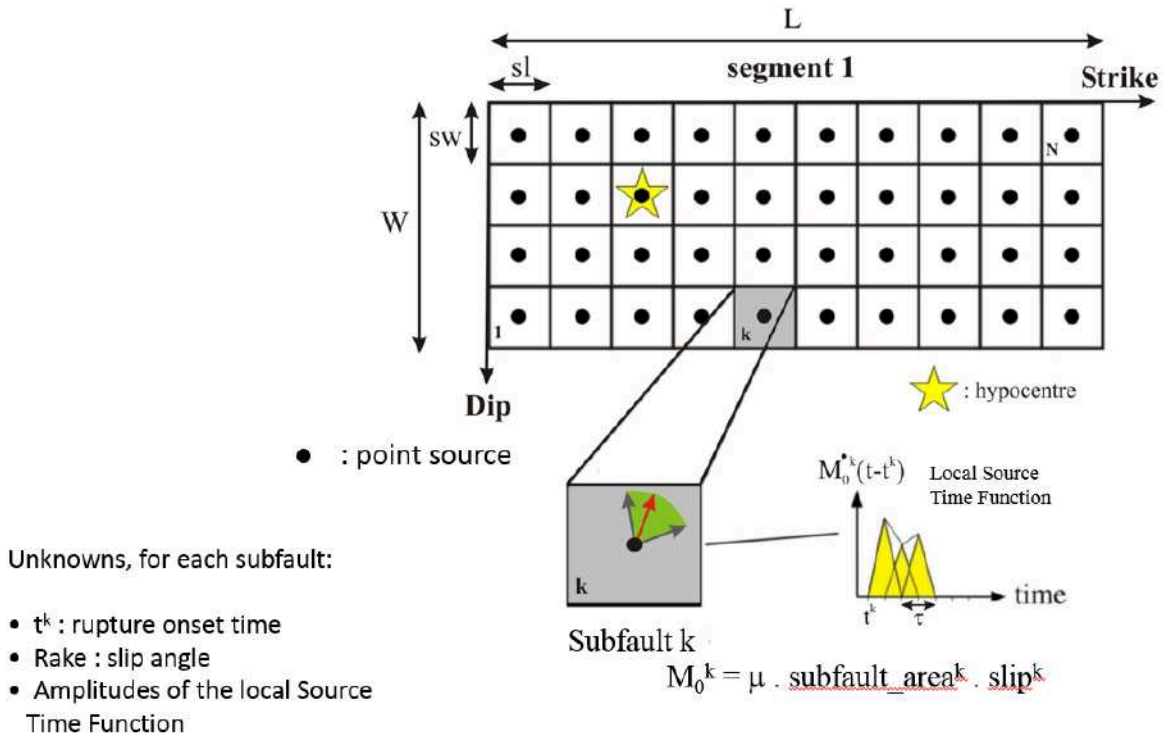

Figure 2-9: *Schéma de la discrétisation du plan de rupture dans le cas d'une source étendue. Illustration pour un seul segment de rupture, caractérisé par ses angles strike et dip (azimut et pendage), ses longueur et largeur totales, L et W. Les sous-failles peuvent être rectangulaires, de longueur sl et de largeur sw. Les points noirs représentent les points sources au centre de chaque sous-faille. L'un d'eux coïncide avec l'hypocentre (étoile jaune). Sur une sous-faille donnée (numéro k), le vecteur glissement (vecteur rouge) peut varier à l'intérieur d'un certain domaine (arc vert), en relation avec les valeurs permises pour l'angle de rake. Une fonction temporelle de la source locale (STF locale, fonction taux de moment sismique) est associée à chaque sous-faille, avec un temps de déclenchement t^k, représentée par une somme de triangles isocèles de largeur τ et se recouvrant mutuellement pour moitié. Subfault_area correspond à une surface sl.sw, et slip au glissement sur une sous-faille, considéré constant (Delouis, 2009).*

Au niveau de chaque sous-faille, la fonction temporelle de la source est représentée par plusieurs fonctions triangulaires et isocèles (**Figure 2-9**), on parle alors de fenêtres temporelles multiples, pour permettre la description du taux de moment sismique au cours du temps, sur une sous-faille. L'aire sous cette fonction correspond au moment sismique M_0^k de la sous-faille. La rigidité du milieu (unique, variable, au choix) et la surface des sous-faille étant fixées, il est alors possible d'obtenir le glissement Δ_u^k sur une sous-faille :

$$\Delta_u^k = \frac{M_0^k}{\mu^k . S^k} \quad (5)$$

2.8.2. Détail de la méthode d'inversion

La procédure d'inversion que nous avons utilisée est celle décrite dans Delouis et al. (2002). Elle repose sur la formulation du modèle de source étendue décrite ci-dessous, qui peut être multi-segments, c'est-à-dire incorporer plusieurs segments de failles, chacun défini par ses angles de strike, dip, rake (le rake pourra varier), et ses dimensions L et W.

2.8.3. Les paramètres à inverser

- le temps de déclenchement de la rupture « t^k » **(Figure 2-9)**

Tout d'abord on définit la notion de vitesse de rupture moyenne entre l'hypocentre et une sous-faille donnée (Vr), comme étant égale à la distance entre le centre de la sous-faille et l'hypocentre, notée « dhypo », divisée par le temps de déclenchement de la sous-faille. On a alors Vr= dhypo/t^k. L'intervalle de temps permis pour le déclenchement d'une sous-faille donnée est borné entre « tmin » et « tmax », associés à deux vitesses de rupture moyenne minimale et maximale. Soit :

$$t_{max} = \frac{d_{hypo}}{Vr_{min}} \text{ et } t_{min} = \frac{d_{hypo}}{Vr_{max}} \quad (6)$$

où « t_{max} » et « t_{min} » sont les temps de déclenchement maximum et minimum de la rupture, « Vr_{max} » et « Vr_{min} » les valeurs maximum et minimum permises pour la vitesse de rupture moyenne dans l'inversion. La vitesse de rupture moyenne peut être d'abord autorisée dans un intervalle de valeurs assez large (par exemple de 1km/s à 3.5 km/s). Il sera réduit par des tests intermédiaires afin d'optimiser l'inversion sans dégrader la modélisation des données.

- *Le rake*

Contrairement au strike et au dip, le rake est un paramètre inversé, afin de permettre des variations de la direction du glissement le long de la rupture. Les valeurs limites sont définies par un degré de liberté autour de la valeur de l'angle définie dans le mécanisme au foyer.

- *l'amplitude de chaque fonction triangulaire*

L'amplitude maximale commune pour chaque fonction triangulaire de l'ensemble des sous-failles du modèle, est déterminée par rapport à la magnitude de l'évènement étudié. La valeur minimale est nulle (ou correspondante à un glissement négligeable). L'amplitude de chaque fonction triangulaire isocèle sera inversée dans cet intervalle de valeurs.

2.8.4. Déroulement de la procédure d'inversion

La méthode d'inversion utilisée repose sur une exploration itérative de l'espace des solutions possibles pour chaque paramètre à inverser. L'exploration, basée sur des tirages aléatoires des valeurs des paramètres entre leurs bornes minimales et maximales permises, est contrôlée par un algorithme de recuit simulé adapté de Corona et al. (1987) et Goffe et al. (1994). Les paramètres à inverser seront donc :

- le temps de déclenchement ;
- le rake ;
- les amplitudes des fonctions sources triangulaires sur chaque sous-faille.

Le nombre total « $Ntot_{param}$ » de paramètres à inverser sera fonction du nombre de fonctions triangulaires isocèles « N_{tau} » allouées à chaque sous-faille, et du nombre de sous-faille « N_{SF} » :

$$Ntot_{param} = (2 + N_{tau}) \times N_{SF} \quad (7)$$

2.8.5. Initiation de la procédure

L'inversion commence à partir d'un modèle de départ, que l'on choisit arbitrairement à l'intérieur des valeurs permises pour les paramètres, ceci avant la mise en place d'un tirage aléatoire dans la suite du calcul. Ce modèle initial permet donc de lancer la procédure d'inversion et de fournir une première fonction cout « Fcout » :

$$Fcout = \sum_{1}^{n_{datasets}} w_j \, RMS_j + bF_{M_0} + c \, F_{smooth} \quad (8)$$

où « $n_{datasets}$ » est le nombre de jeux de données distincts, « j » décrit le jeu de données (dataset) j, «w_j» est le poids attribué au jeu de données « j », «F_{M_0} » est une fonction de minimisation du moment sismique total pondéré par un coefficient « b », « F_{smooth}» est une fonction de lissage pondérée par un coefficient « c ».

La RMS de chaque jeu de données utilisé est définit comme suit :

$$RMS_j = \sqrt{\sum_1^n [\frac{(O_i - C_i)^2}{O_i^2}]} \quad (9)$$

où « n » est le nombre d'échantillon du jeu de données « j », «O_i» les valeurs observées, «C_i» les valeurs calculées de l'échantillon « i »

La fonction d'une minimisation du moment sismique est décrite par :

$$F_{M_0} = e^{[\left(\frac{M_{0_{mod}}}{M_{0_{ref}}}\right) - 1]} \quad (10)$$

où « $M_{0_{mod}}$ » est le moment sismique total du modèle testé, « $M_{0_{ref}}$ » le moment sismique de référence de l'évènement étudié. Cette fonction permet d'augmenter la fonction cout totale « Fcout » des modèles ayant un moment sismique calculé « $M_{0_{mod}}$ » supérieur au moment sismique de référence. La minimisation du moment sismique est nécessaire afin d'éviter que le modèle ne comporte des zones de glissement injustifiées ne contribuant pratiquement pas à l'amélioration de l'ajustement des données.

Dans la procédure d'inversion, la fonction de lissage (sur le glissement et/ou sur la vitesse de rupture) pour une sous-faille donnée est proportionnelle à la racine carrée de la somme des écarts quadratiques entre les valeurs du paramètre à lisser sur l'ensemble des sous-failles voisines (au sens rms donc). Ce lissage peut permettre d'éviter des variations brutales des propriétés de la rupture d'une sous-faille à l'autre.

Le modèle de départ a une influence réduite sur le résultat final, si les paramètres de convergence du recuit simulé (décrit plus loin) sont bien réglés. Il correspond juste à un point de départ pour l'exploration. Il est toutefois nécessaire pour fournir une première fonction cout à laquelle sera comparée celle du premier modèle tiré aléatoirement.

2.9. Exploration basée sur l'approche recuit simulé (simulated annealing)

Le recuit simulé est une approche d'exploration itérative de l'espace des solutions inspirée des cycles thermiques utilisés en métallurgie. Un paramètre de température est introduit, qui permet de contrôler la convergence vers un état stable, censé être le minimum principal de la fonction cout. Cette approche fait partie des méthodes dites de Monte-Carlo, reposant sur la génération de nombres aléatoires. Au début du processus, la température est élevée et des

solutions moins bonnes que celles trouvées précédemment peuvent être acceptées. Cela permet à l'exploration de sortir de minima secondaires. Vers la fin du processus, la température a fortement diminué et la recherche se concentre sur l'exploration du meilleur minimum trouvé, qui est en principe le minimum principal. Quel que soit le paramètre testé, le modèle construit à chaque itération sert de base pour générer les données synthétiques calculées qui entrent dans la définition de la fonction cout. L'objectif est de minimiser la fonction cout.

La méthode repose sur trois boucles imbriquées. La boucle la plus externe définit le nombre de cycles à température constante. La boucle intermédiaire itère sur les différents paramètres à inverser. La boucle la plus interne définit le nombre de fois qu'un paramètre donné est tiré aléatoirement avant de passer au paramètre suivant. A chaque itération, le tirage aléatoire du paramètre courant donne lieu au calcul d'une nouvelle fonction cout (Fcout) (25), tous les autres paramètres étant maintenus inchangés. Cette nouvelle fonction cout est comparée à celle de l'itération précédente. Si la fonction cout a diminué, la valeur du paramètre courant est conservée, sinon on procède au test de Métropolis (voir ci-dessous). Si le test de Métropolis est vérifié, la solution est conservée, sinon elle est rejetée et on revient à la valeur précédente (celle de l'itération précédente). Lorsque les trois boucles imbriquées ont été parcourues, on test si la précision voulue pour arrêt de la procédure est atteinte (convergence vérifiée). Si oui, c'est fini. Si non, on réduit à la fois la température et les intervalles de recherche pour les paramètres, on se replace sur la meilleure solution trouvée depuis le début de l'exploration, et on repart de nouveau sur les trois boucles imbriquées. Ainsi de suite jusqu'à convergence et arrêt.

2.10. Critère de Métropolis

Toute solution induisant une valeur de fonction cout supérieure à la valeur obtenue à l'itération précédente ne sera pas nécessairement rejetée. Le choix d'acceptation ou de rejet d'une telle solution va se faire au travers du critère de Métropolis.

La valeur de la fonction cout à une itération « i » est comparée à la valeur de la fonction cout obtenue à l'itération « i-1 ». Le principe est de rejeter quasiment systématiquement les solutions qui induisent une fonction cout largement supérieure par rapport à l'itération précédente, tout en accordant une forte probabilité de conserver celles pour lesquelles la fonction cout n'est que légèrement supérieure à la précédente.

La probabilité que cette fonction cout soit retenue est :

$$P = e^{\left(\frac{-(f_i - f_{i-1})}{T}\right)} \quad (11)$$

où «f_i» est la fonction cout à l'itération courante, «f_{i-1}» la fonction cout à l'itération précédente, « T » est le paramètre dit de température.

Un tirage aléatoire d'une valeur entre 0 et 1 est réalisé. Si ce nombre est inférieur à la valeur de P calculée, la solution est retenue. Plus la différence entre les fonctions cout «f_i» et «f_{i-1}» est faible, plus la différence « $f_i - f_{i-1}$ » tend vers 0, donc l'exponentielle tend vers 1. Il y a donc une forte probabilité que la solution soit retenue. Plus la différence est grande, plus la probabilité que la solution soit rejetée est grande. On voit également que l'acceptation est plus probable à température élevée (grande valeur de T) qu'à basse température (petite valeur de T).

Un des paramètres de contrôle de l'inversion (à fixer) est le facteur de réduction de la température lorsque les trois boucles imbriquées décrites ci-dessus ont été parcourues. En fonction de ce facteur, on peut réduire ou augmenter le temps de calcul. Une diminution trop rapide de la température peut induire une convergence rapide mais piégée dans un minimum secondaire. Une diminution trop lente de la température augmenterait le temps de calcul sans pour autant permettre d'améliorer l'optimisation de la fonction cout.

2.11. Arrêt de la procédure

La condition pour arrêter le processus est d'atteindre un critère de convergence, à savoir une précision sur la valeur des paramètres déterminée en amont de l'inversion. Cette précision est atteinte lorsque l'intervalle de recherche devient inférieur à la valeur de précision pour l'ensemble des paramètres inversés. Tant qu'au moins un intervalle défini pour un paramètre est supérieur à cette précision, la procédure d'inversion continue. Lorsque cette précision est atteinte, l'inversion est stoppée, et l'ensemble des valeurs ayant fourni la plus petite fonction cout est considérée comme la meilleure solution.

Chapitre 3 : Analyse de séismes modérés récents en Algérie

Introduction

Les mécanismes au foyer par inversion des formes d'ondes présentés dans cette thèse ont été calculés en utilisant deux approches, l'une manuelle et l'autre automatique. Les deux ont été développées par Delouis et Legrand (1999) modifiée par Orgulu et al., (2005) et ensuite Delouis (2014). Avec l'approche manuelle, le traitement des signaux sismiques se fait manuellement où la principale tache est de trouver la meilleure bande passante (le bon filtrage). Une fois la bonne bande de filtrage trouvée, il faut chercher les meilleures solutions du mécanisme au foyer avec le meilleur modèle de vitesse et cela pour chaque station qui est utilisée dans l'inversion. La démarche est de commencer par un modèle de vitesse en ½ espace et ensuite faire des recherches pour des modèles à couches. La source sismique du séisme de Tadjena (16/12/2006) a été calculée en utilisant cette première approche. Pour les séismes qui sont venus après (Beni-Ilmane et les autres), une version automatique et améliorée a vu le jour (FMNEAR, Delouis, 2014) où toute la partie filtrage, lecture des arrivées des ondes P se fait automatiquement la première fois, et c'est en deuxième étape après avoir une première solution automatique que devons réviser les pointés, les filtrages et tout les autres paramètres dans le cas ou la solution n'est pas robuste (Index de Confidence CI < 60%).

Dans ce chapitre nous allons présenter huit mécanismes aux foyers **(Figure 3-1)** obtenus par inversion des formes d'ondes sismiques de séismes modérés qui se sont produits en Algérie (Tell Atlas, la région des Bibans et la transition Atlas saharien-plateforme saharienne), six d'entre eux ont fait l'objet de publications dans les revues Pure and Applied geophysics (Pageoph), Tectonophysics et Seismological research Letters (SRL) dont je suis soit le premier auteur, soit co-auteur. Dans tous les cas, c'est moi qui ai réalisé les inversions des formes d'onde.

- Le premier séisme, de Tadjena, s'est produit dans la région du Cheliff sur une faille (Boukadir) située à l'Ouest de la faille d'El Asnam où s'est produit le séisme du 10 octobre 1980 (Ms = 7.3). La faille de Boukadir est de type inverse et a les mêmes

caractéristiques que celle d'El Asnam à savoir son orientation, sa dimension, ainsi que son pendage qui plonge vers le Nord. Ce résultat a été publié dans Pageoph, Beldjoudi et al. (2011).

- Le deuxième séisme est en fait une séquence sismique de trois séismes modérés de magnitudes de moment respectives (Mw = 5.5, 5.1 et 5.2). Cette séquence sismique s'est produite durant le mois de mai 2010 dans la région des Bibans à l'Est de l'Algérie dans le secteur de Beni-Ilmane. Elle est localisée à l'intérieur d'un pli actif, où nous avons calculé un premier mécanisme en décrochement, un deuxième en faille inverse et un troisième en décrochement aussi. Ces résultats ont été publiés dans Tectonophysics, Beldjoudi et al. (2016).

- Un troisième séisme modéré de magnitude (Mw = 4.1) s'est produit le 22 février 2014, il est localisé au sud de la région de Boumerdes-Zemmouri, dans la localité de Bordj Menaïel. L'inversion des formes d'ondes a donné un mécanisme en décrochement. Ce travail a été publié dans la revue SRL, Semmane et al. (2015).

- Un quatrième séisme modéré (Mw = 5.0) s'est produit sur la bordure Sud de la Mitidja le 17 juillet 2013. Ce séisme fait partie d'une série de séismes modérés qui ont secoué la région de Hammam Melouane. C'est une faille en décrochement qui a joué lors de ce séisme et un papier a été publié en 2017 dans la revue Pageoph, Yelles-Chaouche et al. (2017).

- Un dernier séisme que nous allons présenter dans ce chapitre est celui de la localité de M'ziraa (Est de la ville de Biskra, au Sud d'Algérie). Ce séisme s'est produit le 18 novembre 2016, de magnitude Mw = 5.1, avec un mécanisme au foyer en faille normale orientée E-O, et à une profondeur de 36 km localisé à la transition Atlas saharien et plate forme saharienne. Il a été suivi par une réplique de Mw = 4.7. Ce séisme nécessite une attention particulière et une investigation plus approfondie afin de comprendre ce qui s'est passé, car c'est l'un des rares cas de séisme en faille normale en Algérie. Ce travail n'est pas encore publié.

Nous incluons les deux papiers dont je suis le premier auteur dans ce chapitre (Tadjena et Beni-Ilmane) et nous présentons brièvement le résultat de l'inversion des formes d'ondes pour les autres.

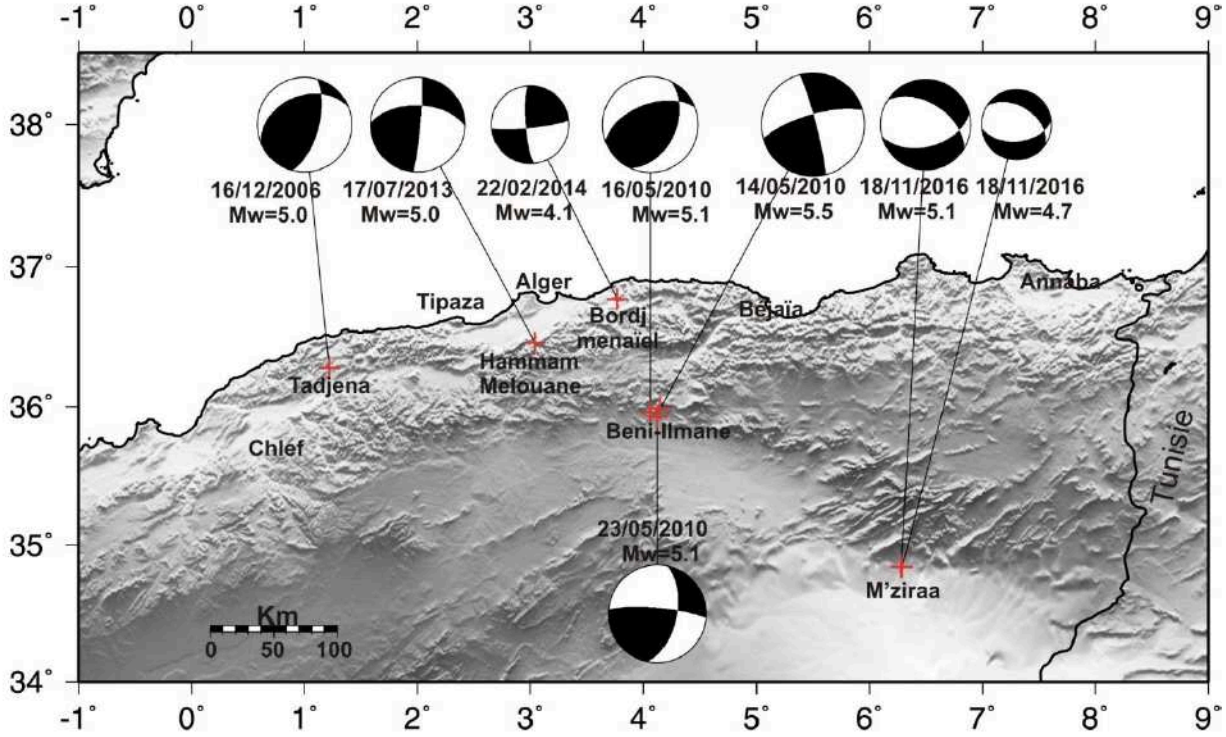

Figure 3-1 : *Carte des mécanismes au foyer des séismes obtenus par inversion des formes d'ondes sismiques. Les croix rouges montrent les épicentres de chaque séisme.*

3.1. The Tadjena Earthquake (Mw = 5.0) of December 16, 2006 in the Cheliff Region (Northern Algeria): Waveform Modelling, Regional Stresses, and Relation with the Boukadir Fault (H. Beldjoudi, B. Delouis, A. Heddar, O. B. Nouar And A. Yelles-Chaouche)

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Pure and Applied Geophysics

The Tadjena Earthquake (Mw = 5.0) of December 16, 2006 in the Cheliff Region (Northern Algeria): Waveform Modelling, Regional Stresses, and Relation with the Boukadir Fault

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Abstract—The Cheliff region has experienced some significant earthquakes in the last century (1937, 1954, and 1980). The most destructive one is that of El Asnam on October 10, 1980, Ms = 7.3 (Io = IX), which destroyed the Chlef city (formerly El Asnam) and its surrounding villages. On December 16, 2006 a moderate earthquake (Mw = 5.0) hit the Cheliff region. The maximum observed intensity (Io = V: MSK-scale) was observed at Abou El Hassen, Benaria, Bouzghaïa and Tadjena. No damages or human losses were recorded. Nevertheless, minor cracks on walls of the old school at Tadjena were observed. The point source focal mechanism of the event was determined by inverting the waveforms of three regional broadband stations of the ADSN (Algerian Digital Seismic Network). It corresponds to thrust-reverse faulting with a strike-slip component. The stress tensor obtained by the inversion of the 15 focal mechanisms available in the Cheliff region exhibits a well constrained compression axis $\sigma 1$ horizontal and trending N145°. The NW dipping nodal plane indicating a NE–SW thrust fault with a right-lateral component (strike, dip, rake = 249, 38, 137) is more compatible with the regional stress tensor than the steep dipping NNE-SSW nodal plane showing reverse faulting with a left-lateral component (strike, dip, rake = 15, 65, 60). Accordingly, the Tadjena moderate size earthquake can be related to the Boukadir active fault bordering the lower Cheliff basin to the north, a situation similar to that of the El Asnam fault bordering the middle Cheliff basin to the north.

Key words: Tellian Atlas, waveform modelling, source parameters, Algeria, Cheliff seismicity, focal mechanism, regional stresses.

1. Introduction

Northern Algeria lies along the Eurasian–African plate boundary and can be affected by strong shallow earthquakes as demonstrated by the events of Algiers in 1365 and 1716 (Io = IX-X) (Mokrane *et al.*, 1994), Orléanville in 1954 (Ms = 6.7) (Dewey, 1990), El Asnam (formerly Orléanville) on October 10, 1980 (Ms = 7.3) (Ouyed *et al.*, 1981, 1983, Benhallou, 1985; Yielding *et al.*, 1989) and Boumerdes on May 21, 2003 (Mw = 6.8) (Ayadi *et al.*, 2003, 2008; Belabbes *et al.*, 2009; Bounif *et al.*, 2004; Delouis *et al.*, 2004; Meghraoui *et al.*, 2004; Semmane *et al.*, 2005; Yelles-Chaouche *et al.*, 2004). The CRAAG (Centre de Recherche en Astronomie Astrophysique et Géophysique) recently installed (2006–2007) a new digital seismic network (Algerian Digital Seismic Network: ADSN) composed of eight broadband (BBVS-60: http://www.geodevice.cn) and two very broadband (GCZ-NT: Institute of seismology China Earthquake Agency. Xiao Hongshan, Wuhan, China, 430071) seismic stations (Fig. 1). The main objectives of this ADSN network are to improve the data quality, to implement a real time alert system (Yelles-Chaouche *et al.*, 2007), and to give the possibility to characterize the seismic sources. On December 16, 2006 at 19h34mn (UTC), the Cheliff region was struck by a shallow moderate-size earthquake, hereafter called the Tadjena earthquake. The seismic moment (Mo) and moment magnitude (Mw) were respectively 3.5×10^{16} Nm and 5.0 (This study). In Table 1 and Fig. 1 we display the different locations of the main shock given by different seismologic centres. From the fault-plane solution provided by IGN (Instituto Geográfico National, Spain, http://www.ign.es/ign/es/IGN/home.jsp), and this study, the earthquake occurred by a thrust-reverse fault oriented NE–SW to NNE–SSW with a strike-slip component like the seismic sources in the Cheliff region (thrust faulting with a left-lateral

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component for the 1980 event, and thrust faulting with a right-lateral component for the 1954 event). The earthquake provoked a great panic among the population of Bouzeghaïa, Tadjena, Abou El Hassen, Chlef and Tenes. Fortunately, no damages or human losses were reported.

The Tadjena earthquake is located in a very active region of the northwestern part of Algeria, precisely in the vicinity of the Boukadir fault which is assumed to be an active fault (Meghraoui, 1982, 1988) sharing many similarities with the El Asman fault responsible for the largest known earthquake in northern Algeria (Fig. 2). From surface observations, both faults appear as reverse faults associated with folding, dipping to the northwest, with a length of several tens of km. The question whether the Boukadir fault could produce a large earthquake similar to the El Asnam one is raised, and our seismological study confirms the seismic and reverse character of the Boukadir fault at depth. The Cheliff region became the center of interest of seismologists after the El Asnam earthquake (1980); many studies were driven by different authors and teams. Ouyed *et al.,* (1981, 1983) and Yielding *et al.,* (1981 and 1989) studied the El Asnam aftershocks sequence. Meghraoui (1982, 1988) conducted seismotectonic and paleoseismology studies in the region. Aoudia and Meghraoui (1995) analyzed the earthquake of Abou El Hassen (1922) located on the Tenes-Abou El Hassen fault in the north of the Cheliff region (Fig. 2). Bezzeghoud *et al.,* (1994, 1996) analyzed the seismicity of the El Asnam region from the data of the Algerian telemetred seimological network and determined the focal mechanism of the Rouina earthquake in the eastern part of the Cheliff region (event 14, Fig. 2) using P-wave first motions. In this paper, we present, for the first time in Algeria, a focal mechanism obtained by modelling regional waveforms using records of three broadband stations of the ADSN network, a methodology which had never been applied in Algeria. This event, together with 14 focal mechanisms previously determined for the Cheliff region, is used to determine the regional stress tensor. The formal inversion of the stress tensor had

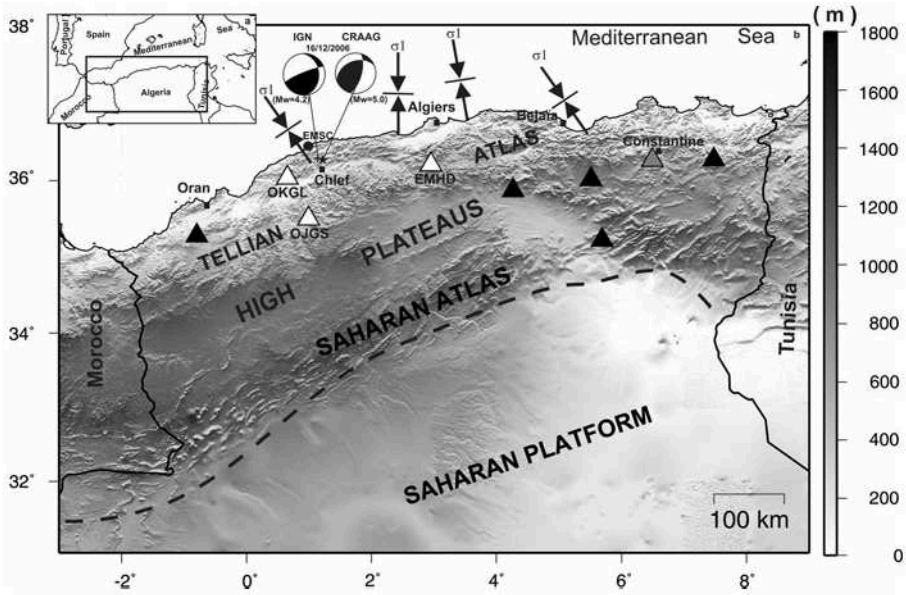

Figure 1

a Map shows regional location of the north of Algeria. **b** Shaded relief (from etopo5; http://www.ngdc.noaa.gov) of northern Algeria showing the different morphostructural units. Algiers, Oran and Constantine are the principal megacities in Algeria. *Black triangles* are broadband seismic stations. *White triangles* are broadband seismic stations used in the waveform modelling. *Grey triangle* is the very broadband seismic station. The star is the CRAAG location of the December 16, 2006 Tadjena event and its focal mechanism obtained in this study. The cross is the IGN location of the December 16, 2006 and its calculated focal mechanism. The *open circle* is the EMSC location of the same event. *Black arrows* pointing inward are the projections onto the horizontal plane of compressive axes σ_1 obtained by different authors. In the eastern part of Algeria σ_1 is N145°E ± 15° (Beldjoudi *et al.,* 2009). In the central part of Algeria, σ_1 is N170°E (Stich *et al.,* 2006; Ayadi *et al.,* 2008). In the western part the σ_1 is N145°E ± 15° (calculated in this study)

Table 1

Earthquake location of December 16, 2006 event as given by different seismological centres

Origin time UTC (HH:MM:SS)	Lat.° N	Lon.° E	Depth (km)	Magnitude	Ref.
19:34:33.0	36.28	1.22	7.00	Md = 5.2	CRAAG
19:34:31.0	36.29	1.15	10.50	Mw = 4.2	IGN
19:34:34.3	36.45	0.91	30.00	mb = 4.8	CSEM

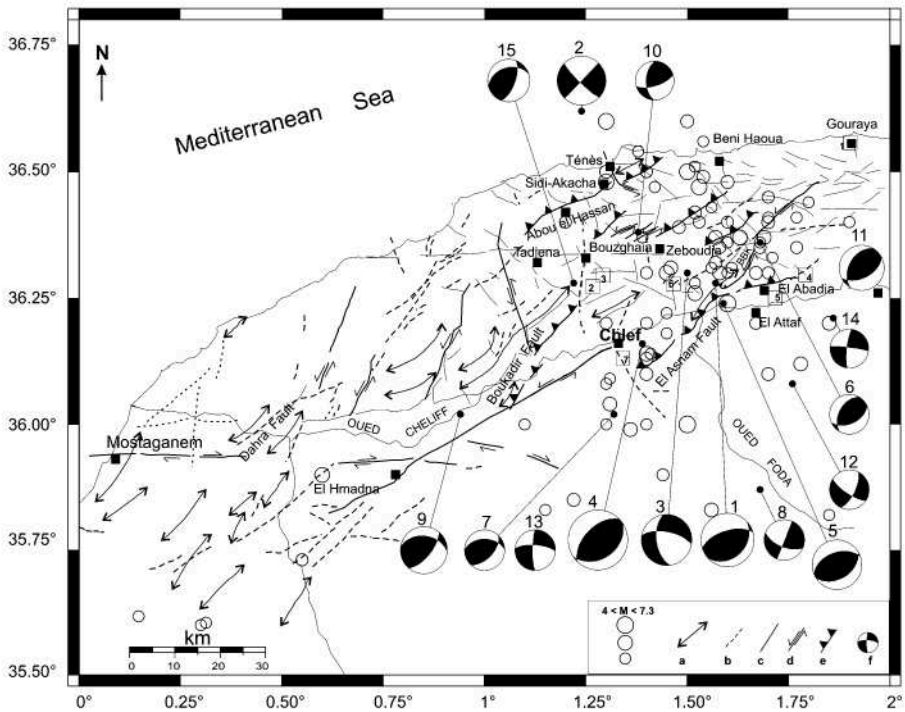

Figure 2

Seismotectonic map of the Cheliff region [after MEGHRAOUI (1988) modified]. Focal mechanism (1 to 15) parameters are detailed in Table 2. **a** Anticline. **b** Probable Faults. **c** Probable Faults (Satellite images and DEM). **d** Strike-Slip Faults. **e** Thrust Faults. **f** Focal Mechanism. *Open squares* 1 to 7 correspond to epicentres cited in the Table 3. *Open circles* show the seismicity (historical and instrumental) recorded from 1365 to 2009 ($4.0 < M < 7.3$)

never been performed for this region of Algeria, previous studies relying only on the analysis of the P- and T-axes of the focal mechanisms (OUYED *et al.*, 1983; LAMMALI *et al.*, 1997). By carrying out the stress inversion we aim at determining precisely the maximum compression direction in western central Algeria and compare it to previous determinations done for areas more to the east (BOUNIF *et al.*, 2003; STICH *et al.*, 2006; AYADI *et al.*, 2008; BELDJOUDI *et al.*, 2009). The present-day plate boundary between Africa and Eurasia being considered to be located in northern Algeria, with a change in plate motions in the last few million years (CALAIS *et al.*, 2003), it is of great importance to determine the shortening direction as seen by earthquakes and its possible lateral variations.

Finally, the relation between the Tadjena event and the active faults in the epicentral area is discussed.

2. *Seismotectonic Setting of the Cheliff Region*

Northern Algeria is formed by four morpho-structural domains, namely: The Tell Atlas, the High Plateaus, the Saharan Atlas and the northern part of

the Saharan platform (Fig. 1). The Tell Atlas is the northernmost part of the African (Nubian) plate, and its tectonic activity expresses the convergence between the African and European plates with a shortening rate of ~5 mm/year (Nocquet and Calais, 2004; Stich *et al.*, 2006). It is characterized by predominantly reverse faulting under ~NNW–SSE to NW–SE compression (Negredo *et al.*, 2002; Calais *et al.*, 2003; Fernandes *et al.*, 2003; Henares *et al.* 2003; McClusky *et al.*, 2003; Nocquet and Calais, 2004; Stich *et al.*, 2006). It consists of a succession of mountain ranges and valleys sub-parallel to the coastline with juxtaposed platforms (alluvial basins) and high topography reaching 1,500 m.

The Cheliff valley is a section of the Tell Atlas domain where compressional tectonic structures are predominant (Meghraoui, 1988; Meghraoui *et al.* 1996; Yelles-Chaouche *et al.*, 2006). The neotectonic structures are represented by a series of northeast-southwest trending folds and associated parallel thrust faults with a NNW–SSE compressional direction (Meghraoui, 1988). This region is characterized by shallow seismicity and active faulting (Fig. 2).

The Cheliff basin is one of the well-known and the most studied areas in the Tell Atlas. It was a site of destructive and large earthquakes such as the 1954 Orléansville Ms = 6.7 (Dewey, 1990) and the 1980 El Asnam, Ms = 7.3 earthquake (Ouyed *et al.*, 1981, 1983; Ruegg *et al.*, 1982; Meghraoui *et al.*, 1988; Yielding *et al.*, 1989; Aoudia *et al.*, 2000). The Cheliff basin includes the El Asnam (or Oued Fodda) fault and related fold. This is a reverse fault oriented NE–SW dipping 50°–60° toward the northwest (Fig. 2). After the 1980 earthquake, 30 km of surface breaks were observed (Ouyed *et al.*, 1981, 1983; Ruegg *et al.*, 1982). From paleoseismological studies a shortening value of 2.2 mm/year was calculated at the eastern part of the Cheliff basin (Meghraoui *et al.*, 1996). Near the northeastern termination of the El Asnam fault, the Bled Bahari Karouche (BBK) fault has been described as an active structure displaying left-lateral motion and thrusting at its southern extremity (Yelles-Chaouche *et al.*, 2006).

The Tenes-Abou El Hassen active fold is located 40 km to the northwest of the Cheliff basin (Fig. 2). The fold has an approximate visible total length of 23 km and an average width of 5 km. A comparison with the El Asnam fold indicates several similarities: (1) The length (about 30 km), the (2) strike (NE–SW) and (3) the asymmetry of folds with a transport direction to the southeast (Aoudia and Meghraoui, 1995). The associated fault is NE–SW oriented and dipping toward the northwest.

The Boukadir active fold 30 km long located between the El Asnam structure at the east and the Tenes-Abou El Hassen at the north (Fig. 2), is associated with a N065 trending thrust fault dipping to the northwest (Aoudia *et al.*, 2000). The fault is composed by three segements: (1) a NE segment with a length of 15 km, (2) a central segment 4 km long and (3) a SW segment with a length of 11 km (Meghraoui, 1988).

After Yelles-Chaouche *et al.*, (2006), the Tenes off-shore area is an active zone where an active structure was identified and moderate seismicity is recorded.

3. *Historical Seismicity*

Referring to the historical seismicity of the Cheliff region, several moderate to large earthquakes struck the region with intensities of VIII to X (Fig. 2; Tables 2, 3). The maximum intensity observed in the region is (X–XI) (Bezzeghoud *et al.*, 1996; Ayadi *et al.*, 2002; Boughacha *et al.*, 2004). On August 25, 1922 a destructive earthquake struck the region, the macroseismic epicentre was localized near the Cavaignac (Abou El Hassen) village which was totally ruined (Fig. 2). An intensity of VIII was assigned by Hée (1923, 1933) for the villages of Flatters (Benaria), Hanoteau (Zeboudja), Montenote (Sidi Akacha) and Tenes (Mokrane *et al.*, 1994; Aoudia and Meghraoui, 1995). Aoudia and Meghraoui (1995) associate the earthquake with the central segment of the Tenes-Abou El Hassen fault.

Another destructive earthquake with Io = IX struck the region on September 7, 1934; the earthquake destroyed the localities of Carnot (El Abadia), Saint-Cyprien (Sidi-Bouabida) and Les Attafs (El Attaf). Landslides were observed in the mountains at

Table 2

Seismic events reported during the 20th century for the Cheliff region

N°	Date	Lat.° N	Lon.° E	Depth (km)	M	Focal Meca Strike Dip Rake	Ref.
1	Sept. 9, 1954	36.28	1.57	7	6.5	253 61 104	ES
2	Sept. 10, 1954	36.62	1.24	20	6.0	44 90 −8	DW
3	June 5, 1965	36.30	1.50	10	6.2	172 56 −32	SH
4	Oct. 10, 1980	36.16	1.39	10.1	7.3	225 54 83	DG
5	Oct. 10, 1980	36.24	1.59	10.0	6.1	58 43 81	HRV
6	Oct. 30, 1980	36.36	1.68	7	4.8	210 46 64	CD
7	Nov. 8, 1980	36.02	1.32	15	5.0	270 45 126	HRV
8	Dec. 5, 1980	35.87	1.68	15	5.0	112 61 −179	HRV
9	Dec. 7, 1980	36.02	0.94	25.8	5.8	277 40 140	HRV
10	Jan. 15, 1981	36.38	1.38	15.0	4.7	181 53 29	HRV
11	Feb. 1, 1981	36.27	1.90	15.0	5.5	210 43 64	HRV
12	Feb. 14, 1981	36.08	1.76	15.0	4.9	26 67 −18	HRV
13	Nov. 15, 1982	35.73	1.15	15.0	5.0	274 70 −169	HRV
14	Jan. 19, 1992	36.21	1.86	8.5	4.7	277 85 −169	BA
15	Dec. 16, 2006	36.28	1.22	7	5.0	15 65 60	This study

ES: Espinoza and Lopez-Arroyo, 1984; DW: Dewey, 1990; SH: Shirokova, 1967; DG: DeSchampset al., 1982; CD: Cisternas and Gaulon, 1982; HRV: Harvard CMT solutions, now from GCMT, Global Centroid Moment Tensor catalog, http://www.globalcmt.org/CMTsearch.html. BA: Bezzeghoud, 1999; Bezzeghoud *et al.*, 1994. Depths are given by Harvard (GCMT) except for numbers 1, 2, 3, 6 and 14 which are given by CRAAG catalogue (Mokrane *et al.*, 1994)

Table 3

Major earthquakes of the Cheliff region observed between 1891 and 1980

No	Date	Time	Lat.° N	Lon.° E	M	Io	Observations	Ref.
1	Oct. 15, 1891	04 00 00	36.50	1.80		X	Gouraya Région	RT (1950)
2	Aug. 25, 1922	11 47 00	36.28	1.27	5.9	VIII	Abou El Hassen	AD and ME (1995)
3	Nov. 19, 1922	17 04 00	36.30	1.30		VIII	Abou El Hassen	RT (1950)
4	Dec. 17, 1926	05 41 00	36.30	1.80		VI	El Abadia	RT 1950) MK *et al.*, (1994)
5	Sept. 7, 1934	03 39 00	36.25	1.72	5.0	IX	El Abadia,El Attaf	MK *et al.* (1994)
6	Sept. 9, 1954	01 04 37	36.28	1.47	6.7	X–XI	Cheliff region	MK *et al.*, (1994)
7	Oct. 10, 1980	12 25 25	36.13	1.34	7.3	IX	Cheliff region	MK *et al.*, (1994)

RT: Rothé, 1950; AD: Aoudia and Meghraoui, 1995; MK: Mokrane *et al.* 1994; M: Magnitude; I_o: Isoseimal maximum intensity

Table 4

Characteristics of broadband stations used in this study

Stations BB	Longitude°	Latitude °	Δ (km)	Azimut °	Filter[a]	Sampling (Hz)[b]
OKGL	0.66 E	36.03 N	58	242	BP 0.03–0.15	4
OJGS	0.99 E	35.50 N	89	194	BP 0.05–0.15	4
EMHD	2.95 E	36.21 N	156	93	BP 0.05–0.15	4

BB broadband; Δ epicentral distance; *BP* band pass;

[a] Buttherworth filter, three poles

[b] After decimation

the north of Carnot (El Abadia). The earthquake was felt in Algiers located at ~135 km (Rothé 1950).

On September 9, 1954 a destructive earthquake (Ms = 6.5) again struck the Cheliff region; this earthquake caused the destruction of the Orléanville city and its surrounding villages. During this earthquake 1,340 people died (Mokrane *et al.* 1994).

Table 5

Focal mechanism parameters obtained in this study for the December 16, 2006 event

Sol.	Depth (km)	Strike1°	Dip1°	Rake1°	Strike2°	Dip2°	Rake2°	P axis		Taxis	
								Azimut°	Plunge°	Azimut°	Plunge°
S1	7	15	65	60	249	38	137	126	15	242	59

Sol. Solution

On October 10, 1980 the strongest earthquake (Ms = 7.3) hit the Cheliff region. This earthquake was the largest recorded in northwestern Africa. It caused 2,633 causalities and material damages. Public Buildings, houses and bridges were severely damaged. Railway was deformed and artificial lakes were formed and a vertical offset of 6 m was measured near Oued El Fodda village (Ouyed *et al.*, 1981, 1983; Yielding *et al.*, 1981, 1989; Ruegg *et al.*, 1982).

4. Location and Macroseismic Effects of the Earthquake of December 16, 2006

The seismic event of December 16, 2006 occurred at 19 h 34 min (GMT) had a moment magnitude Mw = 5.0. It was recorded by 22 stations of the telemetred Algerian network composed essentially of short-period seismic stations and ADSN network composed of eight broadband and two very broadband seismic stations. The location was made using 22 P- and 10 S-waves' arrival times. The closest station was at a distance of 14 km. The event of December 16, 2006 was located using the HYPOINVERSE program (Klein, 1978) based on a regional seismic velocity structure and using data from the local networks. The horizontal and the vertical errors were respectively ERH = 4 km and ERZ = 3 km. The residual time error RMS = 0.1 s.

The December 16, 2006 event produced no damages to buildings and socio-economic infrastructures (schools, hospitals, bridges, etc.). Great panic was observed at epicentral region of Abou El Hassen, Benaria, Bouzghaïa, and Tadjena (Fig. 3). At Tadjena village deep cracks on the walls of the school were observed. Figure 3 shows intensities attributed at each village obtained from data (questionnaires or inquires) collected after the event at the sites (black squares) as well as from field investigation. The event was felt within a radius of 50 km. The maximum intensity (Io = V, MSK-scale) is attributed at the localities of Abou El Hassen, Benaria, Tadjena and Bouzghaïa. In the Chlef city the event was felt by a major part of population, and the intensity attributed was IV.

5. Focal Mechanism and Source Parameters

5.1. Data Processing and Inversion Method

The data used for the determination of the focal mechanism of the mainshock correspond to the seismograms recorded by the ADSN. Displacement records are obtained by integration of velocity seismograms. They are bandpass-filtered between 0.03 or 0.05 and 0.15 Hz in order to remove the low frequency noise and to retain only the low-frequency part of the seismograms which contains the information needed to constrain the fault orientation and the slip direction but which is less affected by the crustal structure and rupture complexities. In this publication we use the records of three broadband stations (Fig. 1).

The inversion method is that of Delouis and Legrand (1999), modified in Orgulu *et al.*, (2005), and adapted to a point source. In the point source assumption, we invert for four parameters, the strike, dip and rake of the focal mechanism and the seismic moment.

The inversion consists of three major steps. The first step corresponds to a coarse search in the overall parameter space of the focal mechanism (strike, dip and rake), the synthetic waveforms being computed using an initial velocity model where the crust is represented by a simple half-space (Table 6). Each parameter in this step is evenly, but coarsely, sampled.

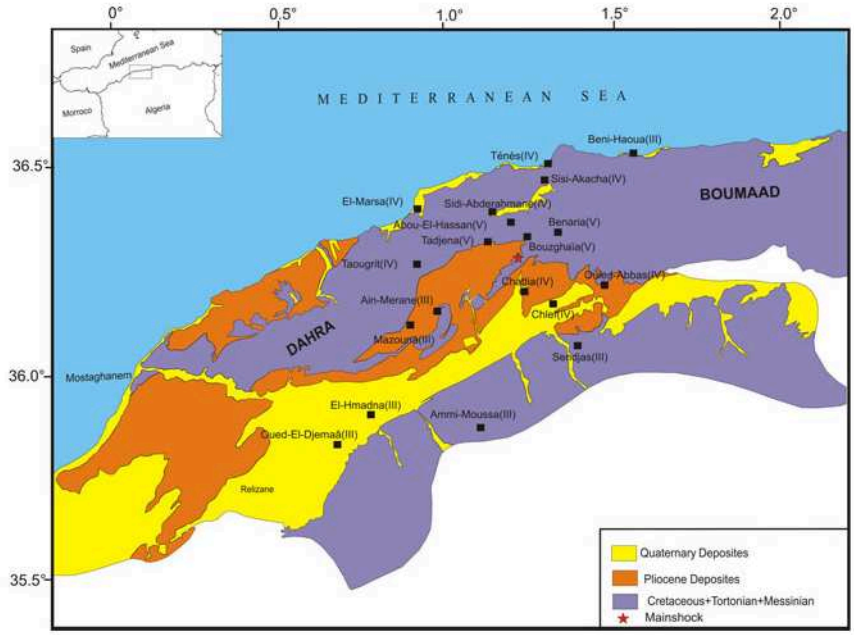

Figure 3
Simplified geological map showing the principal geological units [From MEGHRAOUI (1988), simplified]. The map shows cities and villages of the Cheliff region and their respective attributed intensities (MSK-scale) in relation with the Tadjena earthquake of December 16, 2006

Grid spacing is 20° for strike, 10° for dip, and 20° for rake, corresponding to 2,916 tested solutions.

The earthquake source time function (seismic moment rate) is represented by a single isosceles triangle, whose area corresponds to the seismic moment of the event. We invert the seismic moment in order to match the amplitudes of observed seismograms. The width of the triangular function is fixed and scaled with the a priori value of magnitude.

The criterion of selection in the grid search is the minimisation of the normalized RMS error between the observed and synthetic seismograms (misfit function)

$$\mathrm{RMS} = \frac{\sum_k \mathrm{rms}_k}{\mathrm{nstat}}$$

where nstat is the number of stations, Σ_k is the sum from ($k = 1$) to nstat, with

$$\mathrm{rms}_k = \sqrt{\frac{\sum_j (O_j - C_j)^2}{\sum_j (O_j)^2}}.$$

The Σ_j is the sum from $j = 1$ to N, N being total numbers of samples for the three components of

Table 6

Half space initial velocity model and velocity models found for stations OKGL, OJGS and EMHD respectively

Initial model					
Thicks (km)	V_p (km/s)	V_s (km/s)	Density	Q_p	Q_s
Half space	6.0	3.48	2.65	500	250
Velocity model of OKGL station					
2	2.40	1.26	1.60	120	60
8	5.60	2.93	2.60	470	235
20	7.00	3.66	3.10	700	350
Half space	8.10	4.68	3.30	1000	500
Velocity model of OJGS station					
3.00	3.60	2.21	2.10	240	120
8.00	4.80	2.95	2.35	370	185
19.00	5.80	3.56	2.65	500	250
Half space	8.10	4.68	3.30	1000	500
Velocity model of EMHD station					
0.60	2.00	1.17	1.40	100	50
3.00	4.00	2.35	2.15	270	135
8.00	5.60	3.29	2.60	470	235
18.40	5.80	3.40	2.65	500	250
Half space	8.10	4.68	3.30	1000	500

Thicks: layer thicknesses. V_p and V_s are velocity of P and S waves respectively. Q_p and Q_s are the factor quality of P and S waves respectively

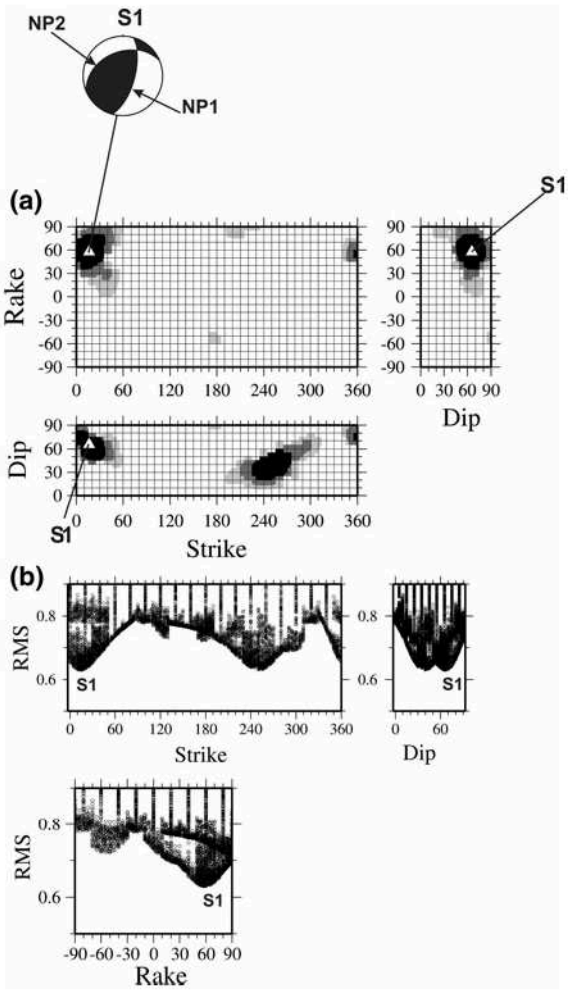

Figure 4
Result of point source inversion. **a** Solutions with the lowest rms errors projected on the (strike-rake), (strike-dip) and (dip-rake) planes. S1 shows the lowest rms (best solution). The focal mechanism corresponding to the best solution (S1) is drawn at the top. The two nodal planes are labelled NP1 and NP2. **b** rms errors plotted as a function of strike, dip and rake

seismograms. O_j and C_j are respectively observed and calculated signals. Synthetic seismograms are computed using the discrete wavenumber method of Bouchon (1981) for a 1D layered velocity model.

In a second step, we search to optimize the velocity model for each individual station using the best focal mechanisms found in the first step. Several thousands of combinations of layer thicknesses, velocities, and V_p/V_s ratios are tested. For each station, the layered model providing the best waveform match is retained (Table 6). Then the coarse search is redone with the new velocity models (one per station).

The third step consists of a fine search around the 45 minima of the RMS misfit function resulting from the coarse search. Those 45 minima correspond to the best solutions obtained for each discrete value of the strike, dip, and rake parameters tested in the coarse search. They sample a large part of the focal mechanism solution space and are not restricted to the vicinity of the very best solution.

In all steps of the inversion procedure, synthetic seismograms are computed for an arbitrary fixed initial value of seismic moment (initial_moment). The amplitude of synthetics and the seismic moment being linearly related, a loop on moment values is implemented within the RMS computation algorithm. Synthetic seismograms are simply multiplied by the ratio (tested_moment/initial_moment) before the RMS calculation, and we finally retain the value of moment (tested_moment) producing the lowest RMS misfit.

5.2. Results

The fault plane solution for the event was obtained using the point source inversion for three broadband station records of ADSN. The stations used are OKGL, OJGS and EMHD (Fig. 1; Table 4). The focal mechanism (FM) corresponds to a thrust or reverse faulting mechanism, depending on the nodal plane, with a strike-slip component. Nodal planes strike in a NE–SW or NNE–SSW direction. The horizontal pressure axis (P) oriented NW–SE (Table 5). The depth of the focus was fixed to that determined by the CRAAG using the Algerian network (7 km, Table 1). In Fig. 4a solutions with the lowest rms errors are projected on the (strike-rake), (strike-dip) and (dip-rake) planes. The group of best solutions of similar quality is depicted in Fig. 4a, labelled "S1". In Fig. 4b, rms errors are plotted as a function of strike, dip and rake separately. Each small circle represents one trial solution. Only rms values smaller than 1.0 are plotted. Uncertainties on the strike and rake parameters can be estimated to ±25° and ±15° for the dip from Fig. 4a and b. The observed and synthetic seismograms for solution S1 are drawn in continuous and dashed line, respectively (Fig. 5). Waveforms are correctly matched, except the latest

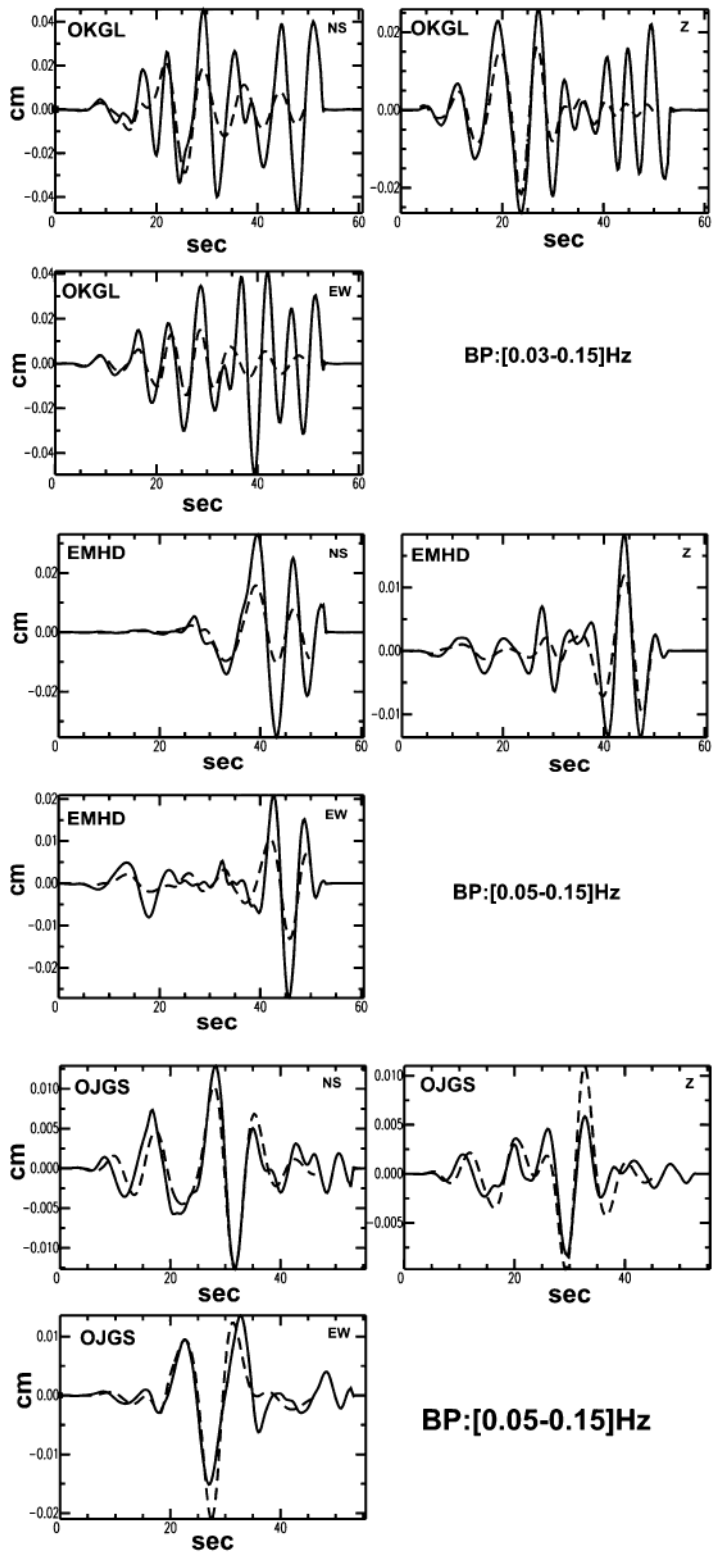

Figure 5
Best fit of displacement (cm) between observed (*continuous line*) and synthetic (*dashed line*) seismograms

waves at station OKGL which display the most complex propagation effects.

6. Regional Stress Inversion

We used the GRIDSIM_STRESS algorithm (Delouis *et al.*, 2002) in order to deduce the orientation of the main axes of the stress tensor ellipsoid. We used the FM (1) of the mainshock calculated in this study, and (2) fourteen events occurred in the Cheliff region between 1954 and 1992 (Table 2; Fig. 2). The criteria for selecting the best stress tensors is to minimize the angular deviation between observed and computed slip vectors on the nodal planes of the focal mechanisms. We define a score, which is the average of the scalar products between observed and computed slip vectors. A scalar product of 1 means perfect matching (no angular deviation). Minimizing the angular deviations is equivalent to maximize the score, whose largest possible value is 1. The stress tensor shape factor R is defined as in Rivera and Cisternas (1990), where the domains $(R < 0)$, $(0 < R < 1)$, and $(R > 1)$ correspond to compressional, strike-slip, and extensional stress regimes respectively. The stress axes orientation and the stress shape factor are explored using a combination of grid search and simulated annealing. The different calculated solutions are plotted in Fig. 6. The value of the best score is 0.99, showing that all focal mechanisms are very consistent with a common stress tensor. The best solutions (black circles) indicate that the compression axis $\sigma 1$ is horizontal and oriented N145°, the $\sigma 2$ axis is near vertical, and the $\sigma 3$ axis is horizontal oriented N55°. On the diagram of score versus shape factor R, we note that the best scores correspond either to the strike-slip regime or to the compression regime. This is related to the possibility for both $\sigma 2$ and $\sigma 3$ to be either near vertical or near horizontal, as shown by the stereographic plots. The $\sigma 1$ axis is best constrained for this case.

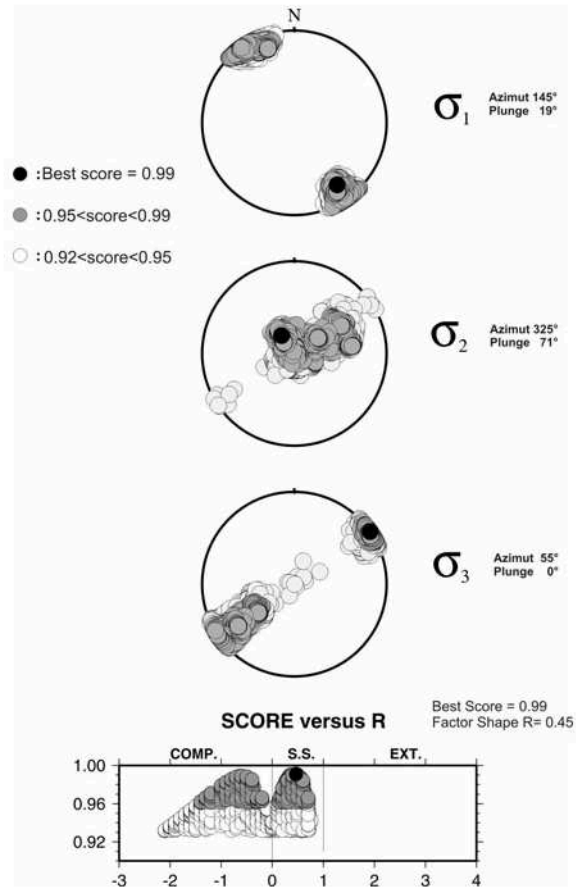

Figure 6
Stress tensor inversion for the Cheliff region. Solutions for σ_1, σ_2 and σ_3 are plotted in three lower hemisphere equal-area stereographic projections. The azimuth and plunge of the principal stress are indicated. The best solution (best score 0.99) is represented by the *black dot*. The score is, here, the average of the scalar products between observed and computed slip vectors. A scalar product of 1 means that the observed and computed slip vectors are identical, 0 means that they are perpendicular; negative scalar products can occur, when the angular difference is more than 90°. Solutions with a score greater than 0.95 and 0.92 are represented by *grey* and *open circles*, respectively. At the bottom, the same solutions are plotted as a function of their R and score value. The shape factor $R = (\sigma_z - \sigma_x)/(\sigma_y - \sigma_x)$, with $\sigma_y \geq \sigma_x$ and σ_z being the stress axis closest to the vertical. The domains corresponding to the compressional $(R < 0)$, strike-slip $(0 < R < 1)$, and extensional $(R > 1)$ regimes are labelled respectively COMP., S.S. and EXT. Here, the R factor corresponding to the best solution is equal to 0.45, indicating a strike-slip regime, but equally good solutions are found in the compressional regime

7. Discussion and Conclusion

The Cheliff region is a very active area where major earthquakes can occur. The region has experienced the strongest and most destructive earthquake of Algeria (El Asnam, on October 10, 1980 with Ms = 7.3) since the instrumental era. The Tadjena event of December 16, 2006 which occurred

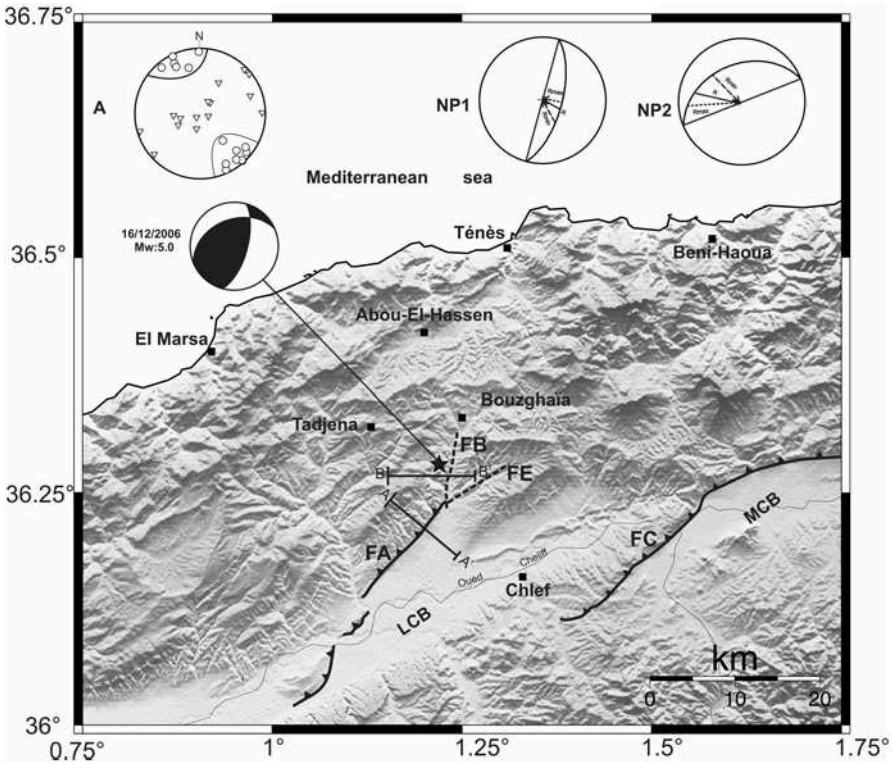

Figure 7

Digital Elevation Model (DEM) of the Cheliff region (SRTM 89 m). **a** Lower hemisphere projection of the P- (*open circles*) and T- (*open triangles*) axes of the regional focal mechanisms displayed on Fig. 2 and Table 2. NP1 and NP2: Nodal planes and slip vectors with their uncertainties for solution S1 corresponding to the focal mechanism of the December 16, 2006 event. *Rmax* rake max, *Rmin* rake min, *R* best rake. (Rmax = 80 and Rmin = 30 for NP1; Rmax = 160 and Rmin = 110 for NP2). *Black star* is epicentre location. *MCB* Middle Chelif Basin. *LCB* Lower Chelif Basin. *FA* Boukadir thrust fault. *FE* extension of *FA* segment *FB* Transverse fault. *FC* El Asnam thrust fault. AA' is cross section (N140°) of FA fault and BB' is cross section (N90°) of FB fault (see Fig. 8 for details)

in the Cheliff region on a distinct thrust fault system from the El Asnam 1980 event was a moderate (Mw = 5.0) and shallow one (depth = 7 km). The maximum intensity of V (MSK-scale) was observed at the villages of Bouzghaïa, Benaria, Abou El Hassen and Tadjena. At Chlef city, the intensity was IV. The focal mechanism of the event, obtained from a point source waveform inversion using the records of three digital regional broadband seismic stations of ADSN, shows thrust-reverse faulting with a strike-slip component. Two focal mechanisms were determined for the Tadjena event, one by the IGN Madrid, the other one by this study. Both solutions differ substantially (Fig. 1). In particular, the NE—SW nodal plane is near vertical (dip 81°) in the IGN solution and shallow dipping (38°) in ours. It should be noted that a complete exploration of the parameter space allowed us to determine that our focal mechanism is well constrained within ±25° for the strike and rake, and ±15° for the dip angle. In the case of the IGN solution, no such assessment is performed. However, the IGN focal mechanism was obtained with three stations located in Ibiza and on the eastern coast of Spain providing a relatively narrow azimuth aperture (53°), a situation presumably not very favourable for a good constraint of the focal parameters.

The P- and T-axes of all the focal mechanisms of the Cheliff region displayed in Fig. 2 are shown on a lower hemisphere plot in Fig. 7a. Remarkably, the P-axes exhibit little dispersion, with an average P-axis near horizontal trending N147° ± 30° in agreement with result published by Ouyed *et al.*, (1983) for the aftershock sequence of El Asnam

Table 7

Application of the best regional stress tensor (shown in Fig. 6) on the nodal planes of solution S1 found for the focal mechanism of the Tadjena earthquake (Fig. 4)

Nodal planes	Strike°	Dip°	Rake°	Rake predicted by the stress tensor °	Score
NP1 solution S1	15	65	60 ± 25	14	0.69
NP2 solution S1	249	38	137 ± 25	126	0.98

When applying the stress tensor to a nodal plane defined by its strike and dip parameters, a value of rake are predicted. The score is the scalar product between the unit slip vectors obtained from the focal mechanism and predicted by the stress tensor (cosine of the angular difference). The rake from the focal mechanism is indicated with its uncertainty of ±25° as inferred graphically from Fig. 4

(N146°) and by Lammali *et al.,* (1997) for northern Algeria.

The stress regime determined for the Cheliff region is either strike-slip or compressional, depending on whether the axis closer to the vertical is $\sigma 2$ or $\sigma 3$. The ambiguity is related to the existence of normal faulting components for a few focal mechanisms, numbers 3 and 12 mainly, which tends to promote stress tensor solutions with a vertical $\sigma 2$ axis (strike-slip regime). However, since none of these events exhibits pure normal faulting, stress tensor solutions having a $\sigma 3$ axis near the vertical (compressional regime) are not precluded. The ambiguity, which may be related to the uncertainty in the focal solutions and to the limited number of events incorporated in the inversion (15), cannot be solved at this stage. Nonetheless, the most compressive axis $\sigma 1$ found by the stress inversion is well constrained to be near horizontal trending N145°, i.e. almost identical to the average P-axis.

This trend is identical to the direction found by Beldjoudi *et al.,* (2009) from the inversion of thirteen focal mechanisms in the eastern part of northern Algeria between longitudes 4.5°E and 8°E. Stich *et al.,* (2006) and Ayadi *et al.,* (2008) inverted the stress tensor with datasets predominated or entirely constituted by the focal mechanisms of the 2003 Boumerdes earthquake (Mw 6.9) sequence. These two studies provided similar results with a N350°E (N170°E) direction of $\sigma 1$ for the central part of northern Algeria, essentially between longitudes 2.5°E and 4.5°E. Stich *et al.,* (2006) already suggested a counter-clockwise rotation of $\sigma 1$ westward along the coast of northern Algeria (longitude <2°E). Our results, indicating $\sigma 1$ trending N145°E between longitudes 0.5 and 2°E confirm this later statement.

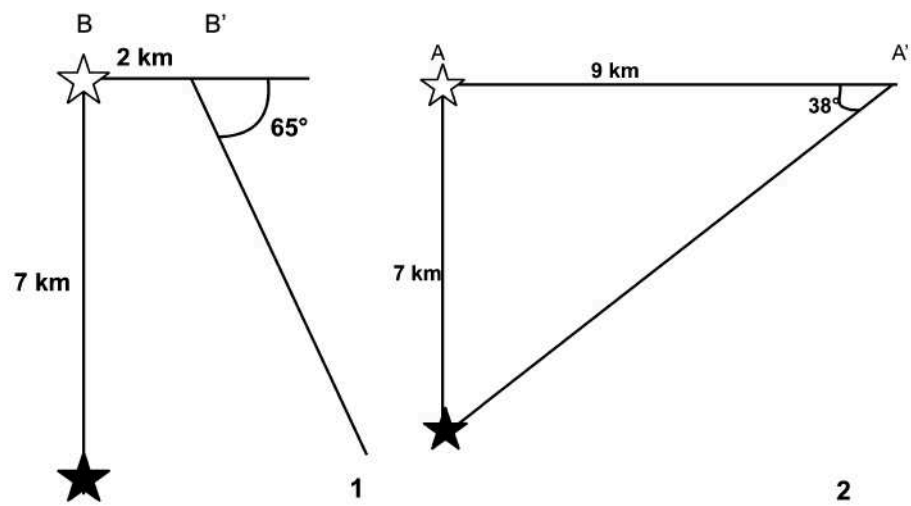

Figure 8

Projection of the hypocentre and of the two alternative fault planes deduced from the focal mechanism. 1 Case of the transverse fault (FB). 2 Case of the Boukadir fault (FA)

Bounif *et al.*, (2004) found a stress tensor with the most compressive axis $\sigma 1$ nearly horizontal and almost oriented NS, from the aftershocks data of the Chenoua (Algeria) earthquake (Mw = 6.0, October 29, 1989), for the near coast area located between longitudes 2.30°E and 2.50°E.

Considered all together, these five different stress inversions suggest that the direction of maximum compression $\sigma 1$ is indeed undertaking a clockwise rotation of about 25° in the central part of northern Algeria with respect to the areas more to the east and to the west (Fig. 1). Whether this rotation is a characteristics of the 2003 Boumerdes sequence only or a more general property of central northern Algeria is difficult to assess given the scarcity of focal mechanisms in this area before the 2003 crisis.

The scalar product between the observed slip vector of the December 16, 2006 event and the slip vector predicted by the stress tensor is 0.69 for nodal plane NP1 dipping to the east (strike, dip, rake = 15, 65, 60, Fig. 4) and 0.98 for nodal plane NP2 dipping to the NW (strike, dip, rake = 249, 38, 137), corresponding respectively to angular deviations of 46° and 10° (Table 7). This gives hints about NP2 being more compatible with the regional stress tensor.

Several neotectonic events have been reported by Meghraoui (1982); Aoudia and Meghraoui (1995), Aoudia *et al.* (2000) and Yelles-Chaouche *et al.*, (2006) in the vicinity of the December 16, 2006 earthquake. Recently Korchi-Heddar (2007) updated the tectonic map by introducing DEM (SRTM) and satellite data. Comparing with the tectonic observations, the fault plane solution noted NP2 (strike, dip, rake = 249, 38, 137) could be associated with the Boukadir fault, indicated "FA" or and its possible extension "FE" on Fig. 7. We measured precisely the azimuth of the Boukadir fault from its map surface expression. The fault strikes N41°. The group of best solutions for nodal plane NP2 of solution S1 (Fig. 4a) spans azimuths between N220° and N270° which correspond to N40° to N90°. The Boukadir fault is, therefore, compatible in azimuth.

On the other hand, NP1 (strike, dip, rake = 15, 65, 60) could be associated with fault noted "FB", transverse to the lower Cheliff basin. The better compatibility of NP2 with the stress tensor determined from the inversion of the 15 focal mechanisms of the Cheliff region points toward the Boukadir fault as being the causative fault. The location of the epicentre to the west of the trace of the FB fault (Fig. 8-1) dipping to the east is not in favour of the FB fault being the causative structure. On the other hand the hypocentre at 7 km depth projects itself well on the Boukadir fault, assuming it is dipping 38° to the NW as indicated by the focal mechanism (Fig. 8-2). The FB fault could be compatible with the event if we assume that the real epicentre is indeed shifted 4–6 km more to the east, a solution less probable but which cannot be completely excluded considering uncertainties on epicentre location. However, more geological field investigations are recommended in the epicentral region in order to better characterize the recent fault activity of both the Boukadir (FA) and the transverse (FB) faults.

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3.2. The Beni-Ilmane (Algeria) seismic sequence of May 2010: Seismic sources and stress tensor calculations (H. Beldjoudi, B. Delouis, H. Djellit , A. Yelles-Chaouche, S. Gharbi, I. Abacha)

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The Beni-Ilmane (Algeria) seismic sequence of May 2010: Seismic sources and stress tensor calculations

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ABSTRACT

A moderate earthquake with a moment magnitude of Mw 5.5 struck the Sub-Bibanique region of eastern Algeria on 14 May 2010, killing three people, injuring hundreds of others, and causing moderate damages in the epicentral area, mainly in the villages of Beni-Ilmane and Samma. The focal mechanism of the seismic source for the first shock, obtained by near-field waveform modelling, exhibits left-lateral strike-slip faulting with the first nodal plane oriented at N345°, and right-lateral strike-slip faulting with the second nodal plane oriented at N254°. A second earthquake that struck the region on 16 May 2010, with a moment magnitude of Mw 5.1, was located 9 km SW of the first earthquake. The focal mechanism obtained by waveform modelling showed reverse faulting with nodal planes oriented NE–SW (N25° and N250°). A third earthquake that struck the region on 23 May 2010, with a moment magnitude of Mw 5.2, was located 7 km S of the first shock. The obtained focal mechanism showed a left-lateral strike-slip plane oriented at N12° and a right-lateral strike-slip plane oriented at N257°. Field investigations combined with geological and seismotectonic analyses indicate that the three earthquake shocks were generated by activity on three distinct faults. The second and third shocks were generated on faults oriented WSW–ENE and NNE–SSW, respectively. The regional stress tensor calculated in the region gives an orientation of N340° for the maximum compressive stress direction (σ_1) which is close to the horizontal, with a stress shape factor indicating either a compressional or a strike-slip regime.

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1. Introduction

On 14 May 2010 at 12:29 GMT, a seismic sequence was initiated in the Hodna–Biban region, eastern Algeria (Fig. 1). Three people died and panic erupted within the populace in the epicentral area. Prior to the event, no instrumental seismological studies had been conducted in the region, with the exception of studies conducted by Hatzfeld (1978), in which the focal mechanisms (FM) of several earthquakes were calculated using teleseismic data. On the other hand, numerous studies have examined historical seismicity (Rothé, 1950; Benhallou, 1985; Mokrane et al., 1994; Harbi et al., 2003, based on macroseismic studies, press records, etc.) and the general geology (Caire, 1957, 1965; Guiraud, 1977, 1990; Meghraoui, 1988; Boudiaf, 1996) of the region. The mainshock of the 2010 sequence was recorded at Mw 5.5, and was followed by aftershock activity that included 1403 events recorded and located by Yelles-Chaouche et al. (2014). Two significant events occurred after the main event, on 16 May 2010 and 23 May 2010, with moment magnitudes (Mw) of 5.1 and 5.2, respectively. This study complements the study of Yelles-Chaouche et al. (2014), and contributes the following points: (1) The FM and Mw values of seismic events in the sequence were re-computed using a full waveform inversion approach (FMNEAR) instead of first motions. (2) A stress inversion study was conducted at a regional scale to examine the response of the sequence to a coherent regional stress tensor, as opposed to a local stress perturbation. (3) A detailed compilation of geological information, synthesised via geological mapping, provides a clear view of the distribution of faults in the epicentral area. The detailed configuration of faults was necessary for relating the different shocks in the sequence to known geological structures. (4) A new seismotectonic model was constructed, taking into account the more detailed information obtained during this study. In particular, we observed differences in the nodal planes of the second shock, which implied a modified relationship to the geological structures. Our FMs show strong similitudes to those found by Global Centroid Moment Tensor (*GCMT*: *http://www.globalcmt.org*) and Istituto Nazionale di Geofisica e Vulcanologia (INGV: http://www.bo.ingv.it/RCMT/), but stable solutions with respect to regional data are needed to ground interpretations in well-constrained focal solutions. The aftershock distribution determined by Yelles-Chaouche et al. (2014) shows two orthogonal directions, E–W and N–S. In this work, we analysed the seismic source of the mainshock and of the two strongest following earthquakes, using broadband seismic records of the Algerian Digital Seismic Network (ADSN), at an epicentral distance of ~25 to 325 km. The analyses of the three seismic sources and the tectonic context lead

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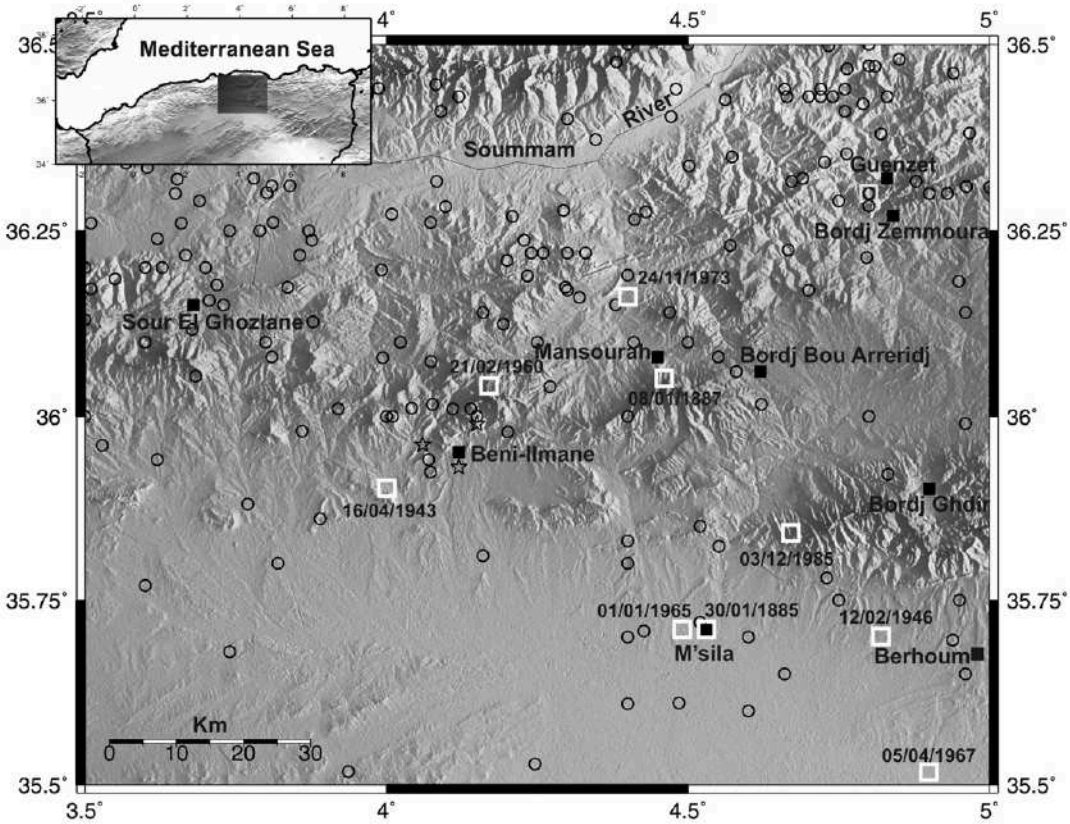

Fig. 1. Historical seismicity in the Biban–Hodna Region from 9 February 1850 to 13 May 2010. White open squares show the main earthquakes (historical and instrumental) reported in the region (see text). Open circles show areas of moderate seismicity. Stars show the three events of May, 2010. Black solid squares indicate cities and main villages.

us to refine the preliminary seismotectonic model proposed by Yelles-Chaouche et al., and to conclude that rupturing occurred on three distinct faults in a complex geological and seismotectonic domain.

2. Seismicity of the Hodna–Biban region

The seismicity catalogue of the Hodna–Biban region dates to the Guenzet–Zemoura earthquake on 9 February 1850. Rothé (1950) reported that this earthquake was felt as far away as Algiers (~170 km from the epicentre). Harbi and Maouche (2009) and Harbi et al. (2003) assigned a maximum observed intensity (Io) of VII (European Macroseismic Scale, EMS) to this earthquake (Fig. 1). Other significant earthquakes (greater than intensity VII) in the region during the period of 1850–1973 occurred near the following sites: (1) M'sila on 1885 and 1965 (Harbi et al., 2003; Hatzfeld, 1978; Rothé, 1950), (2) Mansourah on 1887, 1943, and 1973 (Harbi et al., 2003; Hatzfeld, 1978; Rothé, 1950) (3) Berhoum on 1946, and (4) Melouza on 1960 (Harbi et al., 2003; Hatzfeld, 1978; Rothé, 1950) (Fig. 1 and Table 1). An earthquake on 1 January 1965, located just south of M'sila (Hatzfeld, 1978) in the Hodna region, exhibited a body wave magnitude (Mb) of 5.1. Hatzfeld (1978) calculated the focal mechanism displaying strike-slip faulting using a vertical first plane oriented at N160° and a second plane oriented at N250°, dipping at 78° (Fig. 2).

Others FMs in the Hodna–Biban region were calculated by Hatzfeld (1978) and Beldjoudi in this study, or were obtained from the (GCMT) catalogue for the period of 1973–2012 (Table 2). Earthquakes in the Mansourah region during 1973 and 1974 were amongst the largest in the region, with magnitudes of up to 4.9. For example, an Mb 4.7 earthquake occurring in northeast Mansourah on 29 June 1974 showed a strike-slip FM with right-lateral movement for the first plane (oriented NW) and left-lateral movement for the second plane (oriented ENE–WSW) (Fig. 2, Table 2). An event in the Bordj Bou Arreridj (BBA) region (NE of Mansourah) in 2012 showed thrust fault movement along planes oriented NE–SW (Fig. 2, Table 2); waveform modelling of this event is presented in the Appendix A. The FMs in the region, as determined by various studies (Hatzfeld, 1978, Yelles-Chaouche et al., 2014), are principally strike-slip (60%) and thrusting (40%).

Historical records show that the Hodna–Biban region is seismically active, with the seismicity distributed on numerous active structures. Maximum Io values of VIII–IX have been observed in the area of Mansourah (Roussel, 1973; Bezzeghoud et al., 1996; Boughacha et al., 2004).

Table 1
Main earthquakes observed in the Hodna–Biban region from 9 February 1850 to 24 November 1973.

Date	Intensity	Coordinates		Region
		Latitude°	Longitude°	
09/02/1850	VII	36.30	4.8	Guenzet-Bordj Zemmoura
30/01/1885	VII	35.71	4.53	M'sila
03/12/1885	IX	35.84	4.67	M'sila
08/01/1887	VIII	36.05	4.46	Mansourah
16/04/1943	VIII	35.90	4.00	Mansourah
12/02/1946	VIII	35.66	5.03	Berhoum
21/02/1960	VIII	36.04	4.17	Melouza
01/01/1965	VIII	35.71	4.49	M'sila
05/04/1967	VII	35.50	4.90	M'sila
24/11/1973	VIII–IX	36.16	4.40	Mansourah

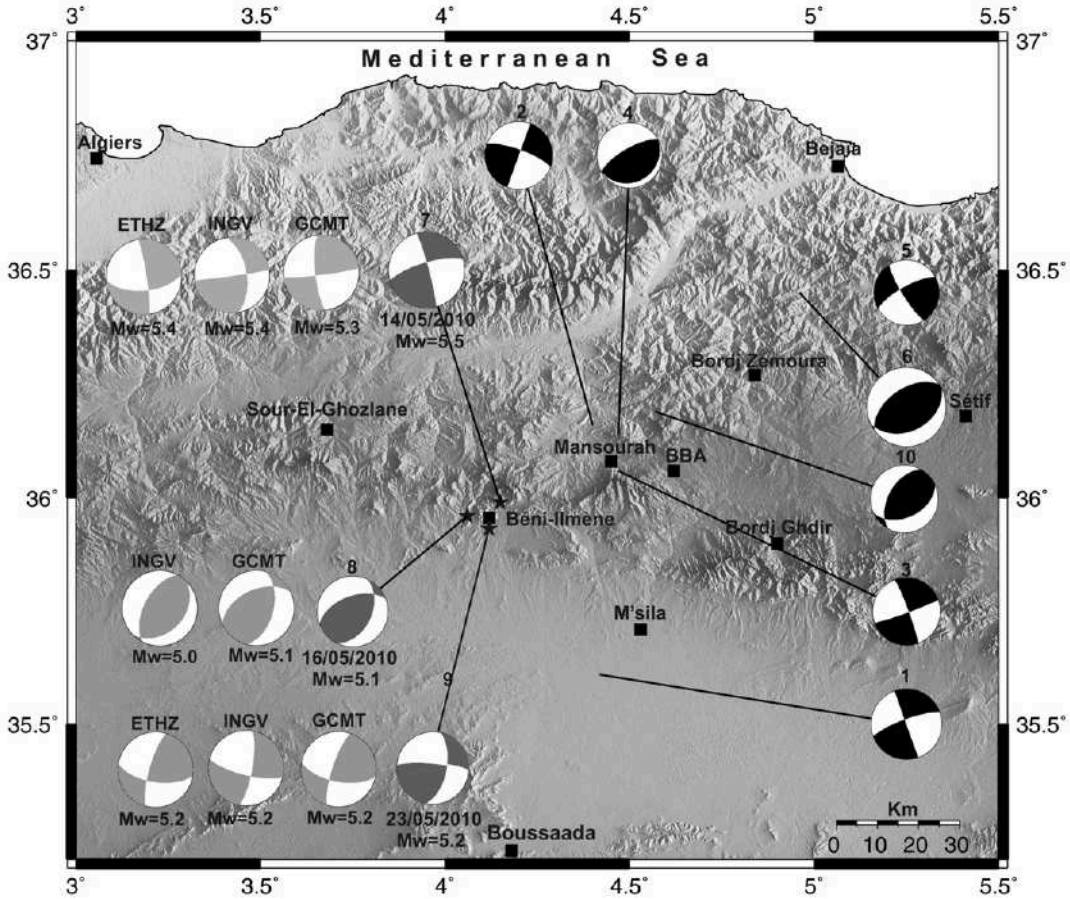

Fig. 2. Focal mechanisms of the Biban–Hodna region, from 1 January 1965 to 5 May 2010 (see Table 2 for references).

3. Earthquake impacts and site effects

On 14 May 2010 at 12:29 GMT, a Mw 5.5 earthquake occurred near the Beni-Ilmane village (SW of Mansourah) in the Hodna–Biban region (Fig. 2). Another Mw 5.1 earthquake struck the same region ca. 42 h later (16 May at 06:52 GMT); this second earthquake was located 9 km SW of the first earthquake. On 23 May 2010 at 13:28 GMT (ca. 6 days after the second

Table 2
Focal mechanisms available in the region and calculated by different authors between 1965 and 2012. Body wave magnitudes (mb) for events 1–5 are given. Moment magnitudes (Mw) are given for events 6–10, and duration magnitudes (Md) for events 11–25.

	Date	Lat.°	Long.°	Depth (km)	M	Plan 1		Plan 2		Reference
						Az.	Dip	Az.	Dip	
1	01/01/1965	35.61	4.42	13	5.1	160	90	250	78	Hatzfeld (1978)
2	24/11/1973	36.16	4.40	15	4.9	20	90	290	76	Hatzfeld (1978)
3	24/11/1973	36.06	4.47	8	4.9	70	86	336	86	Hatzfeld, (1978)
4	25/11/1973	36.14	4.47	19	4.7	240	72	60	18	Hatzfeld, (1978)
5	29/06/1974	36.52	5.21	32	4.7	238	74	148	80	Hatzfeld (1978)
6	10/11/2000	36.45	4.96	10	5.7	64	38	235	52	GCMT Catalogue
7	14/05/2010	35.99	4.15	6	5.5	345	85	254	74	This study
8	16/05/2010	35.96	4.06	6	5.2	25	45	250	55	This study
9	23/05/2010	35.93	4.12	6	5.2	275	80	12	57	This study
10	05/05/2012	36.19	4.57	6	4.9	205	30	66	66	This study
11	14/05/2010	35.92	4.11	6	4.7	88	85	358	87	Yelles-Chaouche et al. (2014)
12	14/05/2010	35.98	4.13	7	4.4	89	66	343	61	Yelles-Chaouche et al. (2014)
13	15/05/2010	35.97	4.09	9	4.0	273	58	88	32	Yelles-Chaouche et al. (2014)
14	16/05/2010	35.97	4.11	8	4.8	345	82	78	22	Yelles-Chaouche et al. (2014)
15	16/05/2010	35.97	4.09	9	4.0	292	71	45	41	Yelles-Chaouche et al. (2014)
16	19/05/2010	35.95	4.09	8	4.1	283	49	76	44	Yelles-Chaouche et al. (2014)
17	21/05/2010	35.95	4.12	8	4.0	85	89	176	78	Yelles-Chaouche et al. (2014)
18	23/05/2010	35.91	4.09	5	4.1	7	85	100	64	Yelles-Chaouche et al. (2014)
19	24/05/2010	35.91	4.11	7	4.9	3	88	93	72	Yelles-Chaouche et al. (2014)
20	24/05/2010	35.92	4.11	6	4.1	3	89	93	55	Yelles-Chaouche et al. (2014)
21	25/05/2010	35.91	4.11	5	4.5	3	87	94	76	Yelles-Chaouche et al. (2014)
22	26/05/2010	35.90	4.11	6	4.2	12	73	280	85	Yelles-Chaouche et al. (2014)
23	26/05/2010	35.91	4.11	6	4.3	179	89	89	64	Yelles-Chaouche et al. (2014)
24	30/05/2010	35.94	4.12	5	4.2	55	51	297	60	Yelles-Chaouche et al. (2014)
25	31/05/2010	35.92	4.11	7	4.6	23	89	113	80	Yelles-Chaouche et al. (2014)

Table 3
The three largest shocks of the 2010 sequence, located by CRAAG (Centre de Recherche en Astronomie Astrophysique et Géophysique), ETHZ (Swiss Federal Institute of Technology in Zurich), GCMT (Global Centroide Moment Tensor), and INGV (Istituto Nazionale di Geofisica e Vulcanologia). Data on the moment magnitude (Mw), seismic moment, strike, dip, and rake are from this study and from various seismological centres.

Date	Seismological centres	Time (GMT)	Lat.°	Lon.°	Depth (km)	Mw	Seismic Moment Nm	Plane 1			Plane 2			RMS	Conf. (%)	No. of Comp.
								Strike	dip	rake	strike	dip	rake			
1st shock 14/05/2010	This study	12 h 29 min	35.99	4.15	6.0	5.5	1.83×10^{17}	254	74	175	345	85	16	0.385	74	21
	ETHZ		35.88	4.14	18.0	5.4	0.16×10^{18}	85	60	−173	352	84	−29	–	–	–
	GCMT		35.90	4.14	12.0	5.3	1.30×10^{17}	83	85	167	174	77	5	–	–	–
	INGV		36.02	4.21	10.0	5.4	1.40×10^{17}	82	88	−148	351	58	−3	–	–	–
	Yelles-Chaouche et al. (2014)							89	75	175	180	86	15	–	–	–
2nd shock 16/05/2010	This study	06 h 52 min	35.96	4.06	6	5.1	5.50×10^{16}	25	45	55	250	55	120	0.61	63	15
	ETHZ		35.84	4.05	–	5.1	-		–			–		–	–	–
	GCMT		35.97	4.19	12.0	5.1	6.45×10^{16}	14	53	53	245	50	128	–	–	–
	INGV		35.88	3.99	10.0	5.0	5.05×10^{16}	45	35	104	209	56	80	–	–	–
	Yelles-Chaouche et al. (2014)							72	9	64	279	82	94	–	–	–
3rd shock 23/05/2010	This study	13 h 28 min	35.93	4.12	6	5.2	7.40×10^{16}	275	80	146	12	57	12	0.472	62	14
	ETHZ		35.84	4.12	18.0	5.2	0.81×10^{17}	99	63	165	196	77	27	–	–	–
	GCMT		35.88	4.17	12.0	5.2	7.14×10^{16}	101	63	165	199	76	28	–	–	–
	INGV		35.91	4.08	12.0	5.2	5.58×10^{16}	104	75	−166	10	76	−15	–	–	–
	Yelles-Chaouche et al. (2014)							104	80	175	195	85	10	–	–	–

earthquake), a third earthquake struck, with a magnitude of Mw 5.2 (Table 3).

The first earthquake killed 3 people and injured 170, and a large number of aftershocks caused panic amongst the local population in the localities of Beni-Ilmane, Ouenougha, and Samma; the sequence of three principal earthquakes of Mw 5.5, 5.1, and 5.2 was especially strong. In the epicentral region, old traditional houses with only a ground floor and built with stone and argillaceous mortar were severely damaged (Fig. 3b). Modern multi-storey buildings built with bricks and concrete (up to 4 storeys in height) were slightly damaged. Rockfall was observed at Djebel Kef El Khert (Fig. 3e), situated on the western margin of the region affected by the first and the third earthquakes, and located above the second earthquake source. Surface fractures were observed to the north and south of Kef El Khert. To the north, at the Kasbat Beni-ilmène cemetery (Fig. 3f and 3g), a tensile crack (trending NNW–SSE) formed in vertical bedding and was ~10 cm wide and ~150 m long. Approximately 1.5 km south of Kasbat Beni-Ilmène, continuous fissures several metres long and trending NNE–SSW were observed in Eocene–Cretaceous rocks (Fig. 3a and 3c). Zazoun et al. (2012) conducted a detailed study of the surface fractures observed in the epicentral area. The fractures, trending NE–SW, were observed over a distance of 5 km and were considered related to induced effects, i.e. not related to surface faulting. Yelles-Chaouche et al. (2014) attributed a maximum Io of VII (EMS-98) to the event.

4. Geological setting

The Biban region, located ~100 km ESE of the capital of Algiers (Fig. 4), corresponds to the southern part of the Tellian Atlas. The region is bounded to the south by the northern front of the Hodna basin and to the north by the Soummam River valley. Morphologically, the Biban region consists of three distinct areas (Caire, 1957): (1) a mountainous

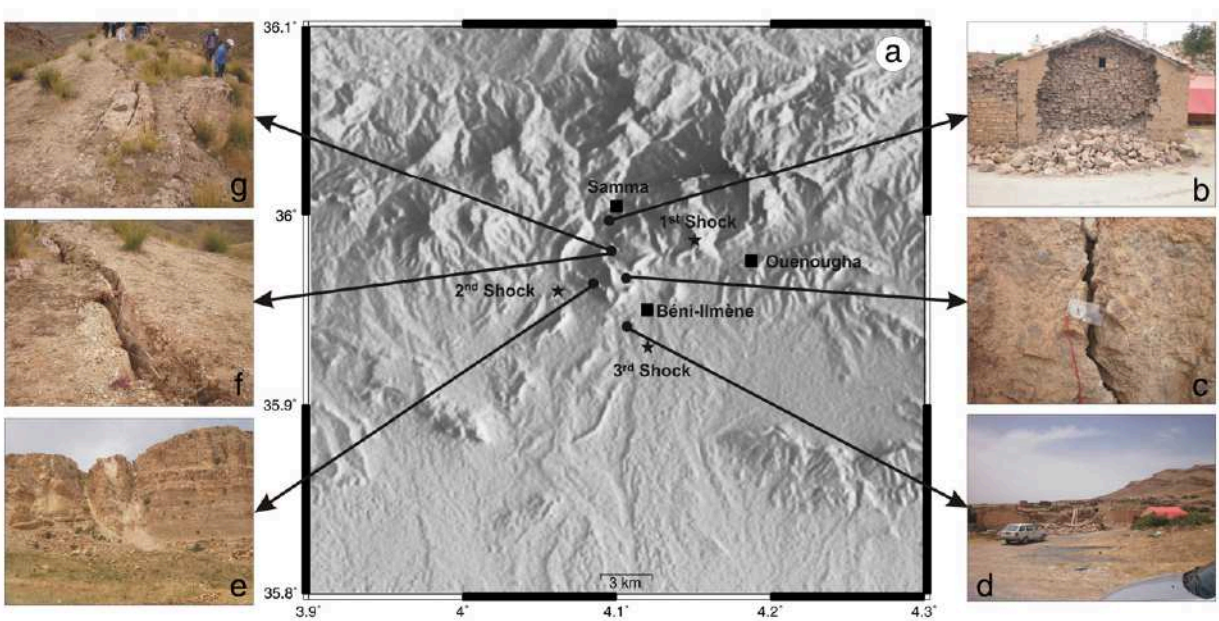

Fig. 3. (a) Site effects in the epicentral area. Stars show the epicentre locations of the three largest shocks of the 2010 sequence. Circles show the locations of site effects. (b) Collapsed wall of an old house at Samaa village. (c) Surface breaks (~3 cm wide) oriented NNW that can be followed for ~50 m. (d) Old house totally collapsed, located south of the epicentral area. (e) Rockfall at Djebel Kherrat. (f and g) Surface breaks (~10 cm wide) oriented NNW that can be followed for ~150 m, at the Kasbat Beni-Ilmane cemetery.

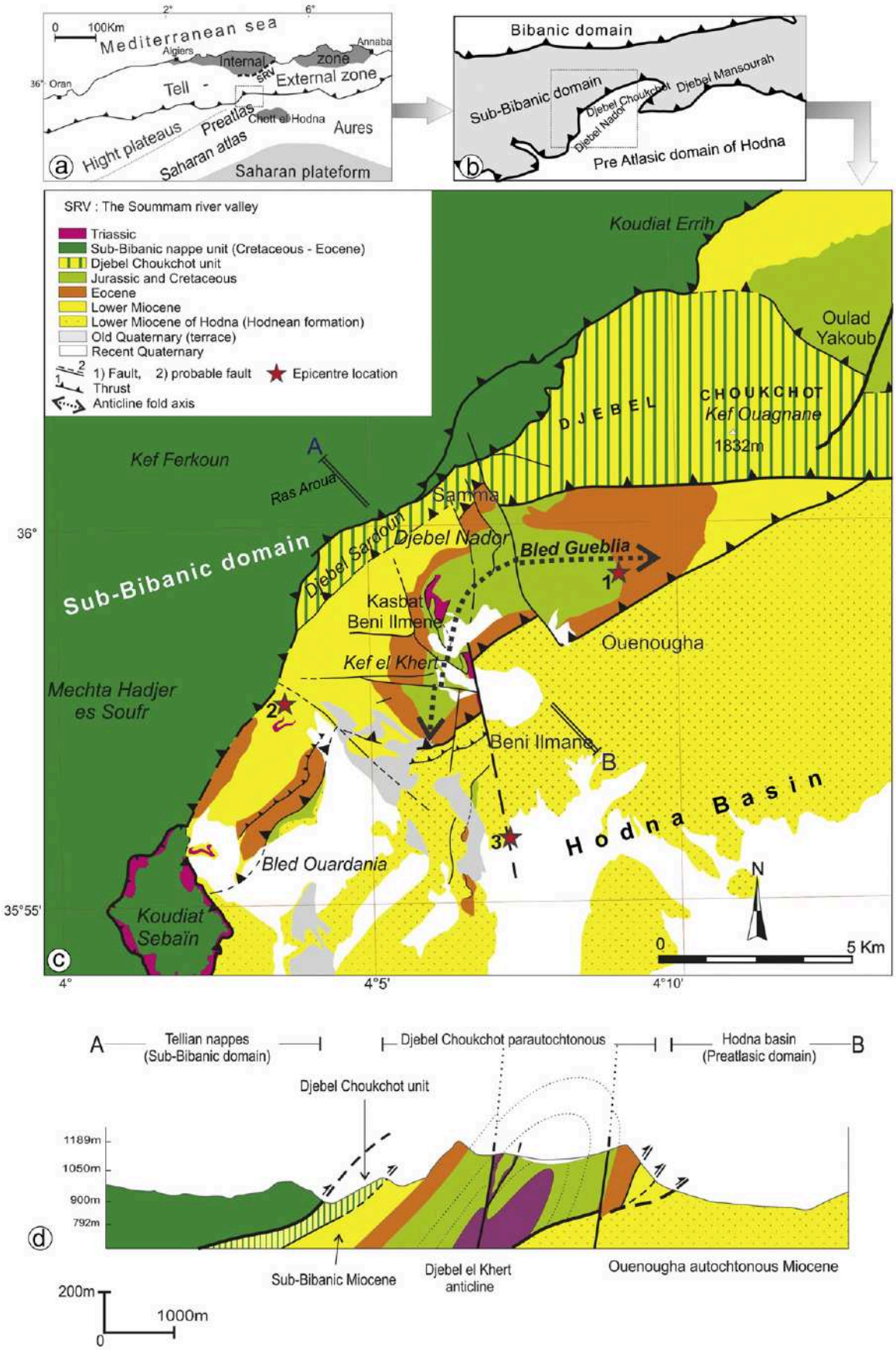

Fig. 4. (Top) Geological map of the epicentral area, modified after Baldini (1966), Kieken (1961), Savornin (1909), and Savornin (1929). a) Location of the epicentral area at a regional scale. b) Location of the epicentral area at a Local scale. c) Geology of the epicentral area. (Bottom) NW–SE geological cross-section (AB).

Table 4
Velocity model used to generate synthetic seismograms. V_p and V_s are the *P*- and *S*-wave velocities, respectively; Q_p and Q_s are the quality factors for *P* and *S* waves, respectively. The standard model is based on the FMNEAR method (Delouis, 2014).

Layer thickness (km)	Vp (km/s)	Vs (km/s)	Density (g/cm^3)	Qp	Qs
0.60	3.30	1.90	2.00	200	100
1.40	4.50	2.60	2.30	350	175
3.00	5.50	3.18	2.50	500	250
25.00	6.50	3.75	2.90	600	300
Half space	8.10	4.68	3.30	1000	500

portion, formed by the Djebel Mansourah and Djebel Choukchot, (2) the Sub-Bibanique portion, and (3) the Bibanique portion (Fig. 4).

The epicentral area of the Beni-Ilmane earthquake sequence is located at the boundary of two distinct geological domains. The area is bounded to the north by the Alpine Mountains and the Tellian Sub-Bibanique nappe (Caire, 1957), formed mainly of Upper Cretaceous and Eocene sedimentary cover. The area is bounded to the south by the pre-Atlasic domain, located on the northern border of the autochthonous Hodnéen Formation, and is covered by Neogene and Quaternary deposits over wide areas. Tectonically, the domain is less deformed than the domains in other regions of northern of Algeria.

Morphologically, the Biban region was shaped by a combination of tectonic and climatic processes (Caire, 1965) that created significant variations in altitude. The Jurassic–Cretaceous and the Paleogene, which consist of calcareous and dolomitic material (Guiraud, 1990), form the high-elevation regions, which include Djebel Choukchot (1798 m) in the north, Kef El Khert (1362 m) and Djebel Nador (1322 m) in the epicentral area, and Djebel Sardoun (1322 m) in the NW part of the epicentral area. To the south, an area of lower relief is formed by valleys and plains of the Beni-Ilmane basin; the elevation (average, 800 m) decreases progressively towards the south and reaches an elevation of 400 m in the Chott El Hodna.

Recent tectonic activity in the Hodna basin has been reported by Grandjean et al. (1966), Guiraud (1977), Meghraoui (1988), Boudiaf (1996), and Kheidri et al. (2007). The faults, which were inherited from Mesozoic–Cenozoic tectonic activity, were reactivated during the Pliocene–Quaternary (Guiraud, 1977; Wildi, 1983; Bracène and Frizon De Lamotte, 2002), and low–moderate earthquake activity is now recorded in the region.

The Beni-Ilmane seismic sequence was located in a faulted fold (Fig. 4) that is strongly deformed in its central and western parts, thus showing the complexity of the geological history off the area. The centre (core) of the faulted fold consists of Lower Cretaceous rocks dipping to the south; an overturned flank consists of Eocene strata that overlie the lower Miocene strata of the Hodna Basin. The average fold trend is NNE–SSW to E–W (parallel to the trends of Atlasic folds). Major active strike-slip faults with dominant N–S to NNW–SSE strikes are also observed; in some cases, Triassic material has been injected into the faults. The faults cross-cut the Kef El Khert fold and to a lesser extent recent deposits. Dextral strike-slip faults oriented NW–SE to E–W and thrust faults striking NE–SW to E–W and dipping to the NW–N are also visible.

Table 5
Characteristics of broadband stations used in the waveform modelling of the first shock (14 May 2010, Mw 5.5). The N, E, and Z designations at the beginning of each station name are the north, east, and vertical components, respectively. BP: band-pass Butterworth filter, using three poles. Δ: epicentral distance.

First shock		
Stations	Δ (km)	Filter (Hz)
NADJF	145	BP 0.02 0.07
EADJF	145	BP 0.02 0.07
ZADJF	145	BP 0.02 0.07
NATAF	25	BP 0.05 0.10
EATAF	25	BP 0.05 0.10
ZATAF	25	BP 0.05 0.10
NCBBR	183	BP 0.02 0.06
ECBBR	183	BP 0.02 0.06
ZCBBR	183	BP 0.02 0.06
NEMHD	88	BP 0.02 0.06
EEMHD	88	BP 0.02 0.06
ZEMHD	88	BP 0.02 0.06
NEADB	169	BP 0.02 0.06
EEADB	169	BP 0.02 0.06
ZEADB	169	BP 0.02 0.06
NOJGS	175	BP 0.03 0.07
ZOJGS	175	BP 0.03 0.07
NOKGL	188	BP 0.03 0.07

Table 6
Characteristics of broadband stations used in the waveform modelling of the second shock (16 May 2010, Mw 5.1). See notes in Table 5.

Second shock		
Stations BB	Δ(km)	Filter(Hz)
EABZH	98	BP 0.02 0.07
NADJF	141	BP 0.02 0.07
EADJF	141	BP 0.02 0.07
NCABS	181	BP 0.05 0.10
NCSVB	143	BP 0.02 0.07
NEADB	164	BP 0.02 0.07
EEADB	164	BP 0.02 0.07
NEMHD	82	BP 0.02 0.07
EEMHD	82	BP 0.02 0.07
ZEMHD	82	BP 0.02 0.07
NOJGS	171	BP 0.03 0.07
ZOJGS	171	BP 0.03 0.07
NOKGL	184	BP 0.03 0.07
EOKGL	184	BP 0.03 0.07
ZOKGL	184	BP 0.03 0.07

5. Focal mechanisms and source parameters

5.1. Data processing and inversion method

Data used for determinations of FMs and source parameters for the three shocks in the Beni-Ilmane sequence were recorded by the ADSN. Broadband records were integrated with displacement and band-pass

Table 7
Characteristics of broadband stations used in the waveform modelling of the third shock (23 May 2010, Mw 5.2). See notes in Table 5.

Third shock		
Stations	Δ (km)	Filter (Hz)
EABZH	88	BP 0.035 0.07
NADJF	134	BP 0.035 0.07
EADJF	134	BP 0.035 0.07
ECABS	179	BP 0.046 0.07
NEMHD	78	BP 0.035 0.07
EEMHD	78	BP 0.035 0.07
ZEADB	152	BP 0.035 0.07
NOJGS	171	BP 0.035 0.07
EOJGS	171	BP 0.035 0.07
ZOJGS	171	BP 0.035 0.07
NOKGL	184	BP 0.035 0.07
NCBBR	173	BP 0.035 0.07
ECBBR	173	BP 0.035 0.07
ZCBBR	173	BP 0.035 0.07

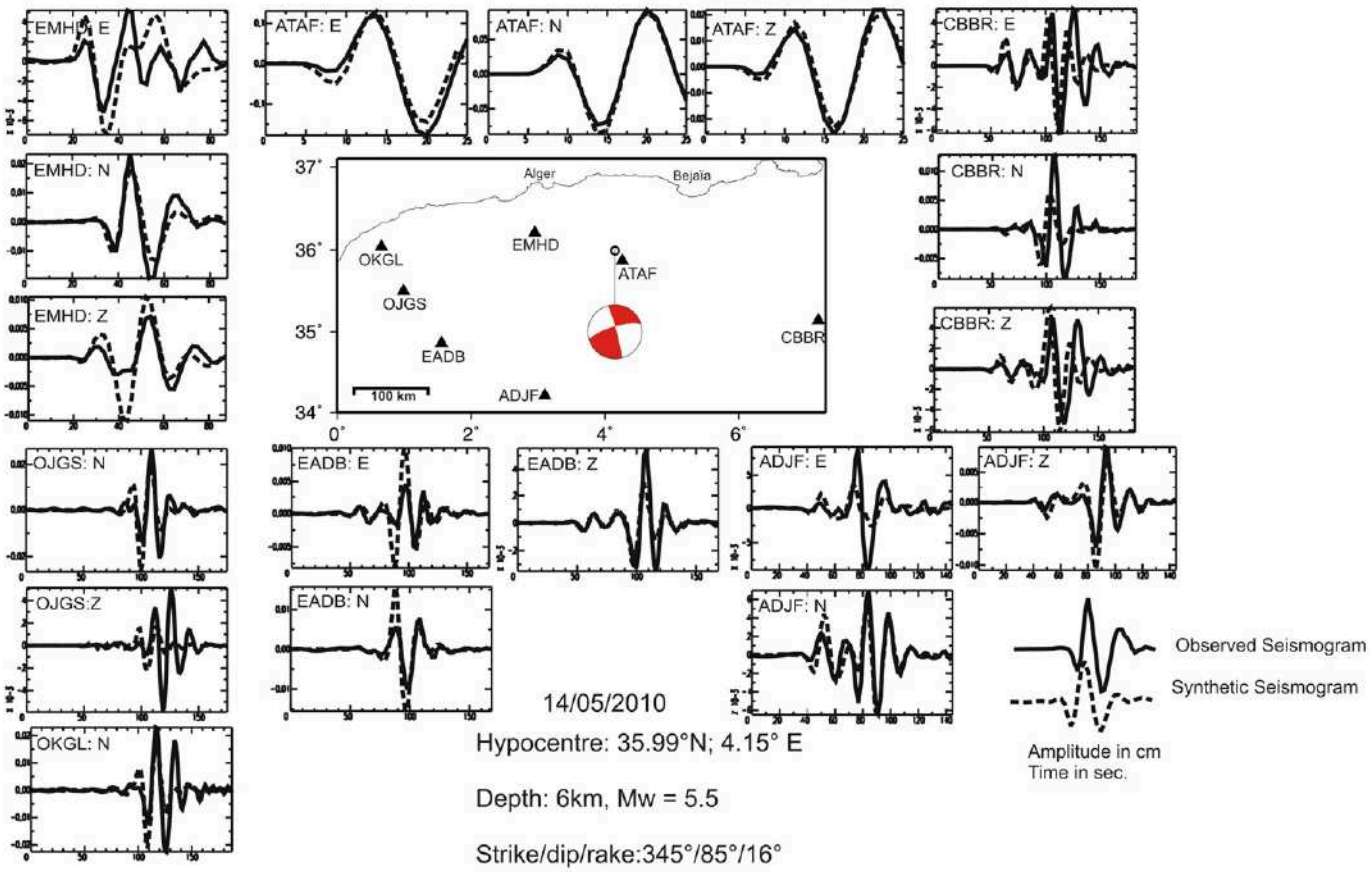

Fig. 5. Results of waveform inversion for the first shock of 14 May 2010 at 12 h 29 min GMT. The inset map shows the focal mechanism. Triangles indicate the broadband stations used to determine the focal mechanism. Synthetic seismograms (dashed lines) and observed seismograms (continuous lines) are superimposed, showing the best fit.

filtered to remove low-frequency noise and reduce the impact of high-frequency signals that are sensitive to the crustal velocity structure. The waveform inversion, performed at relatively low frequencies (ranging from 0.02 to 0.10 Hz), is less sensitive to the choice of velocity model. Synthetic seismograms, generated using the discrete wavenumber method of Bouchon (1981) and designed for 1D velocity models, were filtered in the same way. The velocity model used here and presented in Table 4 is the default model proposed by the FMNEAR method (Delouis, 2014), which can be seen as a kind of standard model for the continental crust. We tested different velocity models in the present case, in particular the best model of Yelles-Chaouche et al. (2014) for the same Biban–Hodna region, but waveforms were best modelled with the default model, which was hence retained for the final results. Observed and synthetic seismograms were initially aligned with respect to the first P-wave arrival. Waveform inversion was performed to determine the focal mechanism and moment magnitude (Mw), employing the near-source records (FMNEAR) method (Delouis, 2014). We describe the method only briefly here. For small to moderate events ($M < 5.5$), the earthquake is represented by a purely double couple point source. The inversion combines two grid search steps (on the strike and dip parameters of the FM), whilst the source time function and the rake (slip angle) are determined by simulated annealing runs. The criterion for selecting a solution is based on a normalised root mean square (RMS) misfit function: $\mathrm{RMS} = \mathrm{SQRT}\,[\sum(\mathrm{obs}-\mathrm{cal})^2/\sum(\mathrm{obs})^2]$, where SQRT is the square root, $\sum$ is the sum over all data points from the seismograms, and obs and cal are the observed and calculated amplitudes, respectively. During the inversion, FMNEAR methods automatically discard components for which waveforms are poorly fitted. As a result, some components (horizontal and vertical) may not be represented in the final solution. The FMNEAR method explores the focal depth and provides the optimal depth from the waveform modelling point of view. Certainly, depth depends on the velocity model, but this dependence is not very high when modelling the full waveforms at low frequency. Depth is re-estimated by FMNEAR, mainly by correcting depth in case of a strongly erroneous starting depth, but it cannot pretend to determine depth better than with an accuracy of a few kilometres.

A confidence index (CI), varying between 0% and 100%, describes the uniqueness and quality of the solution; Delouis (2014) provides a full description. The CI is based on the existence or non-existence of secondary FM solutions that have an RMS misfit similar to that of the best solution, but whose FM solutions are very different from the primary FM solution. If such secondary solutions exist, the CI is low; otherwise it is high. Based on the analysis of tens of earthquakes, a value of CI > 70% indicates a well-constrained solution (Delouis, 2014), and a value of CI > 60% represents an acceptable solution.

Filtering parameters (minimum and maximum frequencies) can be adapted manually and individually for each component of the seismograms (N, E, or Z), or determined automatically. The acceleration spectrum of the whole seismic signal is analysed, and deviations from the expected linear trend with positive slope (near Ω^2) at low frequencies are searched for. If such a deviation exists, the part below the near-linear trend reflects low-frequency noise that must be filtered out. Accordingly, the high-pass (i.e., low-cut) frequency (fmin) is identified as the frequency at which the acceleration spectrum maintains a positive near-constant slope up to the plateau area near the corner frequency. The low-pass (i.e., high-cut) frequency (fmax) is basically two

to four times the fmin value; it is constrained by the magnitude and distance, and by comparing the signal/noise amplitude ratio in the spectral domain. We thus derive the boundary frequencies fmin and fmax, which define the filtering band-pass for individual components (N, E, and Z). In the present study, the filtering parameters were automatically determined using the FMNEAR algorithm (Tables 5–7).

For all events, we assume that the epicentre has been well located, so that the latitude and longitude are fixed. Thirty seismic stations (broadband and short period) were used in the location of the events. Horizontal and vertical error locations respectively ERH and ERZ are ±3 km and root mean square (RMS) is ≤0.3. The depth, on the other hand, is less well constrained. Accordingly, for each event we performed a series of inversions with different fixed depths, from 1 to 20 km, with a step of 1 km. The best result, corresponding to the optimal depth, was selected by comparing the RMS, the number of components finally selected, and the CI.

With FMNEAR, good azimuthal coverage by the seismic stations is expected to help constrain the FM. Nevertheless, given that the full waveforms contain more information on the earthquake source than the first motions, the solution may be well constrained in cases where the azimuthal coverage is limited. Since FMNEAR explores the solution space using a nonlinear search algorithm, it can determine whether the solution is constrained or not under such "limited azimuthal coverage" conditions (see the discussion on the CI, Section 5.1).

5.2. Results

The first shock of the Beni-Ilmane seismic sequence occurred on 14 May 2010 at 12:29 GMT. Seven broadband stations (with a total of 21 components) were incorporated in the inversion procedure. Eighteen components were selected for the inversion (Table 5) to estimate the focal mechanism and the source parameters; the remaining three components were discarded by the inversion. The criteria to discard components are the following: low signal to noise ratio, no adequate filtering range found, or poor fit of the waveforms.

The RMS of the best solution is 0.385 and the CI is 74%. The optimal depth is 6 km. The focal mechanism is represented by plane 1, which describes approximately NNW left-lateral movement; and plane 2, which describes approximately E–W right-lateral movement (Table 3, Fig. 5).

The second shock occurred on 16 may 2010 at 06:52 GMT. Eight broadband stations (24 components) were incorporated in the inversion procedure. Fifteen components (selected by the inversion) (Table 6) were used to calculate the focal mechanism and the source parameters. The best solution (RMS = 0.61; CI = 63%) corresponds to a reverse fault oriented NE–SW with a small strike-slip component, at a depth of 6 km (Table 3, Fig. 6).

The third shock, occurred on 23 May 2010 at 13:28 GMT. The focal mechanism and the source parameters were estimated using nine broadband stations. The focal mechanism shows that plane 1 is oriented approximately E–W and shows right-lateral movement and plane 2 is oriented approximately N–S and displays left-lateral movement (Table 3, Fig. 7). The movements on both planes are associated with a small reverse component. The selected solution (depth, 6 km) has an RMS of 0.472 and a CI of 62%; the number of components selected by the inversion is 14 (Table 7).

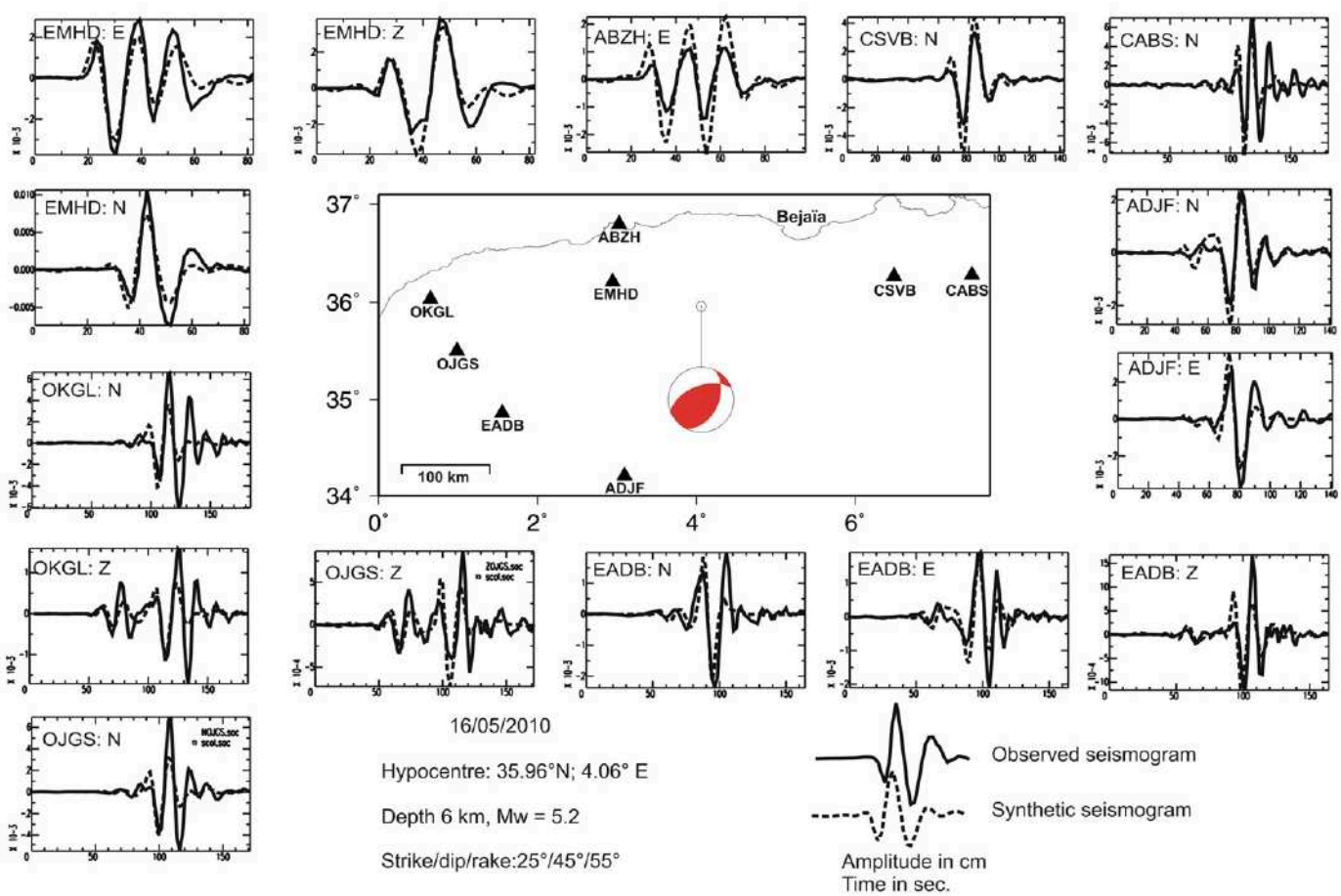

Fig. 6. Same as Fig. 5, but for the second shock of 16 May 2010 at 06 h 52 min GMT.

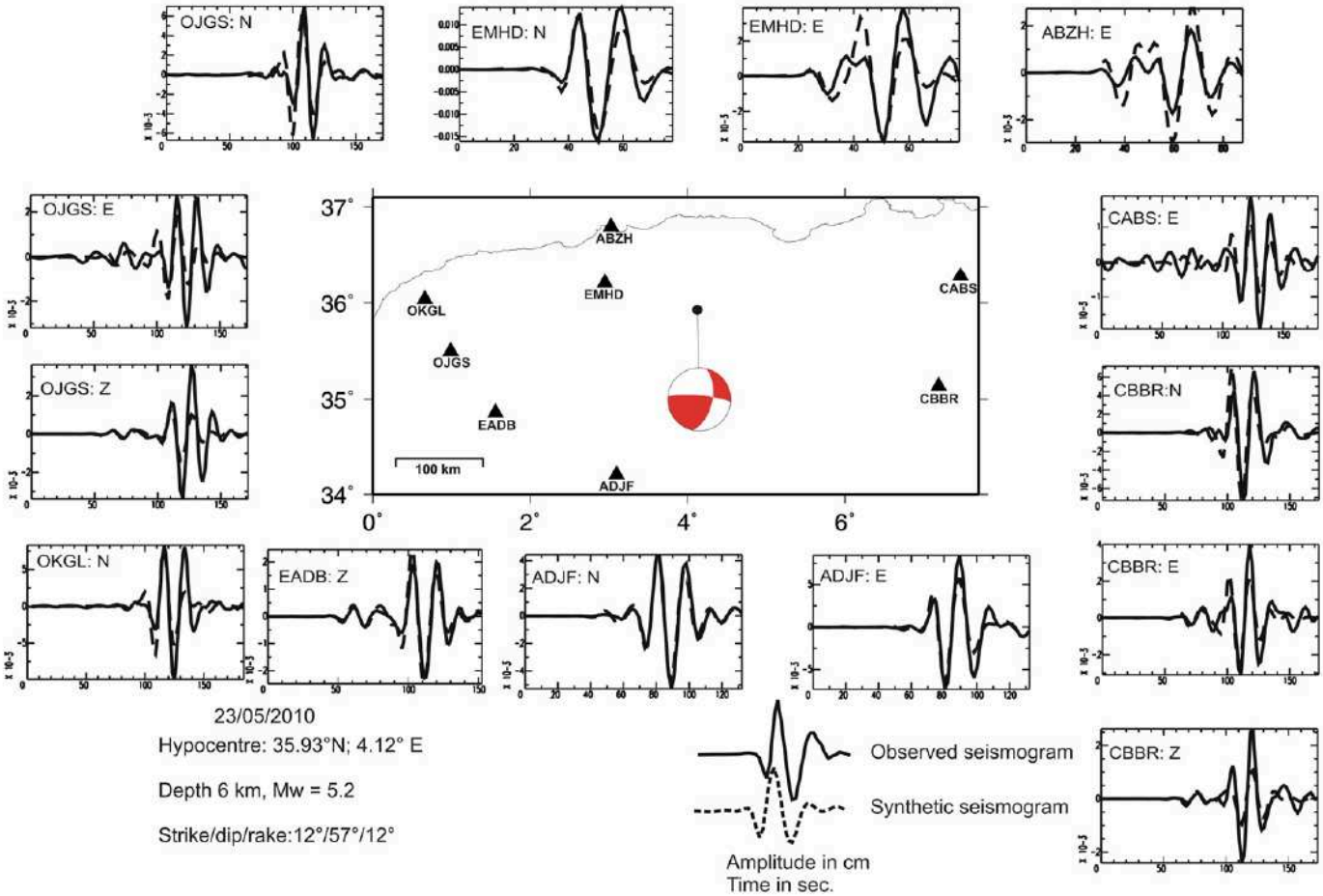

Fig. 7. Same as Fig. 5, but for the third shock of 23 May 2010 at 13 h 28 min GMT.

6. Stress inversion

We used the GRIDSIM_STRESS algorithm (Delouis et al., 2002) to calculate the orientation of the main axes of the stress tensor ellipsoid. The GRIDSIM_STRESS algorithm is used to determine the orientation of the main axes of the stress tensor ellipsoid and the stress shape factor. The criterion for selecting the best stress tensor is to minimise the angular deviations between observed and computed slip vectors on the nodal planes of the focal mechanisms. A score is defined, which is the average of the scalar products of the observed and computed slip vectors. A scalar product of 1 indicates that the observed and computed slip vectors are identical, and a value of 0 indicates that the slip vectors are perpendicular; a negative scalar product can occur when the angular difference is >90°. For the given focal mechanism and trial stress tensor, both nodal planes are tested and the plane providing the largest scalar product is retained when computing the total score. The stress tensor shape factor R is defined as in Rivera and Cisternas (1990), where the domains ($R < 0$), ($0 < R < 1$), and ($R > 1$) represent compressional, strike-slip, and extensional stress regimes, respectively. The shape factor $R = (\sigma_z - \sigma_x)/(\sigma_y - \sigma_x)$, where $\sigma_y \geq \sigma_x$ and σ_z is the stress axis closest to vertical. The shape factor R, which varies between $-\infty$ and $+\infty$, gives information about the stress regime. The connections between σ_x, σ_y, and σ_z and the eigenvalues σ_1, σ_2, and σ_3, respectively, can be easily established for any R value (Delouis et al., 1993). For example if $R \in |0, 1|$, then $\sigma_y > \sigma_z > \sigma_x$ and $\sigma_y = \sigma_1$, $\sigma_z = \sigma_2$, and $\sigma_x = \sigma_3$. The orientation of the three stress axes and the shape factor R are explored using a combination of a grid search and simulated annealing. As the value of a single parameter is explored at a constant increment, the parameter is held fixed in the simulated annealing algorithm whilst the other parameters are free to vary within their physically possible ranges. The orientation of the three stress axes in space is represented by three rotations corresponding to the Eulerian angles. With the Eulerian angles (Φ, θ, Ψ) (Goldstein, 1959), any position in a space of three orthogonal axes can be obtained from an initial position (here, we start from the geographical position N, E, Z), through three rotations only.

We used (1) the focal mechanisms available in the region calculated by different authors (Hatzfeld, 1978; This study) (Fig. 2, Table 2); (2) the focal mechanisms of the three main shocks calculated in this study; and (3) the focal mechanisms of the aftershocks calculated by Yelles-Chaouche et al. (2014) (Fig. 8, Table 2). From the plot of the stress axes solutions (Fig. 9), we estimated an uncertainty of $\pm 10°$ in the azimuth and dip of the σ_1 axis. The uncertainty is graphically estimated from the width of the clusters of solutions in Fig. 9. Uncertainties on the σ_2 and σ_3 axes appear much larger, but are due to the symmetry of the stress ellipsoid for $R = 0$, corresponding to $\sigma_2 = \sigma_3$. In fact, both σ_2 and σ_3 are located in a near-vertical plane whose azimuth is N70° $\pm$ 10°. The best solutions indicate that the axis of maximum compressive stress (σ_1) is approximately horizontal (plunge of 22°) and trends N340° $\pm$ 10°, the intermediate stress axis (σ_2) is near vertical and trends N200° $\pm$ 10°, and the axis of minimum compressive stress (σ_3; the extensional axis in the deviator definition) is horizontal and trends N77° $\pm$ 10°.

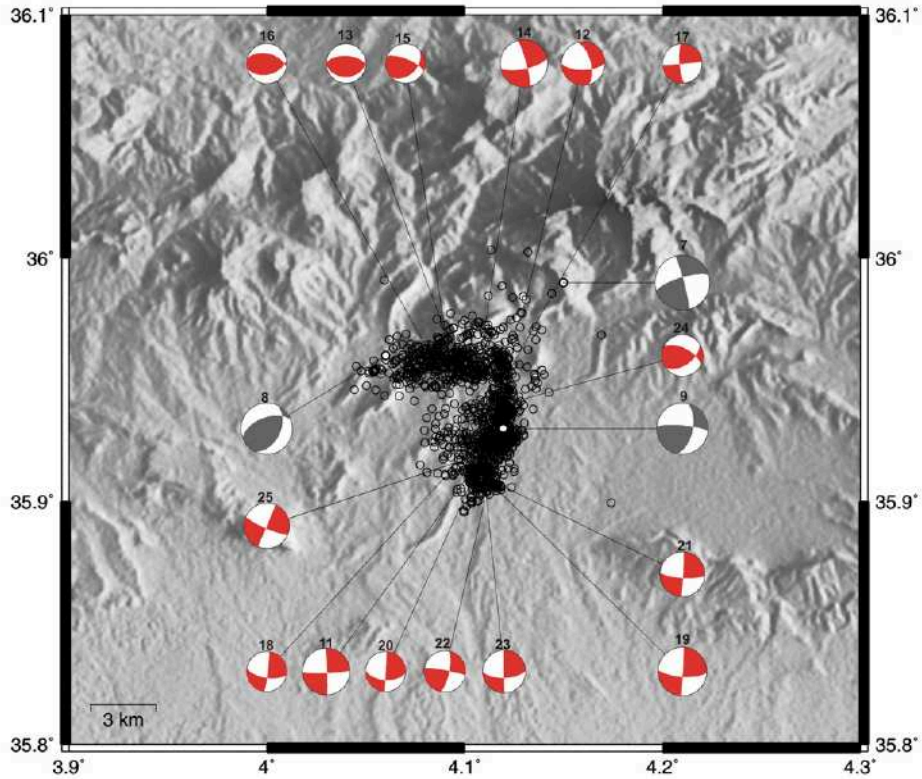

Fig. 8. Focal mechanisms in red calculated by Yelles-Chaouche et al. (2014) using the polarity of the first P-wave arrival. Focal mechanisms in grey were calculated in this study by waveform modelling. Open circles: 1403 aftershocks relocated by Yelles-Chaouche et al. (2014). White dots show the locations of the three important shocks.

On a diagram of score versus shape factor *R*, we note that the best scores correspond to uniaxial compression, at the frontier of the strike-slip and compressional regimes ($R = 0.0$). The score is a measure of the agreement between observed and computed slip vectors on the nodal planes of the focal mechanisms. The ambiguity in the stress regime is related to the possibility that both σ_2 and σ_3 are either near-vertical or

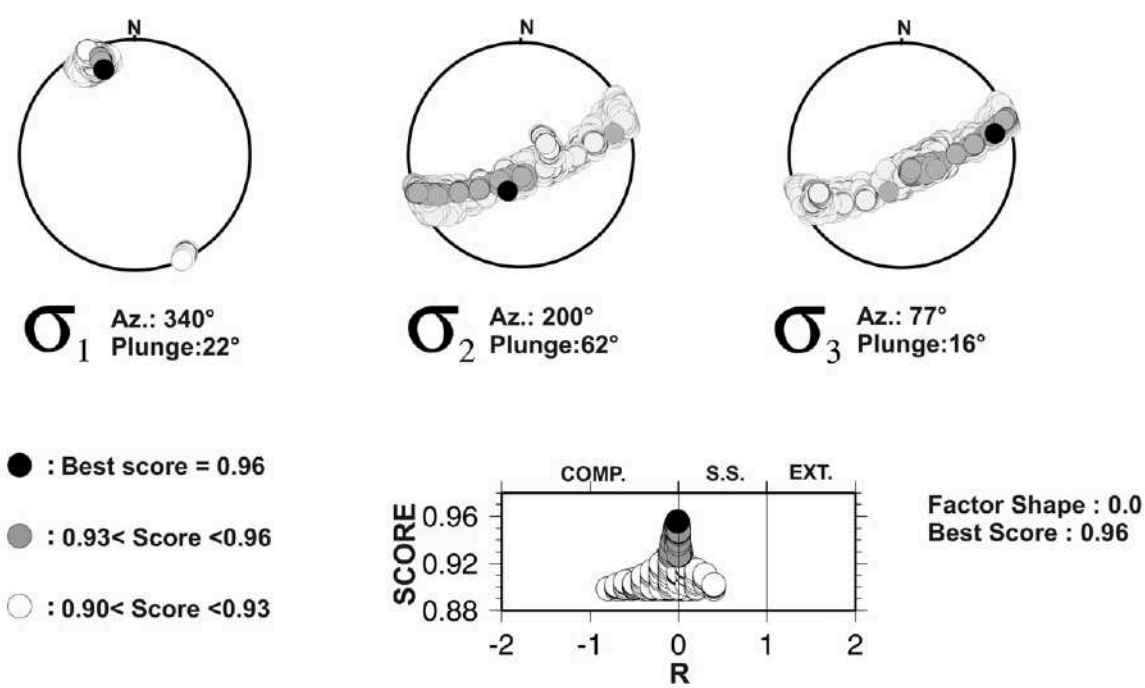

Fig. 9. Solutions for σ_1, σ_2 and σ_3 are plotted in three lower-hemisphere equal-area projections. The azimuth (az.) and plunge of the principal stress axes are indicated. The best solution (best score = 0.96) is represented by a black dot. The score is the average of the scalar products of observed and computed slip vectors. Solutions with slightly degraded scores (0.96–0.93) are shown by grey circles; we consider them as defining the confidence intervals. We also represent other degraded solutions with open circles, corresponding to scores of 0.90–0.93. Overall, plotting the range of solutions provides a view of the uncertainty in the positions of the stress axes and of the values of the stress shape factor *R*.

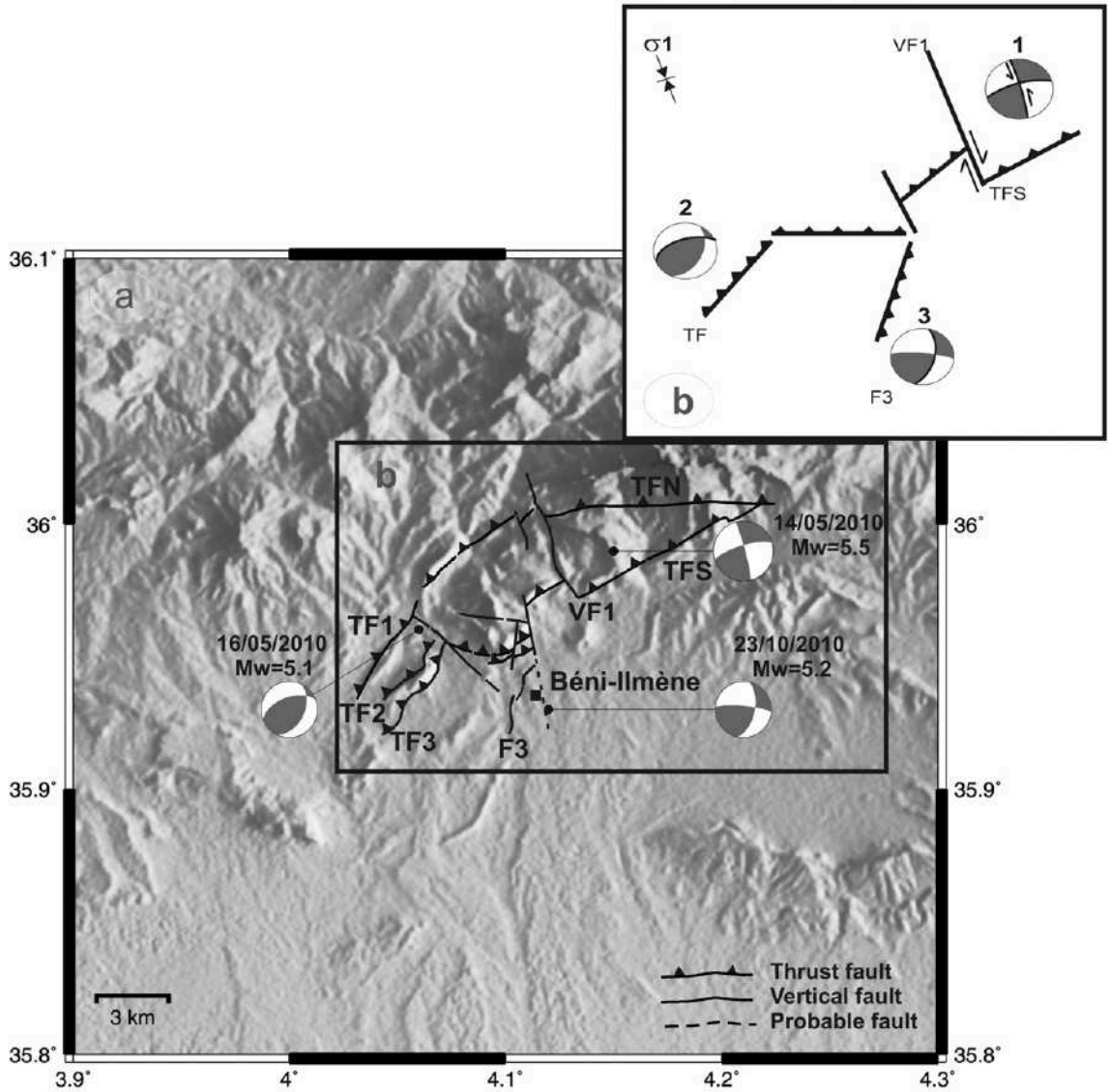

Fig. 10. (a) Map showing the principal faults in the epicentral area and the focal mechanisms of the three largest shocks. (b) Schematic drawing of the main faults and their relationship to the three main shocks in the sequence. The largest shock (shock 1) is related to either fault VF1 or fault TFS; shock 2 is related to fault TF, and the third shock is related to fault F3. The nodal planes assumed to correspond to the identified faults are drawn with thicker black lines (not discriminated for shock 1). The orientation of the principal stress axis σ1 is indicated in the upper left corner.

near-horizontal, as shown in the stereographic plots, whilst the σ_1 axis is best constrained (Fig. 9).

To evaluate whether older data, potentially with larger errors, could bias the results, we carried out two additional stress inversions, one with 7 focal mechanisms available before the 2010 sequence, and one with 18 focal mechanisms for the 2010 sequence itself. The results are very similar to those presented in Fig. 9 for the complete set of 25 mechanisms. We conclude that data quality is not biasing the results, and that the stress field before the 2010 seismic sequence was the same as that during the 2010 seismic sequence.

7. Discussion and conclusion

The Biban–Hodna region is seismogenically active, and the area has experienced many moderate earthquakes. The maximum Io observed in the region is Io VIII–IX (Roussel, 1973; Bezzeghoud et al., 1996; Boughacha et al., 2004). However, no major earthquakes have been recorded in this region of Algeria during the instrumental era. This study represents the first time that near-field seismic broadband records in this area, obtained by the ADSN, have been examined in detail; in addition, a temporary seismic network (short period) was installed in the region to collect additional data (Yelles-Chaouche et al., 2014). The ADSN network allowed us to analyse in detail the seismic sources of the 2010 Beni-Ilmane seismic sequence by modelling broadband records in the near and regional fields.

As mentioned above, our work complements the study of Yelles-Chaouche et al. (2014). We re-computed FMs and moment magnitudes using a full waveform inversion approach; we inverted the stress tensor at a regional scale, and have given a detailed compilation of geological information (synthesised on a geological map), thus providing a clear perspective on the distribution of faults in the epicentral area. Finally, a new seismotectonic model was developed, taking into consideration the more detailed information available in this study.

The solutions obtained for the FMs of the three largest shocks in the sequence are similar to those provided by INGV, Swiss Seismological Service (ETHZ: http://www.seismo.ethz.ch/prod/tensors/mt_autom) and GCMT (Fig. 2, Table 3); however, given the moderate sizes of the events, the confirmation of the solutions with local data was considered important.

The FM solutions for the first and third shocks in this study are comparable to those computed by Yelles-Chaouche et al. (2014) (Fig. 2, Table 3); however, notable differences were found for the second shock. Our solution exhibits reverse motion along nodal planes oriented NE–SW and dipping 45°–55°, with a small component of strike-slip motion (Fig. 2, Table 2), whereas the solution of Yelles-Chaouche et al. (2014) is also reverse, but with nodal planes oriented E–W, with one steep (dip, 82°N) and the other shallow (dip, 9°S). Our solution is closer to the determinations of INGV, ETHZ, and GCMT.

Based on the aftershock distribution, Yelles-Chaouche et al. (2014) described two principal directions, with a cluster oriented NNE–SSW and containing the first and third shocks, and a second cluster oriented approximately E–W and containing the second shock. However, Yelles-Chaouche et al. (2014) did not examine in detail the connection of the clusters to geological structures. According to the geological and tectonic map in Fig. 4, the first shock on 14 May 2010 (Mw 5.5) was located east of vertical fault VF1 (oriented N340°). This fault was reported (Savornin, 1909, 1929; Kieken, 1961 and Baldini, 1966) as a right-lateral strike-slip fault (Fig. 10a and b), located south of thrust fault TFN (striking N270°) and north of thrust fault TFS (striking N240°). Aftershocks in the vicinity of the first shock are too scarce to delineate a preferred fault amongst VF1, TFN, and TFS.

The FM of the first shock indicates strike-slip faulting with a small reverse component. The right-lateral fault plane is striking N254° and the left-lateral plane is striking N345°. The right-lateral plane (N254°) is parallel to the reverse fault TFS (striking N240° ± 14°), whereas the left-lateral plane (N345°) is parallel to the strike slip fault VF1 (striking N340° ± 5°). The two nodal plane solutions can be interpreted in terms of geological and tectonic observations. The right-lateral plane (striking N254°) can be correlated with fault TFS (N240°), and the left-lateral plane striking N345° can be correlated with vertical fault VF1. According to the geological map and cross-section (Fig. 4a and b), the autochthonous Miocene Ouenougha Formation is tectonically overlain by the Djebel El Khert anticline along a shallow-dipping (20°–30°) ramp-flat contact, with the dip of the ramp not exceeding 40°. The FM of the first shock indicates a dip of 74° for the nodal plane (striking N254°), which seems too steep by comparison with the geological interpretation of the faulted contact. On the other hand, the epicentre appears more compatible with fault TFS than with fault VF1; the epicentre is located ~3 km east of vertical fault VF1. However, a horizontal error (ERH) on the order of 3 km is given by CRAAG for the hypocentre location. Fault VF1 appears as a right-lateral fault on maps (Figs. 4, 10a, and b), which is incompatible with the FM. However, VF1 may be interpreted as a transfer fault, with a strike-slip sense depending on the relative rates of movement on present-day thrusts either side of the fault, which are unknown. It is therefore possible that VF1 is kinematically compatible with the left-lateral fault nodal plane striking N345°. All of these elements do not strictly forbid any of the nodal planes from corresponding to the actual rupture plane. At this stage of the study, we cannot determine with certainty which plane was activated during the mainshock, or clearly favour a correspondence with either fault TFS or fault VF1.

The second shock on 16 May 2010 (Mw 5.1) was located near the western termination of the approximately E–W aftershock cluster (Fig. 8). Its focal mechanism indicates reverse faulting, with the first plane striking N25° and the second one N250°, parallel to thrust faults TF1, TF2, and TF3 (Fig. 10a and b). These three faults are located close to one another and are parallel (all striking N235° ± 15°). The epicentre of the second important shock is located 0.8 km and 1.6 km from fault traces TF2 and TF3, respectively. If the hypocentre was located on one of these faults, it should be at a depth of 1 km (TF2) or 2 km (TF3); however, we located the depth of the hypocentre at 6 km. On the other hand, taking into account reasonable amounts of error (of several kilometres for the epicentre and depth), the event might be related to faults TF2 or TF3, and the activated nodal plane would be the one striking, N250°; dipping 55° NW.

The third shock on 23 May 2010 (Mw 5.2) was located at the centre of the approximately N–S aftershock cluster (Fig. 8). The FM shows strike-slip faulting with a small reverse component, with the first plane striking N275° and the second one N12° and dipping 57° to the east. The epicentre is located east of the fault trace (trace F3, striking N20°; Fig. 10a and b). The southern aftershock cluster is striking N18°, parallel to structure F3. We suggest that the event was related to fault F3 and that the second nodal plane (strike, N12°; dip, 57°E) was the activated plane. Fig. 10b summarises the relationships between the three shocks and the known faults.

The regional stress tensor calculated in this study for the Hodna–Biban region, between longitudes 4°E and 5.5°E, indicates either a strike-slip or a compressional stress regime, depending on whether the axis closer to the vertical is σ_2 or σ_3. The trend of σ_1 (N340°), which is well constrained and is identical to the directions found from the inversion of 15 FMs in the western part of northern Algeria between longitudes 0.75°E and 2°E (σ_1, striking N325°; Beldjoudi et al., 2012) and the inversion of 13 FMs in eastern Algeria between longitudes 4.5°E and 8°E (σ1, striking N325°; Beldjoudi et al., 2009).

The Beni-Ilmane seismic sequence exhibits complex characteristics and involved the reactivation of a faulted fold system along three distinct faults associated with the three largest shocks (of M > 5). The relationship between the third shock (23 May 2010; Mw 5.2) and the approximately N–S cluster of smaller aftershocks (M < 5) is particularly striking. The approximately E–W cluster of aftershocks can be partially related to the second large shock (16 May; Mw 5.1), as the cluster is oriented NE–SW near its western termination (Fig. 8). The reason for the absence of aftershocks directly related to the first (14 May; Mw 5.5) and largest shocks remains unknown.

In conclusion, this study has introduced waveform modelling in the near-field for a better characterisation of the seismic sources in Algeria and provided a better determination of the stress tensor in the Hodna–Biban region. It also points to the presence of active structures with a length of about 15 km where events of magnitude > 5.5 can be expected to occur producing strong damage. Through this study we confirm that the Hodna–Biban region is an active one where events with a high magnitude (>5.5) can occur.

Acknowledgements

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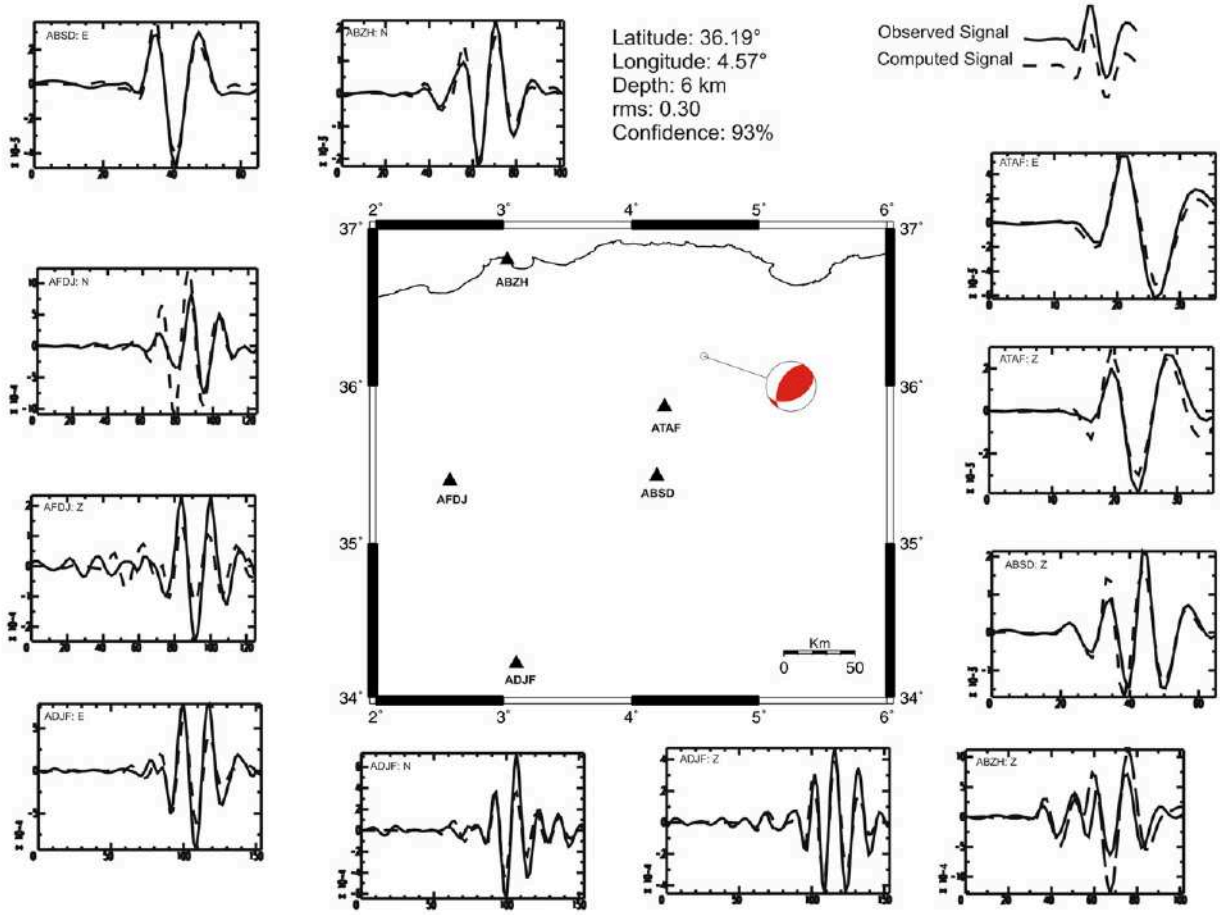
Appendix B. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.tecto.2015.12.021.

Appendix A

Same as Fig. 5 but for the Bordj Bou Arreridj events of 5 May 2012 (Mw 4.9).



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3.3. Séisme de Bordj-Menaïel 2014, Mw 4.1

Caractéristiques de la source sismique

Un séisme modéré a frappé la région épicentrale de Boumerdes-Zemmouri le 22 février 2014 à 19h30 mn (GMT). Le séisme a été localisé 12 km au SE de l'épicentre du séisme de Boumerdes-Zemmouri du 21 mai 2003 (**Figure 3-2**). La magnitude de moment de l'événement Mw = 4.1. L'inversion des formes d'ondes a donné un mécanisme au foyer en décrochement senestre pour le plan orienté NS et dextre pour le plan orienté EW. La solution obtenue est bien contrainte avec un degré de confidence (Confidence Index : CI) égal à 78% et une RMS égale à 0.56 pour la profondeur de 7 km (**Figure 3-3a et 3-3b**). Nous cherchons à obtenir une solution qui ait à la fois une fonction erreur quadratique d'ajustement des formes d'ondes (RMS) la plus petite possible, et un indice de confiance le plus élevé possible pour s'assurer que la solution est bien contrainte. 14 composantes ont été utilisées dans l'inversion (**Figure 3-4**). Au cours de l'inversion (FMNEAR), certaines composantes mal modélisées sont éliminées, souvent en raison de la difficulté de trouver un filtrage satisfaisant. Pour plus de détail concernant cet événement le papier intégral est dans la partie annexe de la thèse.

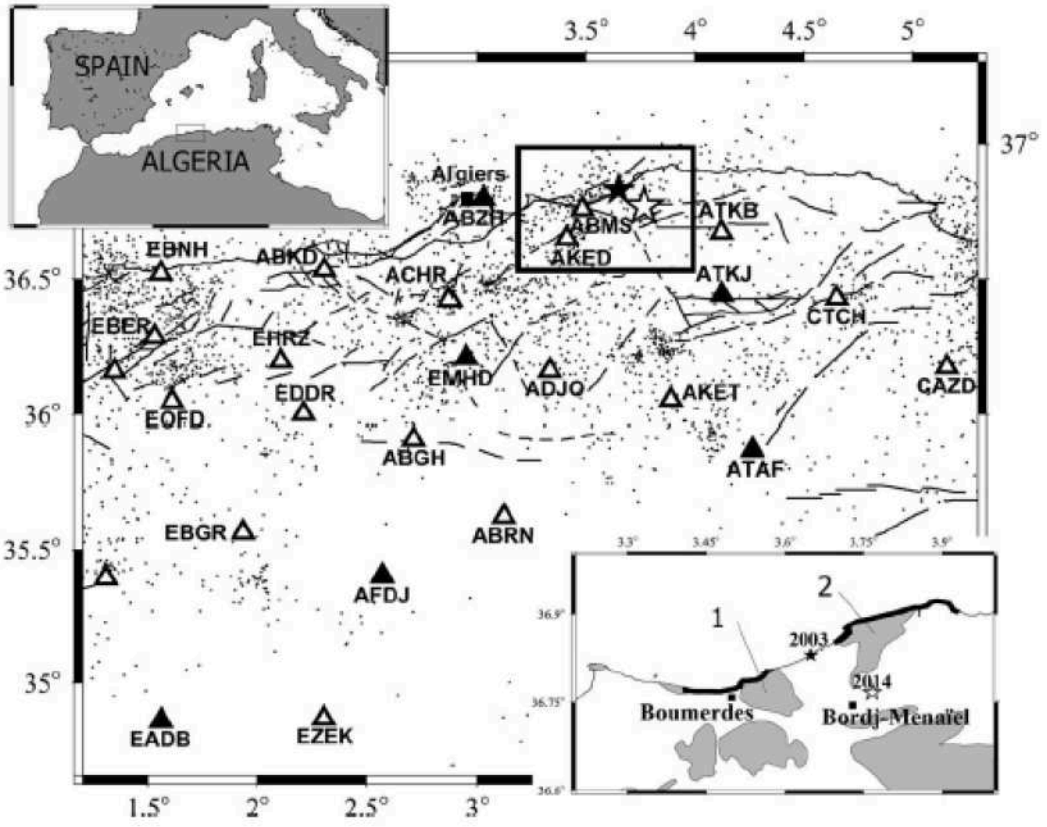


Figure 3-2 : *La localisation du séisme du 22 février 2014 est donnée par l'étoile blanche. L'étoile noire est l'épicentre du séisme de Boumerdes-Zemmouri du 21 mai 2003. Les triangles noirs montrent les stations large bande. Les triangles blancs montrent les stations courtes périodes. La sismicité depuis 1992 jusqu'à mars 2014 est représentée par les points noirs. Les traits représentent les failles actives reportées par Guemmache (2010), (D'après Semmane, Benabdeloued, Beldjoudi et al., 2015). Les lignes épaisses dans l'encadré en bas montrent le soulèvement côtier suite au séisme de Boumerdes-zemmouri de 2003. Les surfaces en gris sont celles du socle et 1 et 2 correspondent aux régions des deux aspérités trouvées dans les modèles de glissement proposés (Semmane et al., 2015).*

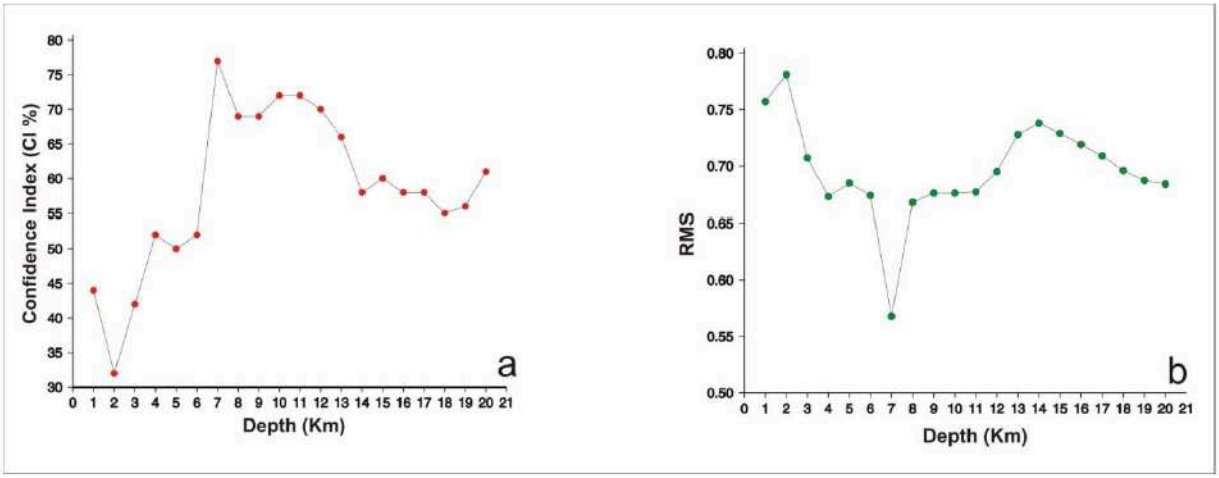


Figure 3-3 : *Qualité de la solution obtenue. a) Degré de confidence en fonction de la profondeur. b) RMS en fonction de la profondeur.*

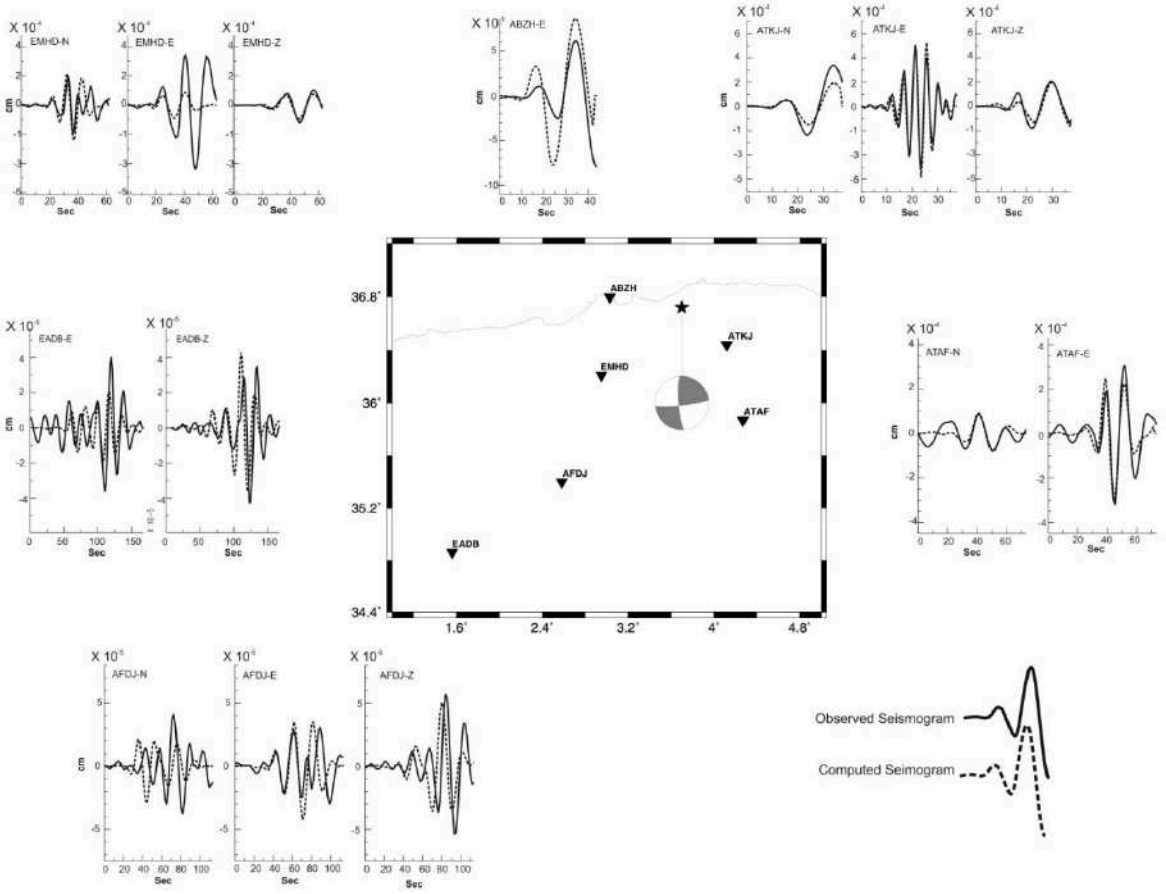


Figure 3-4 : *mécanisme au foyer obtenu par inversion des formes d'ondes sismiques de l'événement de Bordj-Ménaïel du 22/02/2014 (Algorithme FMNEAR, Delouis, 2014). En triangles noirs les stations large bande utilisées dans l'inversion. L'étoile montre l'épicentre. Les traits continus représentent les sismogrammes en déplacement observés et en traits discontinus les sismogrammes calculés (synthétiques).*

3.4. Séisme de Mw 5.0 Hammam Melouane, 2013

Caractéristiques de la source sismique

Le 17 juillet 2013, un séisme s'est produit tout près d'Alger sur la bordure sud de la Mitidja plus exactement à 30 km au sud d'Alger et à environ 50 km de l'Epicentre de Boumerdes-Zemmouri (**Figure 3-5**). De Magnitude Mw = 5.0, le séisme a causé une panique au sein de la population des villes d'Alger, Blida et Boumerdes. Heureusement il n'a pas fait de victimes à

part quelques dégâts sur des maisons et des infrastructures publiques dans la localité de Hammam Melouane (Yelles-Chaouche, Haned, Aidi, Beldjoudi et al., 2017).

L'inversion des formes d'ondes de ce séisme, montre que la source sismique est une faille décrochante, senestre pour le plan NS ou dextre pour le plan EW (**Figure 3-6**). La meilleure solution est obtenue pour une profondeur de 7 km avec un index de confidence (CI) de 80% et une RMS=0.511 (**Figure 3-7a et 3-7b**). Pour plus de détail voir le papier intégral en annexe.

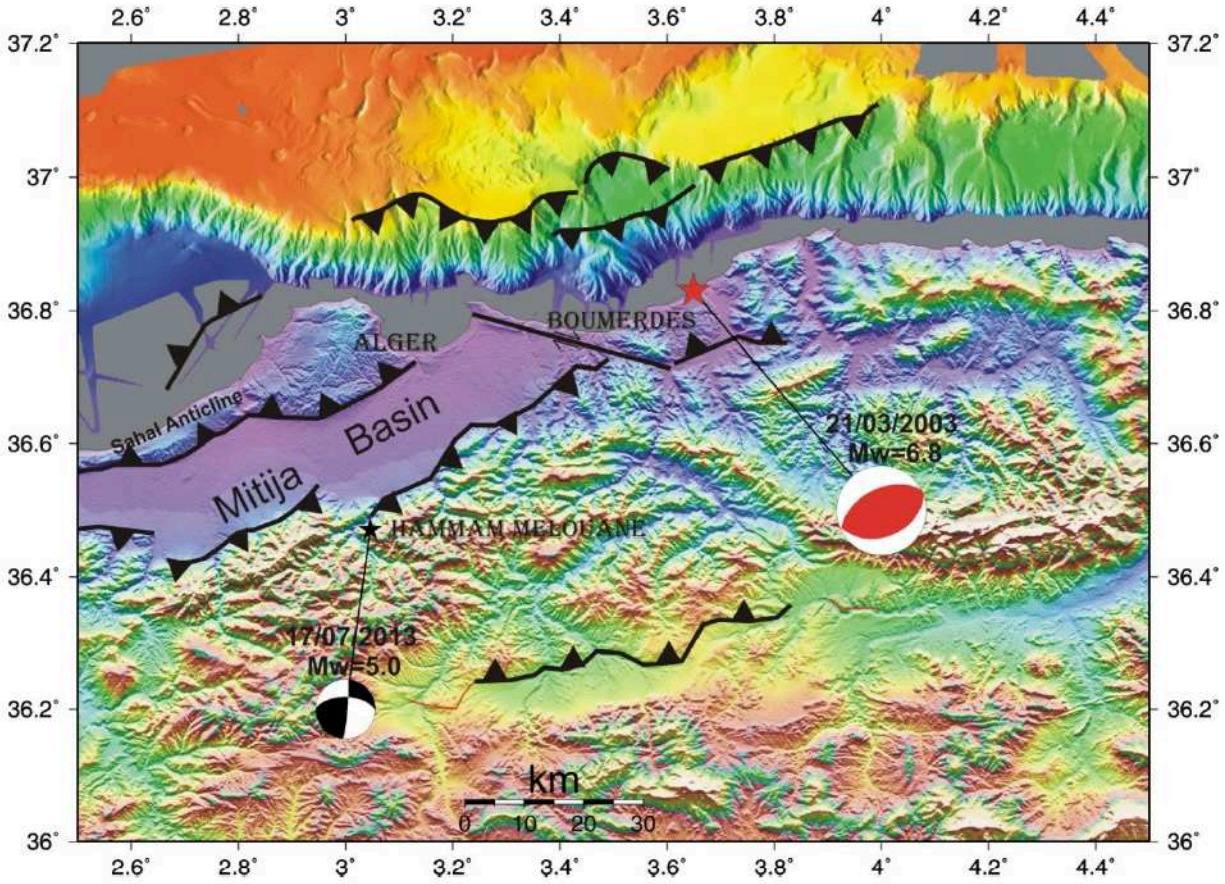


Figure 3-5 : *carte qui montre la localisation du choc du 17/07/2013 (étoile noire) avec le mécanisme au foyer obtenu par modélisation des formes d'ondes. L'étoile rouge montre l'épicentre du séisme de Boumerdes de 2003. Les principales failles actives dans la région de la Mitidja et ses environs sont représentées par les traits en noir, les triangles montrent la direction du pendage.*

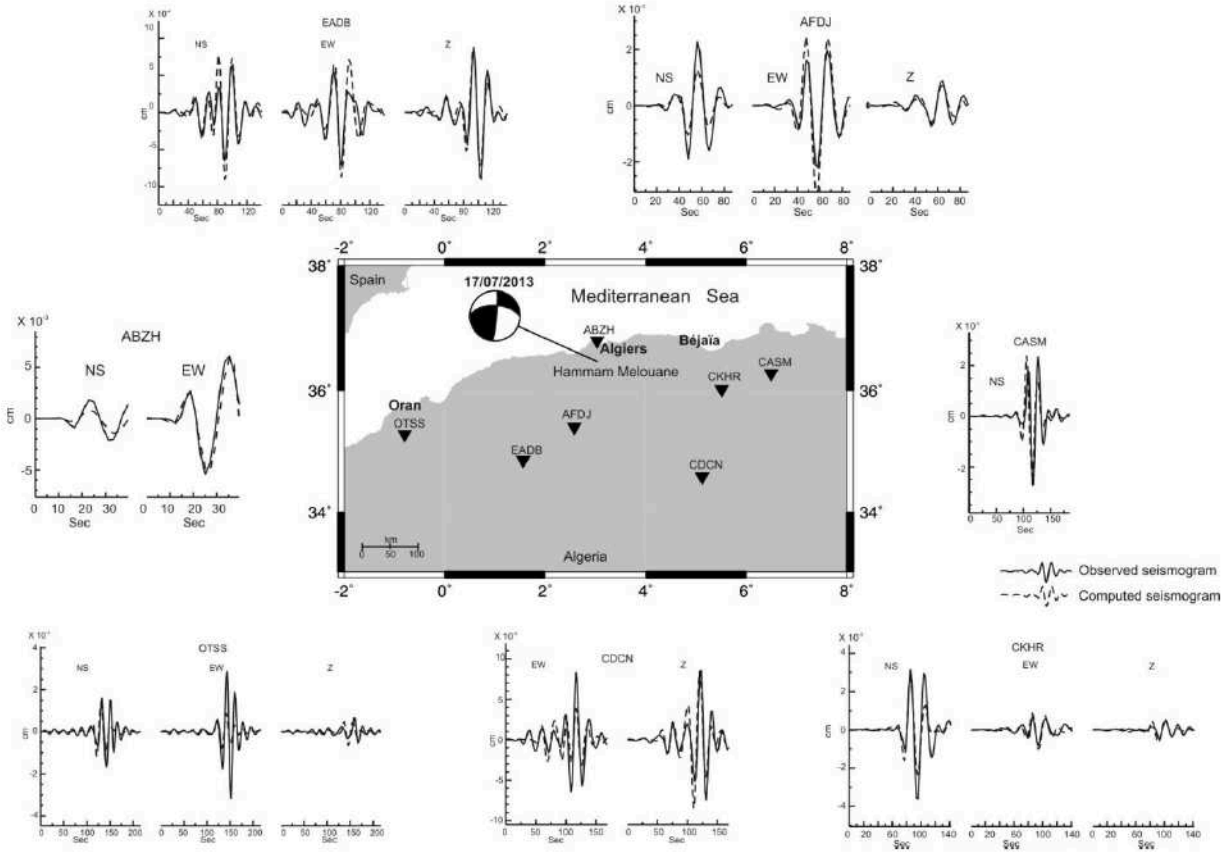
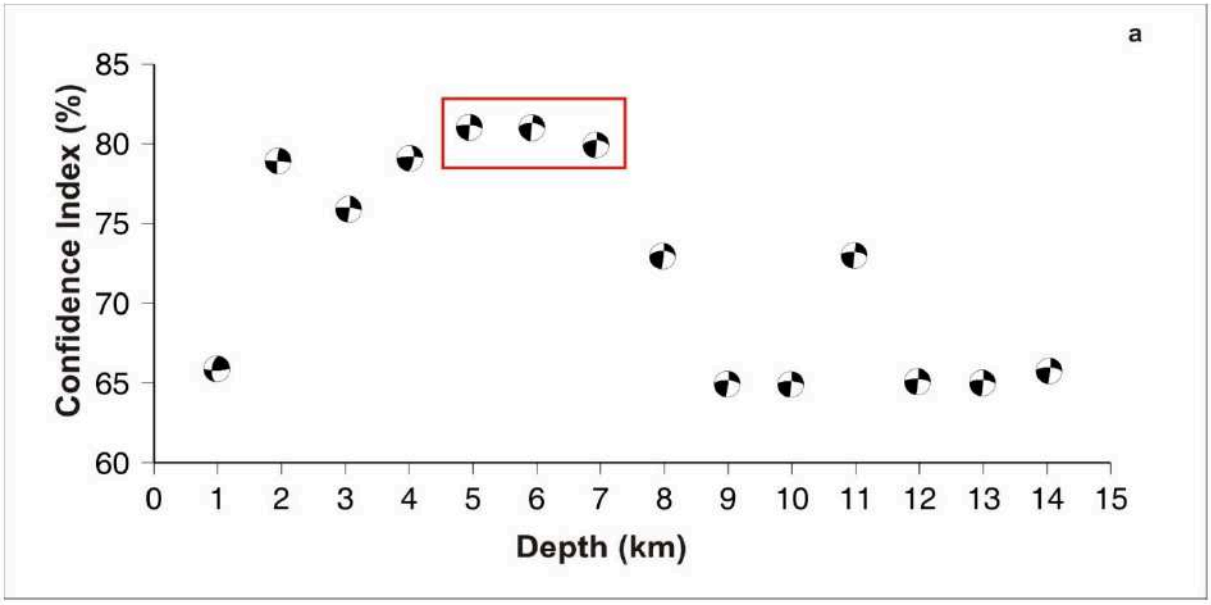


Figure 3-6 : *Modélisation des formes d'ondes sismiques du séisme de Hammam Melouane du 17/07/2013. Sur la carte nous représentons toutes les stations large bande utilisées dans l'inversion (Triangles inversés) ainsi que leurs codes, le mécanisme au foyer obtenu. Les formes d'ondes sont représentées par la superposition des traits continus (observés) et les traits discontinus (calculés : synthétiques).*



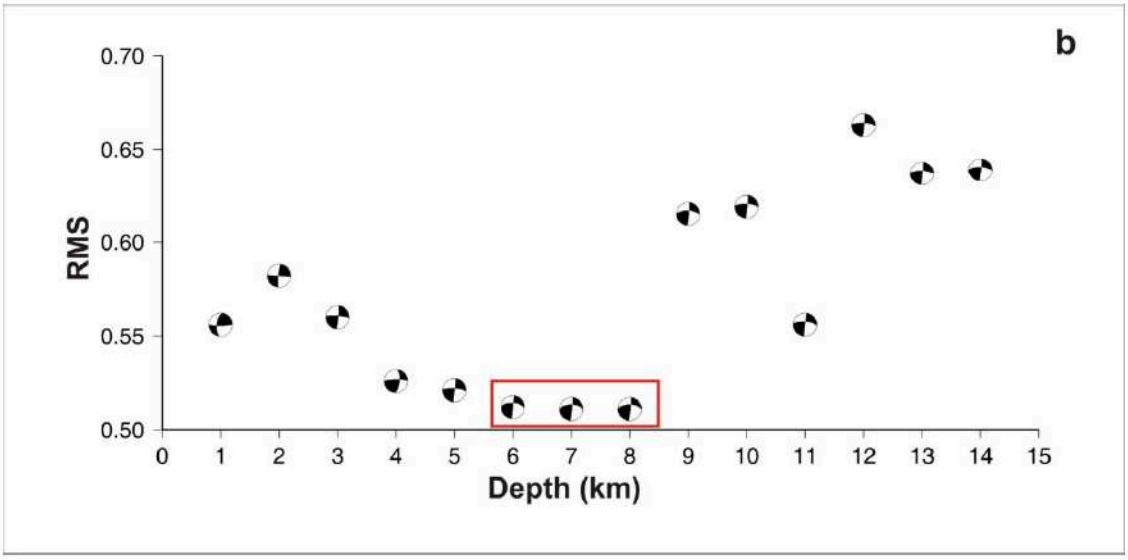


Figure 3-7 : *Qualité de la solution focale obtenue. a) Indice de confidence en fonction de la profondeur. b) RMS en fonction de la profondeur. Encadré en rouge, les meilleures solutions.*

3.5. Séisme de Biskra 2016, Mw 5.1

Caractéristiques de la source sismique

Le séisme de M'ziraa localisé à 350 km SE d'Alger et à 50 km à l'Est de la ville de Biskra, ville de l'Atlas saharien (le sud de l'Algérie). Ce séisme s'est produit le 18 novembre 2016 avec une magnitude de moment Mw = 5.1. Localisé à la latitude 34.84° et la longitude 6.28° et à une profondeur de 40 km (**Figure 3-8**). Le mécanisme au foyer a été calculé avec 24

composantes sismiques, avec un degré de confidence CI=78% et une RMS=0.651. Le mécanisme au foyer est celui d'une faille normale orientée EW avec une légère composante de décrochement pour une profondeur de 36 km. Quelques heures après une réplique de magnitude Mw=4.7 a secoué la région épicentrale aux coordonnées suivantes latitude 34.84° et longitude 6.29°. Un mécanisme au foyer en faille normale orientée EW a été obtenu par inversion des formes d'ondes (**Table 3-1**) avec un index de confidence CI= 81% et une RMS=0.496.

Evénements	Strike1	Dip1	Rake1	Strike2	Dip2	Rake2	CI(%)
Choc principal	75	50	-125	302.4	51.1	-55.7	78
Réplique	85.0	50.0	-115.5	301.8	46.4	-62.7	81

Table 3-1 : *paramètres de la source sismique du choc principal et de la réplique majeure. Les valeurs de strike, dip, rake sont en degrés.*

Le mécanisme au foyer obtenu est celui d'une faille normale orientée EO. La profondeur du séisme est relativement profonde par rapport aux séismes algériens. Ce séisme fera l'objet d'une étude plus détaillé afin de comprendre cette crise sismique survenue dans la région de transition entre l'Atlas saharien et la plateforme saharienne.

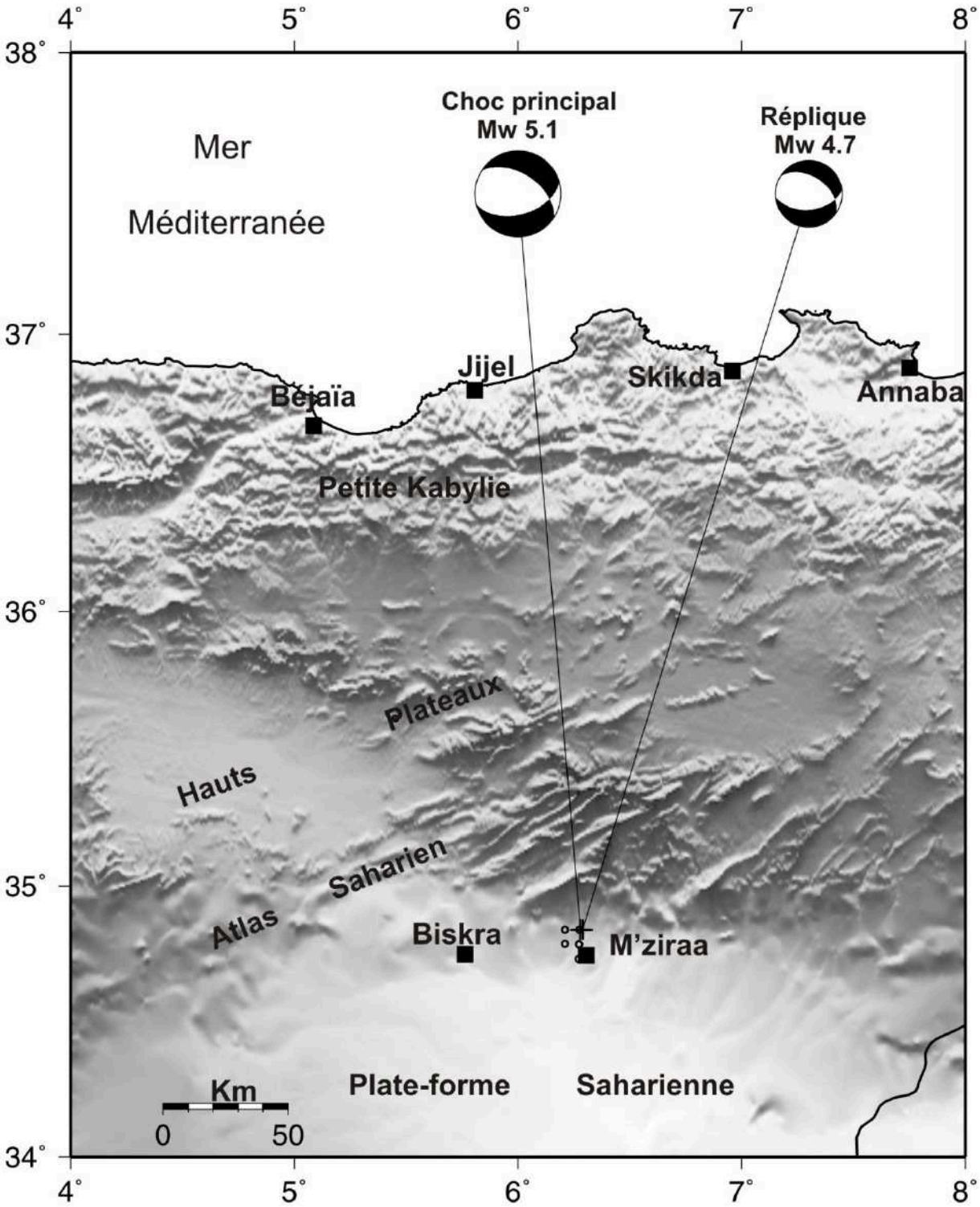


Figure 3-8 : *mécanismes au foyer du choc principal (Mw 5.1) et de la réplique principale (Mw 4.7) du 18/11/2016 dans la localité de M'ziraa de la région de Biskra. Les cercles vides montrent les quelques répliques qui ont suivi le choc principal. Les carrées noirs montrent la localisation des villes.*

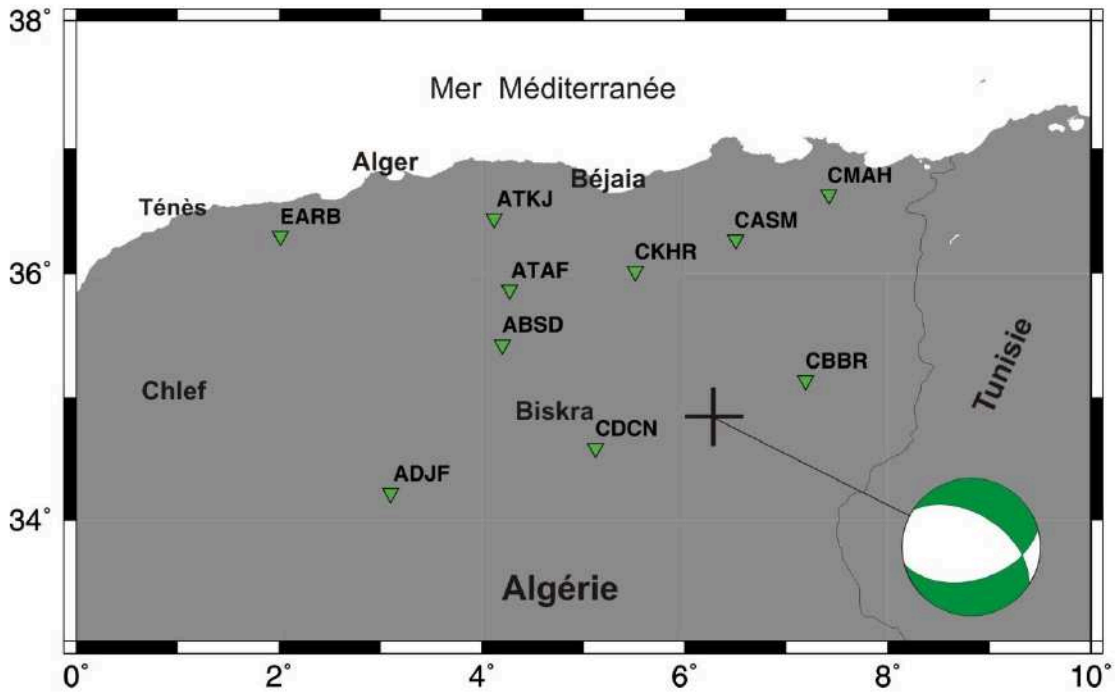


Figure 3-9 : *Stations large bande du réseau ADSN utilisées dans l'inversion des formes d'ondes sismiques et le mécanisme au foyer obtenu.*

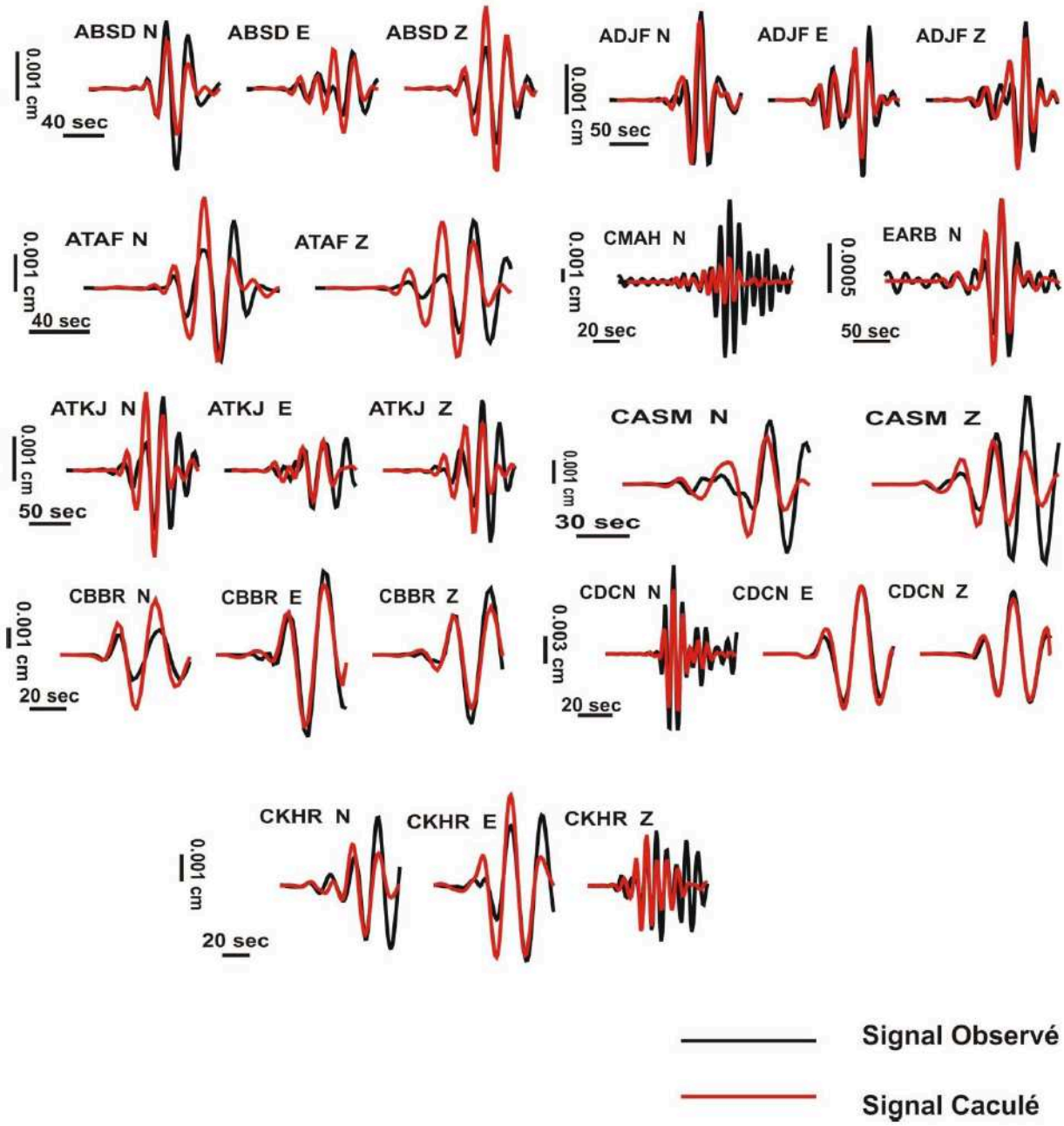


Figure 3-10 *: Modélisation des formes d'ondes du choc principal pour les stations mentionnées sur la* ***Figure 3-9*** *du choc principal.*

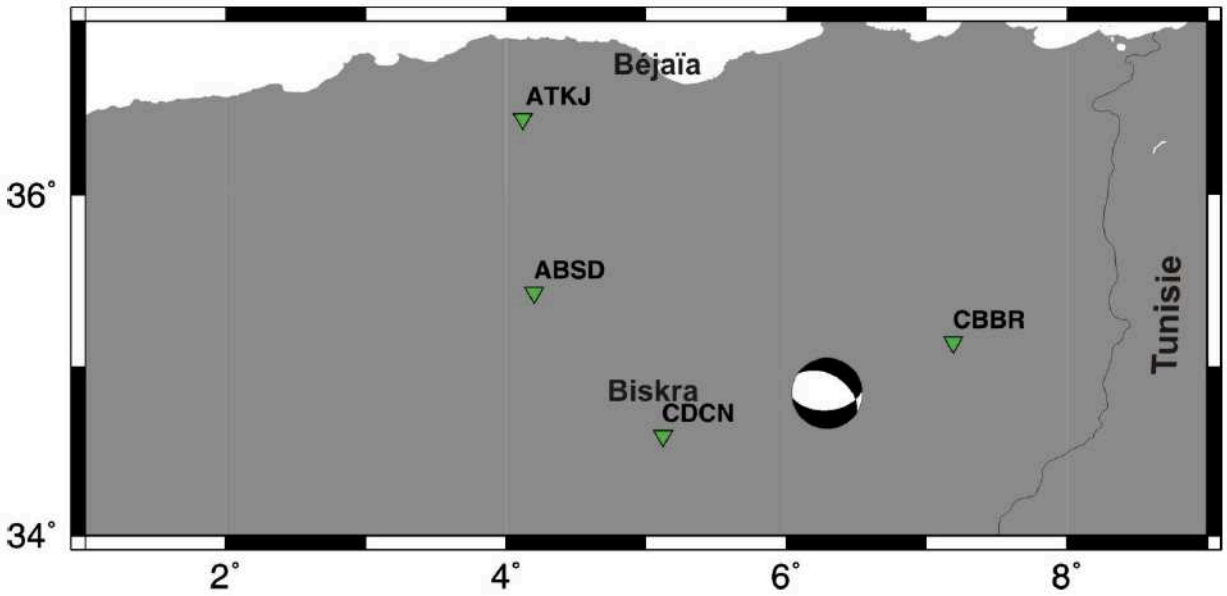


Figure 3-11 : *Stations large bande du réseau ADSN utilisées dans l'inversion des formes d'ondes sismiques et le mécanisme au foyer obtenu de la réplique.*

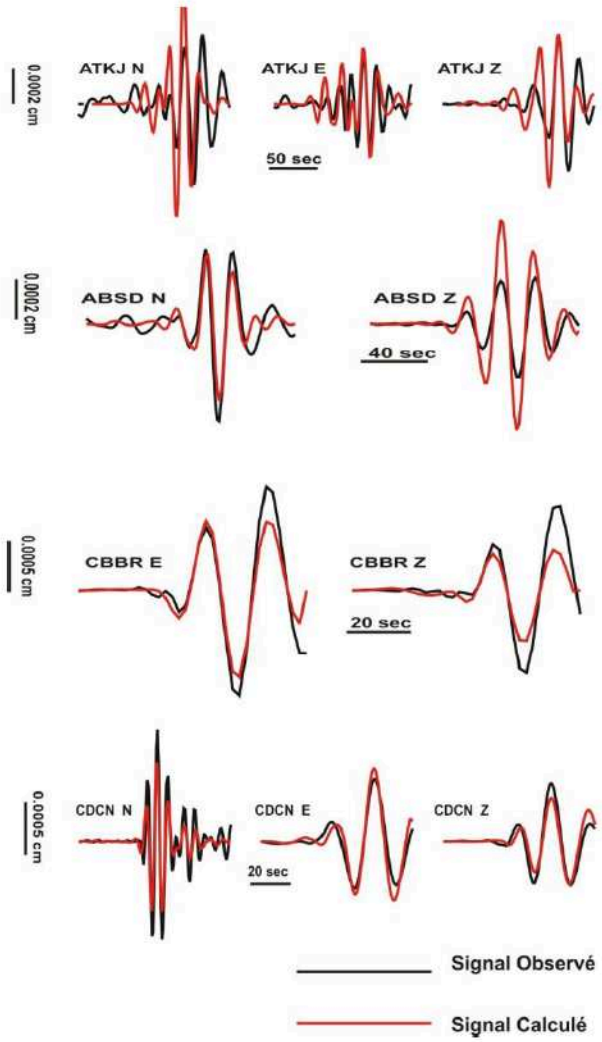


Figure 3-12 : *Modélisation des formes d'ondes de la réplique pour les stations mentionnées sur la* ***Figure 3-11****.*

Chapitre 4: Le séisme de Boumerdes-Zemmouri de 2003

Nous présentons ce chapitre sous la forme d'un article en préparation.

Reassessing the rupture process of the 2003 Boumerdes-Zemmouri earthquake (Mw 6.8, northern Algeria) using teleseismic, strong motion, InSAR, GPS, and coastal uplift data

Introduction

The Boumerdes-Zemmouri earthquake occurred ~50 km East of Algiers on May 21, 2003, at 19h44 (UTC). It is the second strong earthquake that struck northern Algeria, after the El Asnam earthquake (Ms=7.3) on 10 October 1980. The Boumerdes-Zemmouri event was the subject of many studies, among which several focused on the seismic source (Yelles et al., 2004; Meghraoui et al., 2004; Delouis et al., 2004; Semmane et al., 2005; Belabbes et al., 2009), or on the aftershock sequence (Bounif et al., 2004; Ayadi et al., 2008; Ouyed et al., 2011; Kherroubi et al., 2017). The 2003 rupture developed below the coastal area, partly under the sea, and Deverchère et al. (2005) described active structures offshore which may have been activated during the 2003 earthquake.

Regarding the analysis of the rupture process of the 2003 earthquake, different kinds of observations and measurements have been inverted jointly, notably different combination of purely static displacement data among GPS, InSAR (Synthetic Aperture Radar Interferograms) and coastal uplift, in Yelles et al. (2004), Meghraoui et al., 2004, and Belabbes et al., 2009. Two studies combined seismological and static displacement data, Delouis et al. (2004) and Semmane et al. (2005). However all the available kind of data were never inverted altogether. In particular, strong-motion, teleseismic, and InSAR data were never combined.

The motivation of this work is mainly twofold; i) combine for the first time a more comprehensive set of data, including teleseismic, strong motion, GPS, InSAR, and coastal uplift, in order to constrain in the best possible way the rupture model; ii) explore for the first time the possible coseismic activation of faults described offshore in Deverchère et al. (2005). Mahsas et al. (2008) proposed that post-seismic slip inferred from GPS measurements occurred between 0 and 5 km depth near the top of the coseismic rupture. These authors

suggest that the post-seismic deformation may have involved slip on the shallow flat and ramp, and folding of the soft sediments, described in Deverchère et al. (2005).

In this study, we explore kinematic rupture models of increasing complexity, taking into account previously published models and structural complexities proposed in the above mentioned studies. In order to define an optimal model, we will assess to which point model complexity is justified by the data. Indeed, most of the available data in the near source area provide only a partial coverage of the rupture zone. GPS stations are confined to the area between Algiers and Zemmouri, only three strong motion stations located within 50 km epicentral distance were exploitable, the InSAR data present many discontinuities (loss of coherency), and coastal uplift observations are restricted to an almost straight line oriented WSW-ENE. Therefore, constraining the slip distribution is a real challenge and combining the different datasets having complementary coverage is the best approach that we can undertake. Nonetheless, we expect that it will be particularly difficult to constrain the rupture on the farthest offshore area.

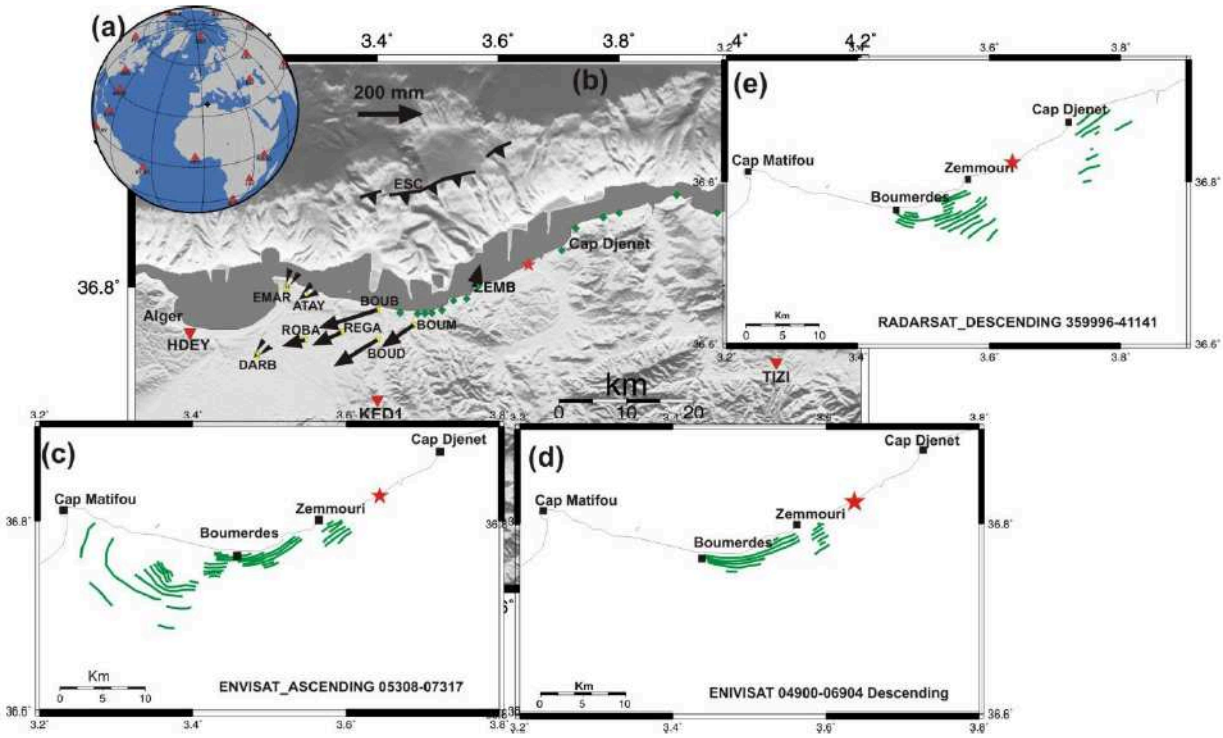


Figure 1 : *Map showing the location of the datasets used in the joint inversion. a) Red triangles are teleseismic stations. Black cross is the epicentre location. b) Yellow squares are the GPS stations location. Black arrows are the horizontal coseismic displacement vectors recorded at these stations. Green diamonds show the coastal uplift (CU) measuring points. Red inverted triangles are the strong-motion stations. Red star is the epicentre location of Boumerdes-Zemmouri earthquake on 21 May*

2003. ESC: offshore escarpments B1 and B2 from Deverchère et al. (2005). c)-d)-e) Green lines are digitized fringes from InSAR image: ENVISAT-Ascending, ENVISAT-Descending and RADARSAT-Descending, respectively.

Data

Strong motion (SM)

We use three strong-motion accelerograms from the Algerian earthquake engineering research Centre (CGS: Centre de Génie Parasismique) (**Figure 1b**). Two of them (TIZI and HDEY) are analogic (SMA-1 instruments) and KED1 is a digital one (Altus, Etna). The stations were not linked to GPS time, they worked on triggered mode. Unfortunately, the first P wave was lost at all three stations (KED1, HDEY and TIZI, **Figure 1a and 1b**). We estimated the arrival time of the first P wave at each of the three stations by conducting a series of strong motion inversions. Since this estimation cannot be very accurate, we will include in the inversion procedure the possibility of a time shift by up to 2s of the seismic traces by performing a correlation between the observed and computed signals. Accelerograms are integrated twice to obtain displacement and bandpassed from 0.1 Hz to 0.25 Hz. The EW component of TIZI station is affected by another problem, a discontinuity in the signal (**Figure 1c**), which could not be corrected. This component was hence excluded from the inversion by assigning it a zero weight.

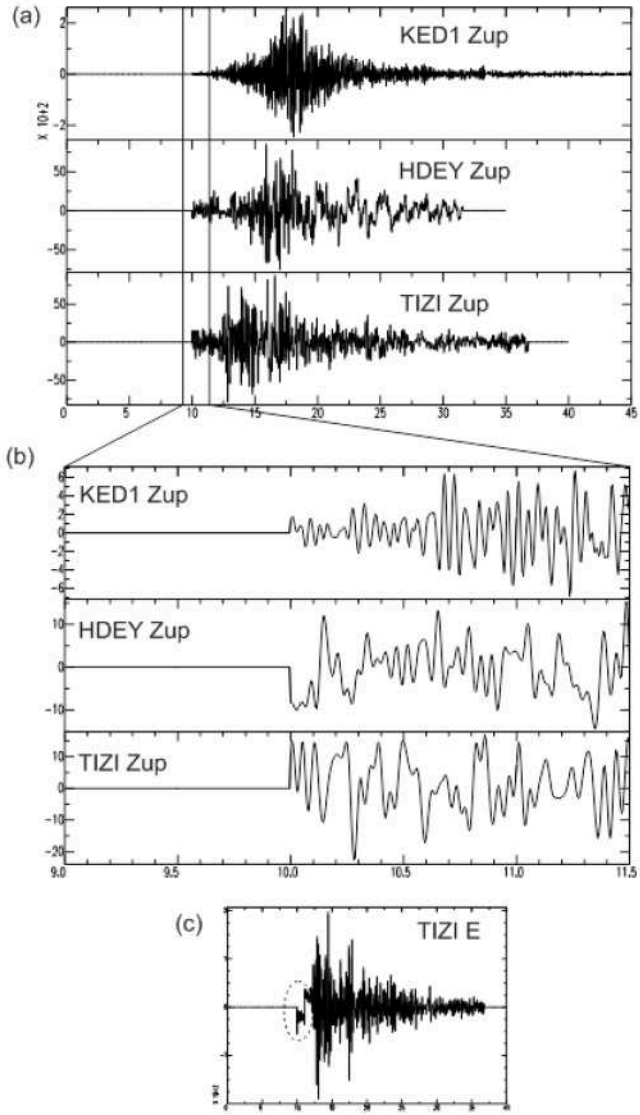


Figure 2: *particularity of the strong motion seismograms of the Boumerdes mainshock. a) Complete acceleration time series for the vertical component of KED1, HDEY and TIZI stations. The three traces are aligned to start at t=10s for the purpose of illustration. b) Zoom on the initial part of the same seismograms. Original seismograms had no absolute time reference and started were time is equal to 10s on the graph. We completed the seismogram before original start (between 9 and 10s on the graph) with zeroes. As can be seen, original seismograms started abruptly meaning that the initial P wave was lost. C) Complete acceleration time series for the East component of TIZI station. We see that within a few seconds after start at t=10s, the seismogram displays two successive amplitude shifts (within the dashed ellipse), which bias the seismogram. This component will be discarded from the inversion.*

Teleseismic (TELE)

Broadband seismograms records of the mainshock at teleseismic distance are obtained from the IRIS and ORFEUS datacenters (**Figure 1a**). The broadband stations are well distributed

azimuthally. Seismograms were deconvolved from the instrument response, integrated to displacement, and equalized to a common magnification and epicentral distance. Signals were bandpassed from 0.01 Hz to 0.8 Hz (P waves) or to 0.4 Hz (SH waves). Finally, waveforms are windowed around P and SH wave trains separately.

GPS

The coseismic horizontal displacements were obtained from Yelles et al. (2004) who processed 9 GPS sites (**Figure 1b**). These sites are located between Algiers and Zemmouri, covering only part of the rupture zone. Static displacement for geodetic data

Coastal uplift (CU)

Coastal uplift data are from Meghraoui et al. (2004). A continuous white band at rocky headlands was measured on the shoreline between Boumerdes and Dellys after the Boumerdes-Zemmouri Earthquake (**Figure 1b**). A maximum of 0.75 m of uplift was observed east of the Boumerdes City and a minimum close to 0 near cap Djenet.

Synthetic Aperture Radar Interferograms (SAR)

The SAR interferograms were processed by Belabbes et al. (2009). Fringes from two descending and one ascending interferograms were digitized and incorporated in the inversion. Since fringes are discontinuous, we defined seven groups of fringes, inside which fringes could be counted sequentially. We attributed a line-of-sight (LOS) displacement values to each fringe, making some assumption about the position of the zero fringe. However, we allowed a LOS shift as a free parameter in the inversion for each of the seven groups. These LOS shifts allow to correct for possible inaccuracy in the assumption of the zero fringe, which may be difficult to identify in certain areas. The interferograms correspond to the following tracks (Belabbes et al. 2009): Envisat descending numbers 04900-06904, between 20030206 and 20030626; Envisat ascending numbers 05308-07312, between 20030306 and 20030724; Radarsat descending number 355996-40798, between 20020927 and 20030829. These three interferograms cover the epicentral area from Cap Matifou in the West to Cap Djenet in the East, but with large gaps in between (**Figure 1c, 1d & 1e**).

Common characteristics of fault models and inversion procedure

The rupture of the Boumerdes earthquake was modeled using the kinematic fault approach of Delouis et al. (2002). It allows for multiple segments of rupture, each defined by its dimension, strike and dip angles, discretized by a regular mesh of point sources to model seismological data, and by a mesh of rectangular subfault dislocations to model geodetic data. Each point source is defined by a local source time function (local STF) representing the seismic moment rate, a direction of slip (rake angle), and a rupture onset time. Each local STFs is represented by three isosceles triangular functions, of width 3s, mutually overlapping. One of the point source coincides with the earthquake hypocenter. The parameters to be inverted for are, for each point source (i.e. subfault), the rupture onset time, the rake angle, and the amplitudes of the three triangular functions. We include seven additional parameters, corresponding to the allowed shifts of the seven fringe groups of the InSAR data. Simplest models will incorporate a single rupture plane (or fault segment), while the more complex ones will include several rupture planes (fault segments) with possibly different dimensions, strikes and dips.

Exploration of the parameter space is carried out with a simulated annealing algorithm (Delouis et al., 2002). It requires the a priori definition of bounding values for the inverted parameters. The convergence of the simulated annealing procedure is based on the minimization of a cost function (Fcost), corresponding to the weighted sum of the normalized RMS (root mean square) misfit function of each dataset, plus two functions, one minimizing the total seismic moment (F_{Mo}), the other smoothing the slip distribution (F_{smooth}).

$$Fcost = \sum_{1}^{n_{datasets}} w_j \, RMS_j + bF_{M_0} + c \, F_{smooth}$$

where « $n_{datasets}$ » is the number of datasets, j corresponds to the j^{est} dataset, «w_j» is the weight of the j^{est} dataset, b and c are two coefficients allowing to increase or decrease the effect of the two associated functions.

The normalized RMS misfit function of each data set is defined as:

$$RMS_j = \sqrt{\sum_1^n [\frac{(O_i - C_i)^2}{O_i{}^2}]}$$

where « n » is the number of data points in dataset « j », , «O_i» and «C_i» the observed and computed values of data point « i », respectively.

The function minimizing the seismic moment is:

$$F_{M_0} = e^{[\left(\frac{M_{0_{mod}}}{M_{0_{ref}}}\right)-1]}$$

with « $M_{0_{mod}}$ » the total seismic moment of the model, « $M_{0_{ref}}$ » a seismic moment of reference.

The smoothing function (F_{smooth}) is simply the normalized sum of the differences in slip values between each subfault and its adjacent subfaults.

Since each dataset is normalized the RMS function, we use the same weight (w_i) for each dataset.

The smoothing coefficient (c) is adjusted by performing synthetic data inversions and retaining the coefficient providing the best fit between the inverted models and the synthetic input model.

Synthetic (computed) seismograms at local to regional distances (strong motion data) are generated with the discrete wavenumber method of Bouchon (1981) designed for 1D velocity models. Velocity models were initially derived from published models (Bounif et al., 2004, Semmane et al. 2005), but we modified them to obtain an optimized model per station, in particular adding a low velocity shallow layer below HDEY and TIZI (**Table 1**).

For the teleseismic record, synthetic seismograms are computed using the ray theory with the method of Nabelek (1984) with simple half space velocity models at the source and at the station.

To compute synthetic static ground displacement, used to model the geodetic data, we follow the formulation of dislocations embedded in elastic half space of Savage (1980).

Table 1: *Velocity models used in the modeling of strong motion records*

Layer thickness (Km)	Vp (km/s)	Vs (km/s)	Density (g/cm^3)	Qp	Qs
Velocity Model for KED1 station					
0.5	4.5	2.6	2.5	150.00	75.00
10.00	5.00	2.89	2.60	300.00	150.00
23.50	6.50	3.71	2.70	400.00	200.00
0.00	8.00	4.62	2.78	1000.00	500.00
Velocity Models for HDEY and TIZI stations					
0.50	3.50	1.38	1.90	150.00	75.00
3.00	4.50	2.57	2.30	300.00	150.00
9.00	5.00	2.86	2.40	400.00	200.00
8.00	5.50	3.14	2.55	500.00	250.00
10.00	7.00	4.00	3.10	700.00	350.00
0.00	8.00	4.57	3.40	1000.00	500.00

Faults models and results

The main parameters (strike, dip, length, width) of the various models tested are given in **Table 2**. The RMS misfit functions relative to the different datasets used in the joint inversion are presented in **Table 3**. In all fault models, we consider the position of hypocenter as fixed. The epicenter had been relocated on the coastline at 36.83°N and 3.65°E by Bounif et al. (2004). Hypocentral depth is the same as in Delouis et al. (2004), i.e. 6.5 km. It is well constrained by the teleseismic data. In all cases, the rake is allowed to vary between +80° and +110°, except for the N100 segment (see below) where rake is in between +75° and +105°. We will not show the slip vectors on the slip maps since they do not vary much. Rupture timing can vary according to a minimum and maximum rupture velocity, respectively Vrmin = 1.6 km/s and Vrmax = 3.5 km/s. For a given subfault, the rupture velocity considered here is the distance to the hypocenter divided by the rupture onset time.

One segment model (1seg)

As a starting model, we use the same model as published in Delouis et al. (2004), hereafter called "Delouis-2004". However, this model is tested here with two additional datasets, strong motion and InSAR. The fault model dimensions are the same, 60 km along strike and 24 km along dip, and the model is positioned at the same coordinates (**Figure 3**). Nonetheless, the rupture model is discretized more finely that Delouis-2004, with subfaults measuring 4km x 4km instead of 6 x 6 km. This is related to the fact that we include the near fault SM and SAR data.

We carried tests varying the strike of the fault around the value used in Delouis-2004 (70°), but with joint inversions of all the datasets (TELE, SM, GPS, SAR, CU). Strikes between 65° and 75° produced equivalent quality results, from the data fitting point of view. Consequently, we retain 70° as the optimum strike, as in Delouis-2004. On the other, hand, fault dip was fine-tuned to 40° (instead of 45° in Delouis -2004). It is a compromise since a value of 45° improves slightly the modelling of the GPS data but degrades to the modelling of the seismological data. The top of the fault model reaches the surface offshore.

Regarding the positioning of the fault model with respect to the coast, models located farther offshore like the one used in Semmane et al. (2005), produce a clear degradation of the modelling of the teleseismic data. The reason is that a model located more offshore imposes a deeper hypocenter, the epicenter being kept fixed, which becomes incompatible with the pP and sP reflected phases of the teleseismic records.

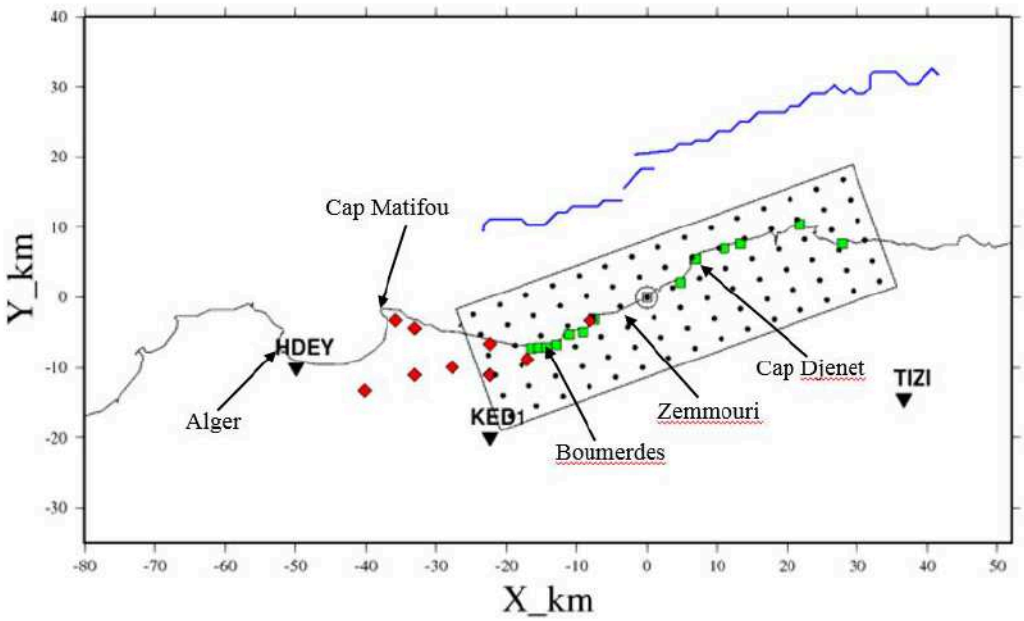


Figure 3: *Surface projection of model 1seg (rectangle), with black dots corresponding to the point sources at the center of the subfaults. Grey square within an open circle: subfault corresponding to the hypocenter; black line: coastline; blue lines: offshore escarpments B1 and B2 described in Deverchère et al. (2005); red diamonds: GPS stations; green squares: points where coastal uplift was measured (Meghraoui et al., 2004); black inverted triangles with names: strong motion stations. Y axis is oriented towards the North.*

The slip distribution is shown on **Figure 4.** It is similar to that of Delouis-2004, with a main slip patch west of the hypocenter, with slip reaching 280 cm, and a shallower patch east of the hypocenter. However, we note that slip extends slightly farther West. Regarding data modelling, we encountered a difficulty to achieve a satisfactory fit of the complete set of data. We will not show the modeling of all the datasets with this initial model but focus on specific problems. The fit of GPS data is slightly degraded with respect to that in Delouis et al. (2004) (**Figure 5**). This is due to the dip value of 40° instead of 45° and to the fact that we include the additional SM and SAR datasets. The overall average rupture velocity, weighted by the slip values, is 2.5 km/s.

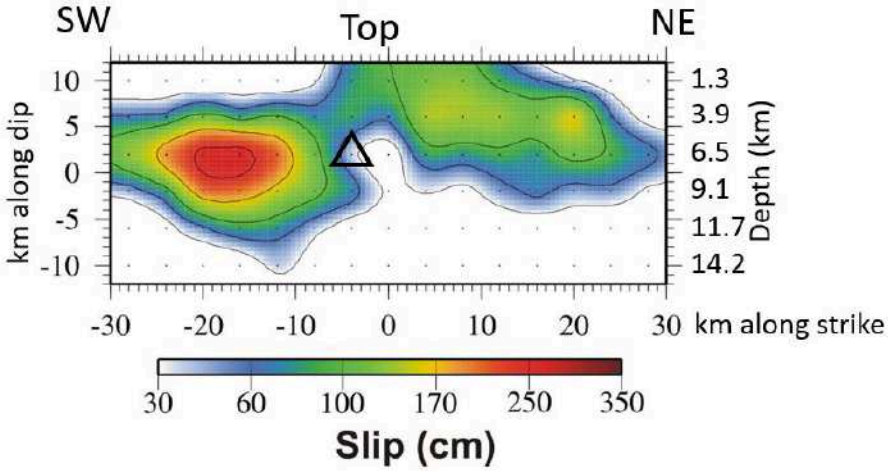


Figure 4: *Slip distribution from the joint inversion of all datasets with 1seg model. The open triangle indicates the point source corresponding to the hypocenter. Black dots correspond to the point sources at the center of the subfaults*

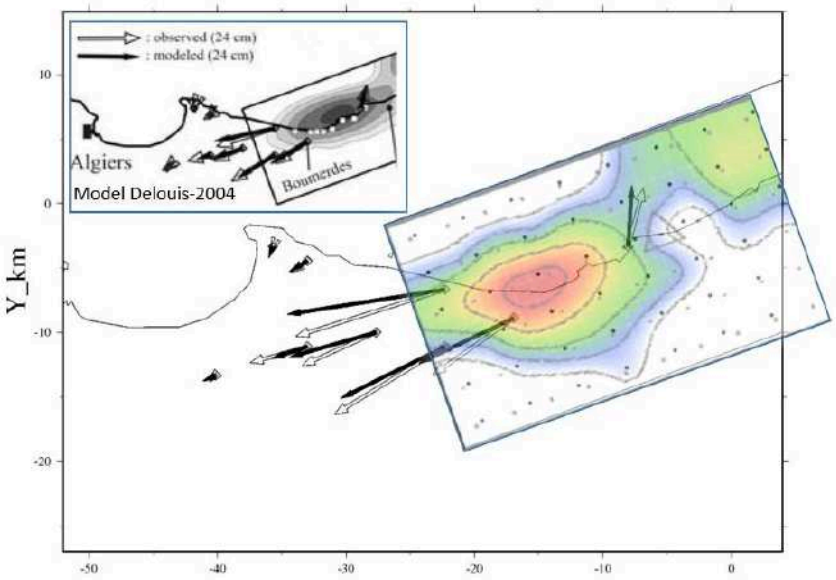


Figure 5: *Zoom on the modelling of the GPS horizontal vectors (main frame) from the joint inversion of all datasets with model 1seg, with only part of the slip model of Figure 4 shown, together with the result from Delouis et al. (2004) in the upper left corner for comparison. Same color scale for slip as in Figure 4.*

Additionally, we observe a difficulty to model correctly SM station HDEY (**Figure 6**). The modelled signal bypasses the first pulses on the horizontal components and the overall amplitudes are not matched at that station. The same difficulty to model HDEY was found in Semmane et al. (2005).

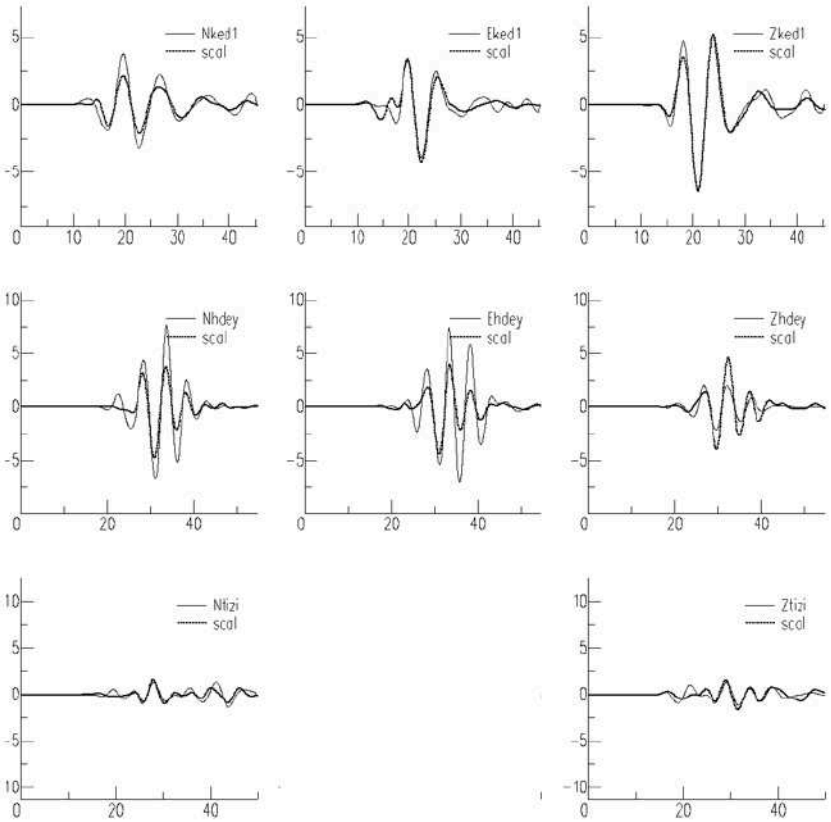


Figure 6: *Modelling of displacement waveforms from the strong motion records obtained with the joint inversion of all datasets with model 1seg. The East component of station TIZI was is not shown since it was discarded due to artifacts in the seismogram (Figure 1c). Amplitudes in cm, time in seconds. Grey line: observed seismogram; heavy black line: computed seismogram.*

The SAR data are well modelled, with a very low normalized RMS misfit function (**Table 3**). It seems that the SAR data do not require more complexity in the fault model, but the problems in the modelling of the GPS and SM (mainly HDEY) data suggest that the simple 1seg model cannot satisfy all the datasets together.

We tested also models with a longer fault, extending farther to the South-West, but it did not produce substantial improvements.

Two segment model with a change in strike (2seg)

In order to invert jointly purely geodetic data (SAR, GPS, and CU), Belabbes et al. (2009) proposed a rupture plane which curves in its western part. Their slip model trends NE-SW (~N60°) between Dellys and Boumerdes, and changes to WNW-ESE (~N100°) between Boumerdes and Cap Matifou. It is remarkable that the trend of the coastline changes in the

same way near Boumerdes, so that their model follows approximately the curve of the coastline.

In this section, we test such a change in strike of the rupture plane, investigating in particular its effect on the modelling of the seismological data which were not included in Belabbes et al. (2009). We decompose the model in two segments, a first one trending N070° (*N70 segment*), identical to our previous model 1seg, and a second one trending N100° (*N100 segment*). We tested different positions for the *N100 segment*, moving it in the inland direction or locating it more offshore, with a dip angle of 40°, 45°, or 50°. These tests were carried out with joint inversions of the whole dataset (TELE, SM, GPS, SAR, CU). The dip and position providing the best overall fit of the data are 45° and the location shown on **Figure 7**, respectively.

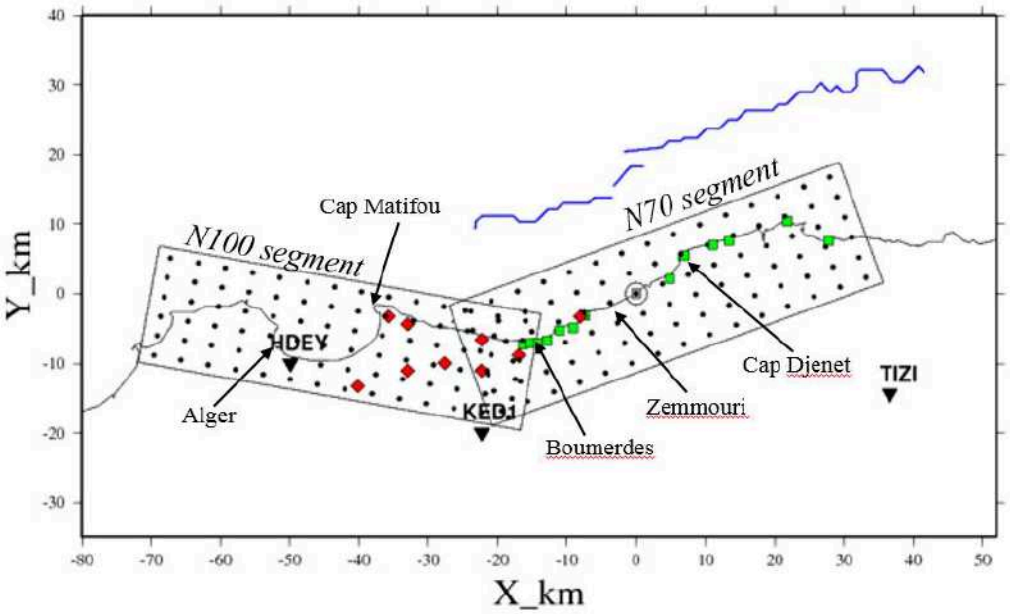


Figure 7: *Surface projection of model 2seg. For other elements, same caption as Figure 3.*

The two fault segments exhibit some overlap in the Boumerdes area, but as shown by **Figure 8** the resulting slip maps do not overlap.

The slip distribution (**Figure 8**) shows two slip patches on the *N70 segment* containing the hypocenter. The eastern patch is shallower, located between 0 and 9 depth, with a maximum slip of 230 cm. The western patch on this same segment is deeper, between 4 km and 12 km depth, and slip reaches 270 cm at its center. *The N100 segment* displays also two slip patches,

a small one in the East of the segment located between 4 km and 8 km depth and a larger one in the western part of the segment, located between 0 and 10 km depth. Maximum slip lies between 110 and 140 cm in both patches of the *N100 segment.*

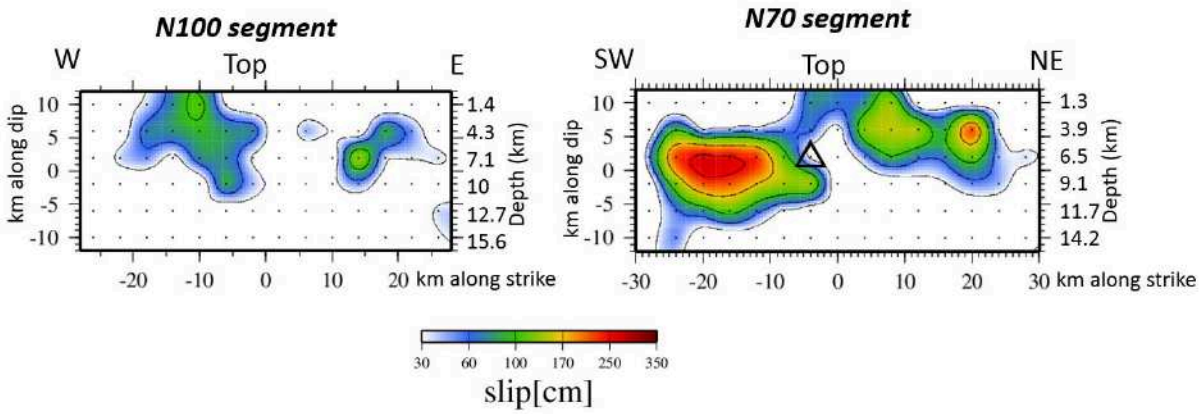


Figure 8. *Slip distribution from the joint inversion of all datasets with 2seg model. For other elements, same caption as Figure 4.*

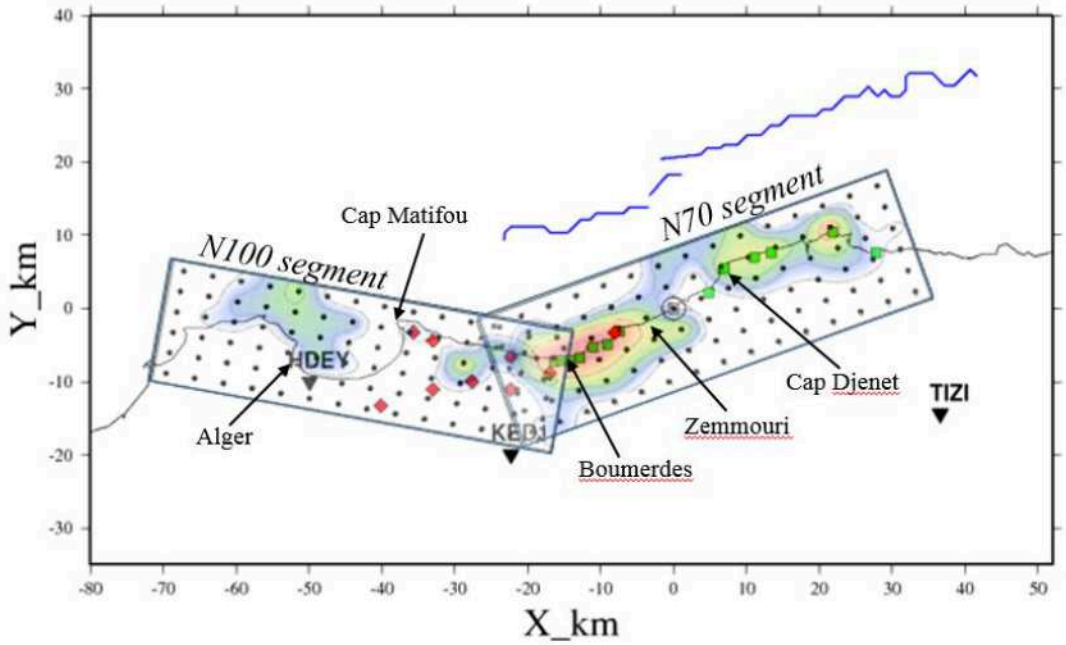


Figure 9: *Surface projection of the slip distribution obtained by joint inversion of all datasets with model 2seg. For other elements, see captions of Figures 3. Same color scale for slip as in Figure 8, but drawn with transparency..*

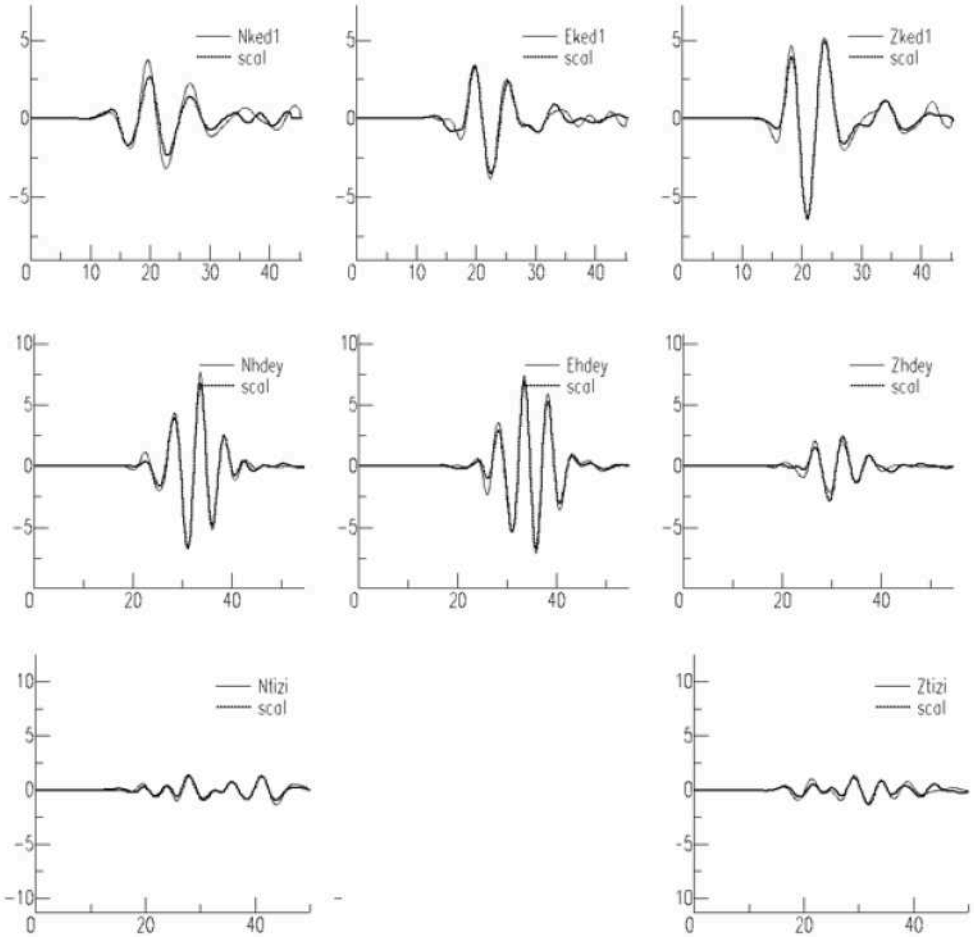


Figure 10: *Modeling of displacement waveforms from the strong motion (SM) records obtained with the joint inversion of all datasets with model 2seg. The East component of station TIZI was is not shown since it was discarded due to artifacts in the seismogram (Figure 1c). Amplitudes in cm, time in seconds.*

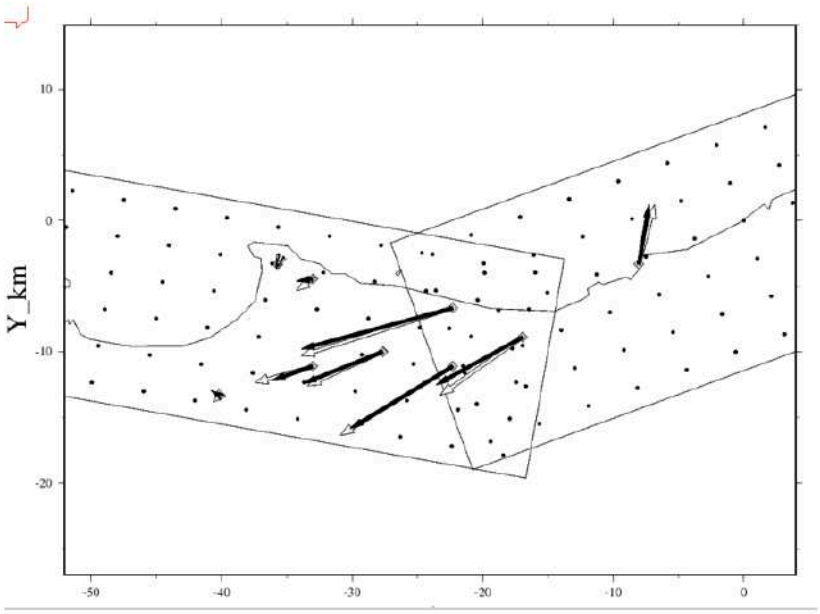


Figure 11: *Modelling of the GPS horizontal vectors from the joint inversion of all datasets with model 2seg.*

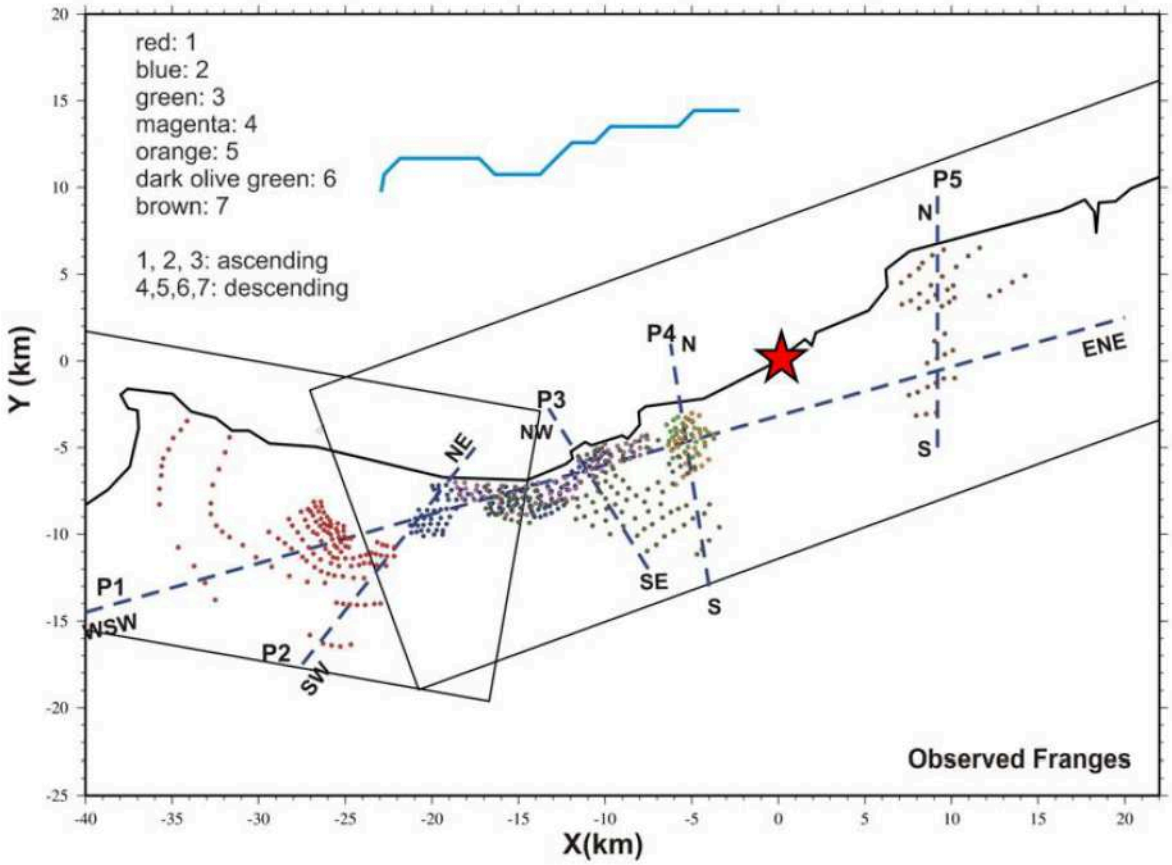


Figure 12: *InSAR data points sampled on the fringes of the three SAR interferograms included in the inversion. InSAR points are distributed among 7 distinct groups ploted with different colors. Black dotted lines are 5 profiles used to show the InSAR data fit in Figure 13. Red star: hypocenter of the*

mainshock; black line: coastline; blue lines: offshore escarpments B1 and B2 described in Deverchère et al. (2005). Rectangular frames: part of the surface projection of the boundaries of model 2seg.

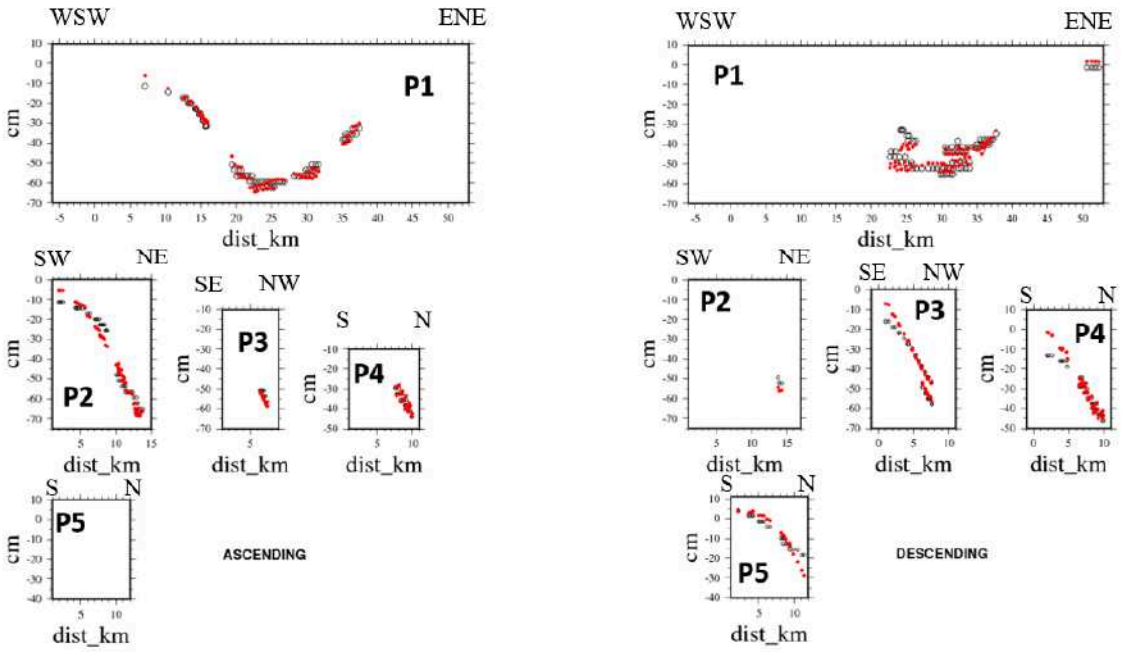


Figure 13*: Modelling of the InSAR data from the joint inversion of all datasets with model 2seg, seen along 5 profiles. Left: from the ascending interferogram; right: from the two descending interferograms. Open circles: observed data; red points: computed data. Vertical axis is the LOS (line of sight) displacement. Negative LOS values mean shortenning of the ground-to-satellite distance and are explained at the first order by coseismic uplift. Horizontal axis is distance along profile, with an arbitrary origin.*

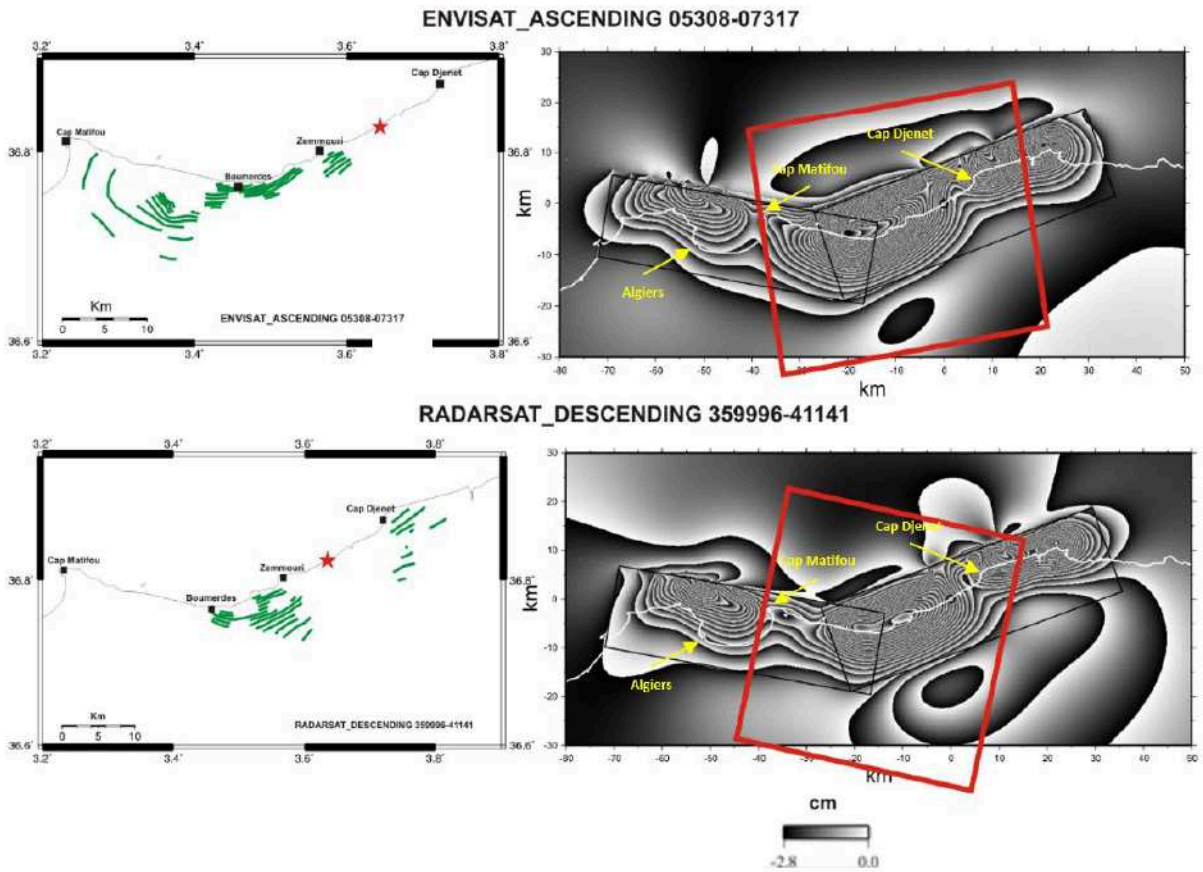


Figure 14: *Top left: digitized Envisat ascending fringes (green lines). Red star: mainshock epicenter. Top right : computed Envisat ascending interferogram from the joint inversion of all datasets, each fringe corresponding to 2.8cm of LOS displacement. The red frame is the positioning of the SAR track. Black rectangles show the surface projection of the 2seg model. White line: coastline. Bottom left and right: same as top, but for the Radarsat descending interferogram.*

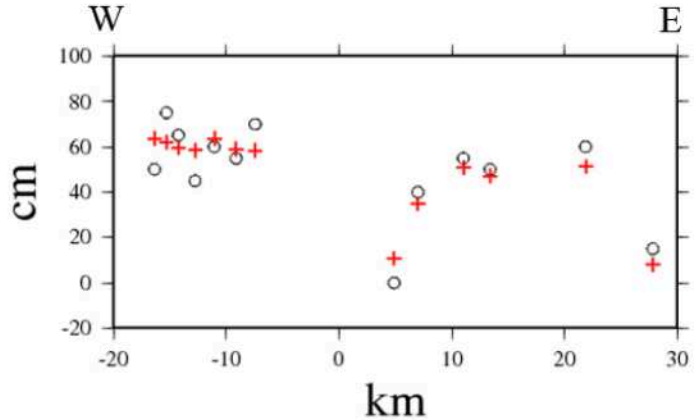


Figure 15 : *Modeling of the coastal uplift (CU) measurements from the joint inversion of all datasets. Measuring points are located along the coast (Figure 7). Horizontal axis: kilometers in the EW direction, with an arbitrary origin. Vertical axis: costal uplift.*

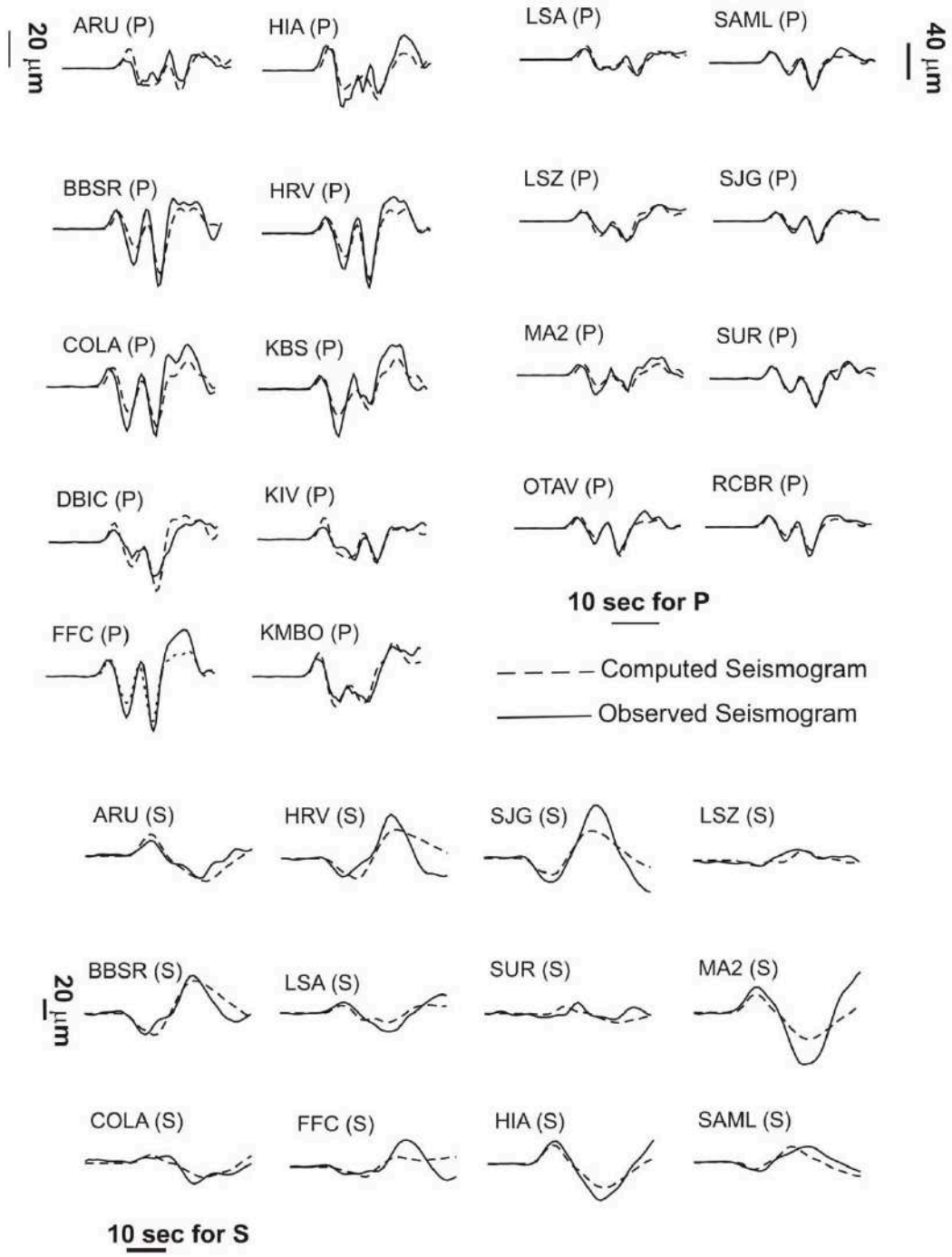


Figure 16: *Waveform fit for the P and SH waves in displacement, from the joint inversion.*

The modelling of the SM data, in particular HDEY station, is strongly improved (**Figure 10**), while the clockwise rotation of the computed GPS largest vectors relative to the observed ones found with the 1seg model (**Figure 5**) is strongly reduced (**Figure 11**).

Figure 17a shows the space and time évolution of the rupture, with cumulative snapshots every 3sec. This figure provides a vision of the propagation of "significative" slip, considered

here as larger than 30 cm. The actual rupture front, related to the propagation of smaller values of slip, and connected to the rupture onset time of the subfaults, may be slightly ahead of it.

As shown by the reference rupture front propagating at 2 km/s (red dashed line of **Figure 17a**), significant slip propagated at a velocity not very far from 2 km/s but with notable variations.

On the N70 main segment, rupture is clearly bilateral, and it propagated faster toward the NE than toward the SW (see slip ahead of the 2 km/s front in the NE at t=9 and 12s). On this segment, slip evolves only marginally after t=15s.

Rupture on the N100 segment starts at about t=12s. On this segment, we observe a shallow slip patch initiating in the West well ahead of the 2 km/s front at t=18s. As indicated on the Figure, rupture velocity at this point is of the order of 3.5 km/s.

The global source time function (STF) shows a main pulse with a duration of 15s, corresponding essentially to the rupture of the main N70 segment, and a continuation until 30s corresponding mainly to the rupture of the secondary N100 segment (**Figure 17b**).

If we compute rupture velocity as the distance from the hypocenter divided by the rupture onset time of the subfaut, and if we average the rupture velocity of several subfaults weighted by their slip values, we find the following:

- the overall average rupture velocity is 2.6 km/s,
- on the N70 segment, average rupture velocity is also 2.6 km/s, but it is 2.8 km/s NE of the hypocenter and 2.5 km/s SW of it,
- on the N100 segment, average rupture velocity is 2.5 km/s.

Average rupture velocity on the N100 segment is not really meaningful given the differenciated behaviour of the two slip zones observed on this segment. The westernmost slip patch is initiating very early since the corresponding rupture front is associated with a rupture velocity of 3.5 km/s (see above).

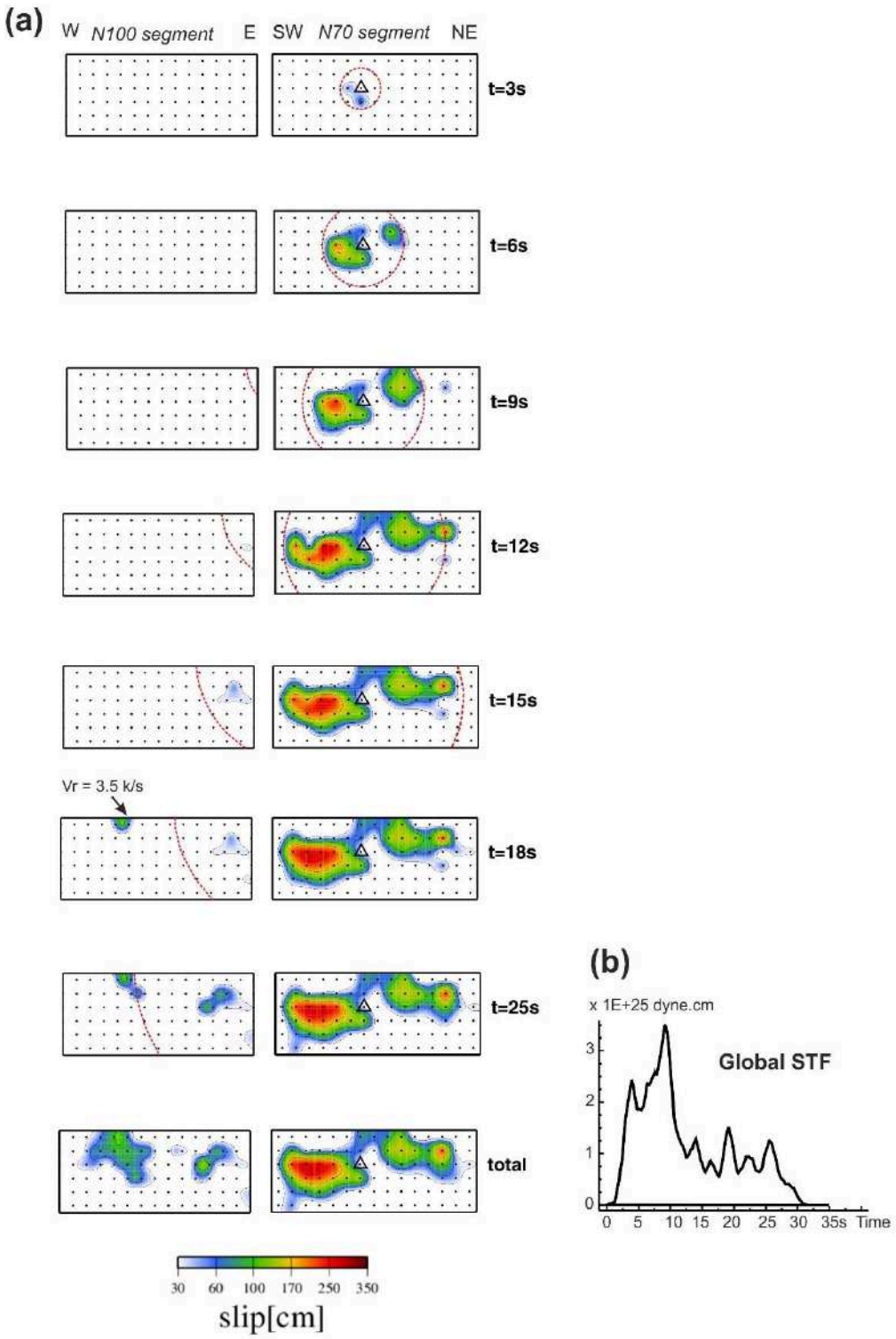


Figure 17: *(a) space and time évolution of the rupture from the joint inversion of all datasets with model 2seg. Slip is displayed on the two segments with cumulative snapshots every 3 seconds. "Total" means the final slip distribution once rupture ended. The red dashed line shows a rupture front corresponding to a constant rupture velocity of 2 km/s, for reference. The initiation of the westernmost slip patch on the N100 segment is associated with a rupture velocity of 3.5 km/s as indicated at t=18s. (b) corresponding global source time function (STF).*

If we carry out the same inversion but with a zero weight for station HDEY (i.e. station not taken into account), the slip area north of HDEY, i.e. north of Algiers in the western part of the *N100 segment*, disapears in the resulting model (**Figure 18**). However, as shown by **Figure 19** , the fit of HDEY records is then poor, the computed waveforms lacking amplitude in their last part. The same mismatch was observed in the modelling of Semmane et al. (2005) for the HDEY station (**Figure 20**).

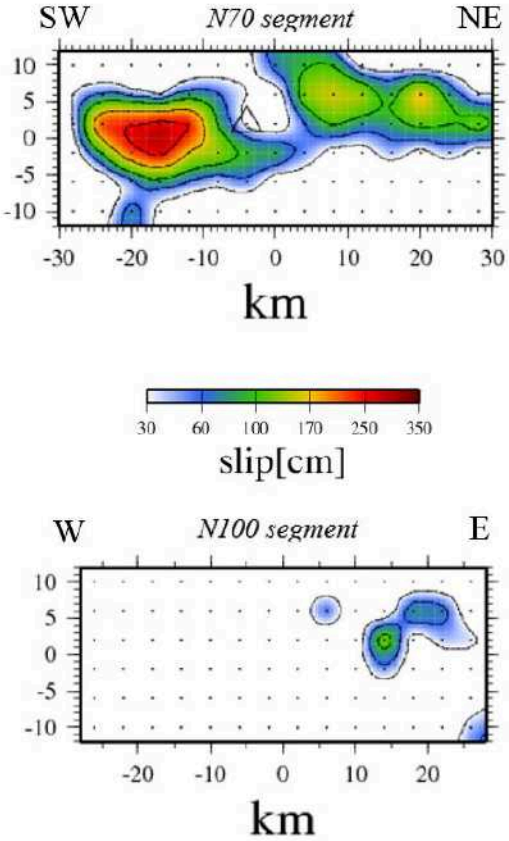


Figure 18: *Same as Figure 8 but without station HDEY to which a zero weight is attributed.*

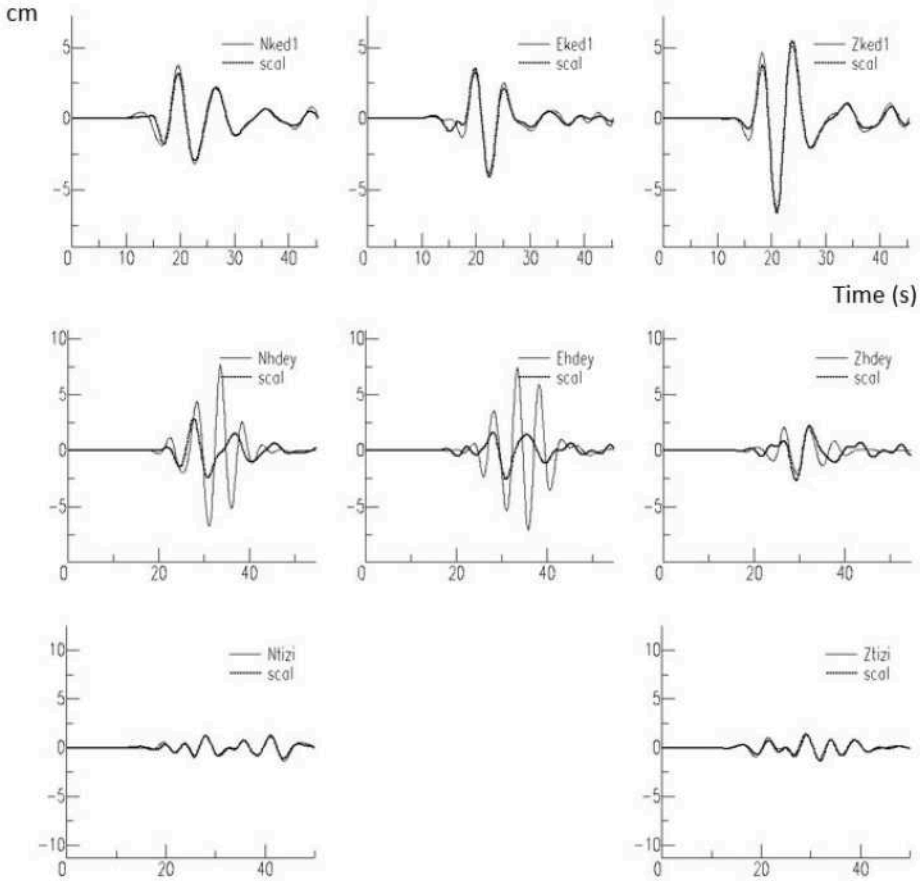


Figure 19 : *same as Figure 10 but without station HDEY to which a zero weight is attributed.*

Figure 20 : *extract of Figure 3 of Semmane et al. (2005) showing the modelling of the HDEY station that they obtained. Continuous line is the observed seismogram and dashed line the computed one. Note that modelling shows a high degree of misfit, as shown in our Figures 6 and 19, but not as in our Figure 10.*

On the other hand, the GPS data remain well modeled (see RMS value in **Table 3**). We conclude that the westernmost slip patch on the *N100 segment*, located offshore Algiers, is

required by station HDEY, and only by that station. Note that the SM station HDEY is the only measuring point available in the West near Algiers.

We questioned whether the existence of this westernmost slip patch could be an artefact related to an inappropriate estimation of the P wave arrival on HDEY. We ran different inversions with different timing for HDEY, as well as for the other stations, and this slip patch was systematically found.

We note that the RMS of the teleseismic data increases when SM station HDEY is not used in the inversion (**Table 3**). This means that station HDEY, when used, helps finding a solution event more compatible with the teleseismic data. This comforts us in the idea that station HDEY contains a meaningful information and has to be retained.

Four segment model with flat and ramp offshore (4seg)

Through the combined analysis of high-resolution bathymetry and seismic reflection surveys, Deverchère et al. (2005) identified escarpments outcropping offshore the Boumerdes-Zemmouri area displaying evidence of recent activity. According to these authors, the escarpments are likely related to a propagating fault system at shallow depth, displaying a ramp – flat geometry, associated with folding and growth of the Plio-Quaternary sediments.

We constructed a model following the geometry of faults offshore as proposed by Deverchère et al. (2005), with a shallow ramp reaching the surface where these authors locate their B1 and B2 escarpments (**Figure 21**). The *ramp segment* , striking N70° and dipping 50°S, connects at about 3 km depth to a *flat segment* striking also N70°, dipping 8°S, which in turn connects to our main *N70 segment* at about 5 km depth (**Figure 22**). The parameters of the complete model are presented in **Table 2**.

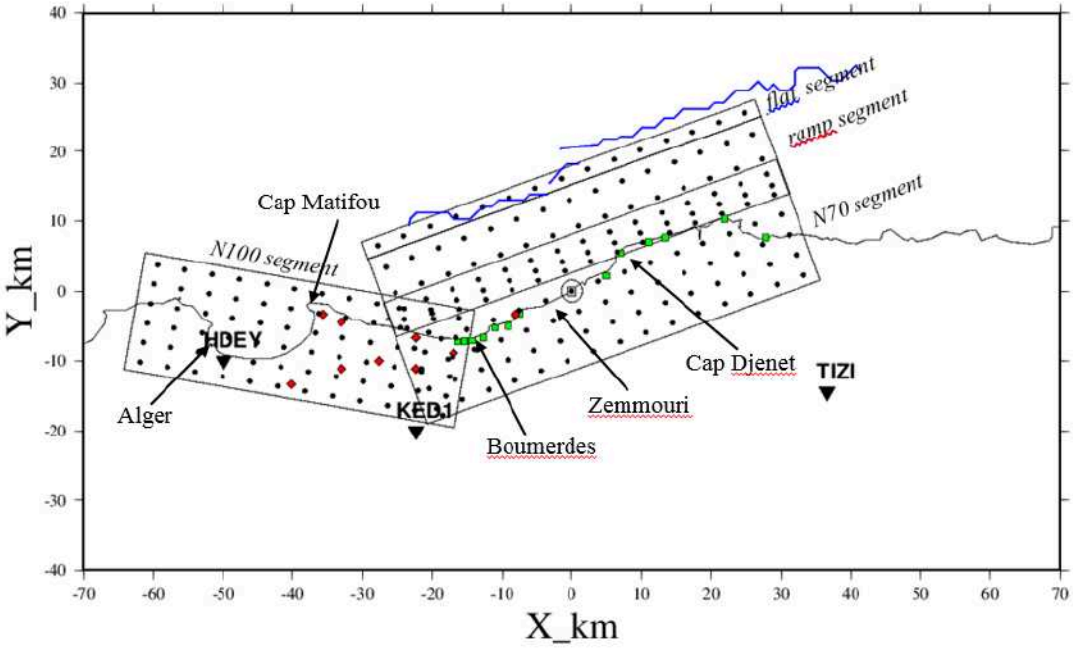


Figure 21: *Surface projection of model 4seg. For other elements, same caption as Figure 3.*

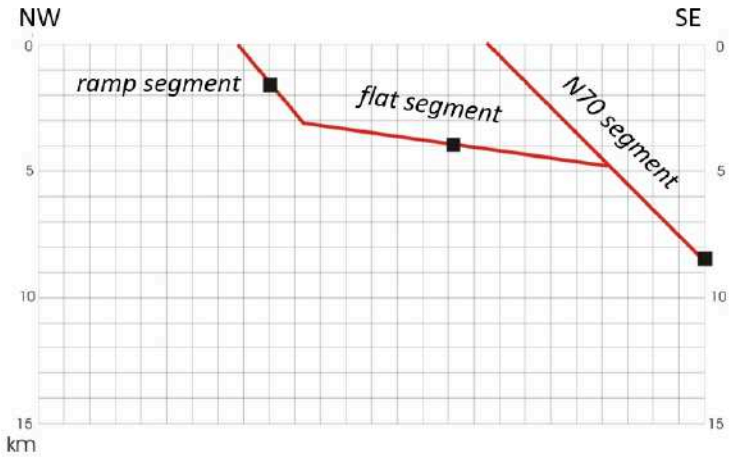


Figure 22*: vertical cross section perpendicular to azimuth N70, showing the position of the three fault segments oriented N70, called the N70 (main) segment, ramp segment, and flat segment. Black squares are located at the middle of each segment.*

The resulting slip model is presented in **Figure 23**.

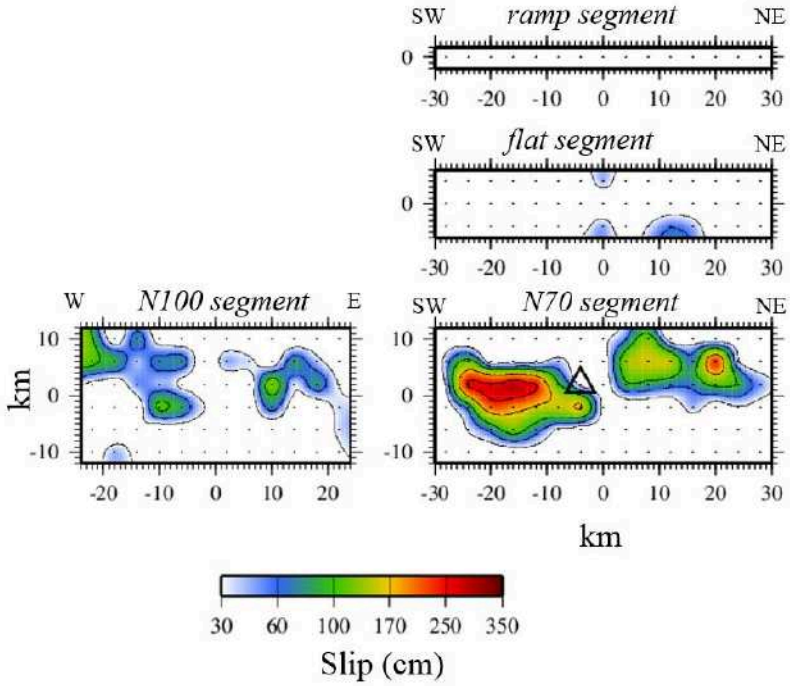


Figure 23: *Slip distribution from the joint inversion of all datasets with 4seg model. For other elements, same caption as Figure 4.*

As shown by **Figure 23**, the slip distribution on the N70 segment of the 4seg model is similar to what was obtained in the previous 1seg and 2seg models (**Figures 4 and 8**). On the *N100 segment*, we observe again two slip areas, as in the 2seg model (**Figure 23**), although the precise shape of the westernmost asperity is different. This indicates that this westernmost slip patch, which is a persistent feature when SM station HDEY is considered, cannot be resolved in too much detail. This will be confirmed by the resolution tests.

Slip on the two offshore segments is very limited, below 20 cm on the *ramp segment,* and reaching 50 to 60 cm on localized small slip patches on the *flat segment* (**Figure 23**).

Data modelling is only marginally improved with respect to the 2seg model, as shown by the RMS misfit values in **Table 3**. Considering that the 4seg model is more complex than the 2seg model, with more parameters, without noticeable improvement of data modelling and very few slip on the supplementary *ramp* and *flat segments,* it is difficult to ascertain the validity of the 4seg model. Model 2seg, simpler and fitting the data equally well, may be considered as a preferred model. This is why we do not show the modelling of the different datasets with the 4seg model.

Table 2: *Geometry and dimension of the fault model*

	Strike (deg.)	Dip (deg.)	Length (km)	Width (km)	Length and Width of subfaults (km)
Model one segment					
Segment 1	70	40	60	24	4 x 4
Model one segment extended					
Segment 1	70	40	80	24	4 x 4
Model two segments					
Segment 1 (*N70 segment*)	70	40	60	24	4 x 4
Segment 2 (*N100 segment*)	100	45	56	24	4 x 4
Model 4 segments					
Segment 1 (*N70 segment*)	70	40	60	24	4 x 4
Segment 2 (*N100 segment*)	100	45	48	24	4 x 4
Segment 3 (*flat segment*)	70	8	60	12	4 x 4
Segment 4 (*ramp segment*)	70	50	60	04	4 x 4

Table 3: *Root Mean Square misfit function (RMS) and seismic moment (Mo) in for the different inversions presented in this study. TELE: teleseismic, SM: strong motion; SAR: InSAR; CU: coastal uplift.*

	1seg model	2seg model	2seg model without HDEY	4seg model
RMS TELE	0.486	0.475	0.509	0.456
RMS SM	0.515	0.216	0.199	0.220
RMS SAR	0.087	0.079	0.081	0.080
RMS GPS	0.181	0.115	0.106	0.108
RMS CU	0.181	0.167	0.167	0.167
M0 (dyn.cm)	2.96×10^{26}	3.67×10^{26}	3.15×10^{26}	3.73×10^{26}

Resolution tests with synthetic data

In order to assess how the different datasets resolve the slip distribution and to which extent the joint inversion allows to better resolve it, we carried out a series of inversions of synthetic data. A synthetic slip model is constructed, with slip patches located on the different segments of the model. This synthetic model is used to generate synthetic data which are finally inverted with the same parameterization as for the real data inversions. Synthetic inversions where carried out with different coefficients of smoothing, searching for the coefficient which minimized the root mean square difference between the synthetic and inverted slip distributions, both for the slip amplitude and rupture timing. All the results presented in this study, obtained with synthetic or real data, where carried out with this optimal coefficient of smoothing.

We applied this approach to the 2seg and 4seg models. The first one because it is our preferred model explaining most completely all the real datasets, the second because we need to assess to which extent slip on the distant offshore segments (flat and ramp segments) can be resolved.

2seg model

For the 2seg model we performed six inversions of synthetic data, testing all five datasets (TELE, SM, SAR, GPS, CU) individually and jointly (**Figure 24**). The synthetic slip model is characterized by five asperities with different shape distributed on the two segments numbered a1 to a5 (**Figure 24**).

The inversion of the teleseismic synthetic data exhibits a "blurred" and spread out version of the three slip patches on the N70 segment, and a weak and mislocated slip zone on the N100 segment.

The inversion of strong motion synthetic data resolves better the three slip patches on the N70 segment, although with a lack of amplitude on a1 and a3. On the N100 segment, the two slip patches a4 and a5 are found, but weaker and deeper than in the synthetic model.

The inversion of synthetic coastal uplift data retrieves three asperities on the N70 segment, with a tendency to overestimate the maximum slip on a1 and a3. Furthermore, a1 is replaced by a twin slip zone, with a deep part near the bottom of the model which is mislocated. Nothing is retrieved on the N100 segment. It is related to the absence of coastal uplift measuring points in the West above this segment (**Figure 7**), and to the fact that the total seismic moment is minimized, avoiding effectively spurious slip where slip is not required by the data.

The inversion of GPS data does not find slip on the eastern part of the N70 segment (slip zones a2 and a3), also because GPS measuring points are absent in this area (**Figure 7**). However, it resolves approximately slip zone a1 and the two slip zones on the N100 segment (a4 and a5).

The inversion of the synthetic SAR data retrieves logically slip zones a1 and a5 which are in the area where most of the SAR data points are located, and not the a3, a2, a4 slip zones which are located where too few SAR points exist. The shape of a1 and a5 is nonetheless found only approximately, with a tendency to be spread out.

Finally, the joint inversion of all the synthetic datasets together is providing the overall best result, retrieving all the five slip zones, although with a less resolved character for the extreme ones, a3 and a4, with respect to a1, a2, and a5.

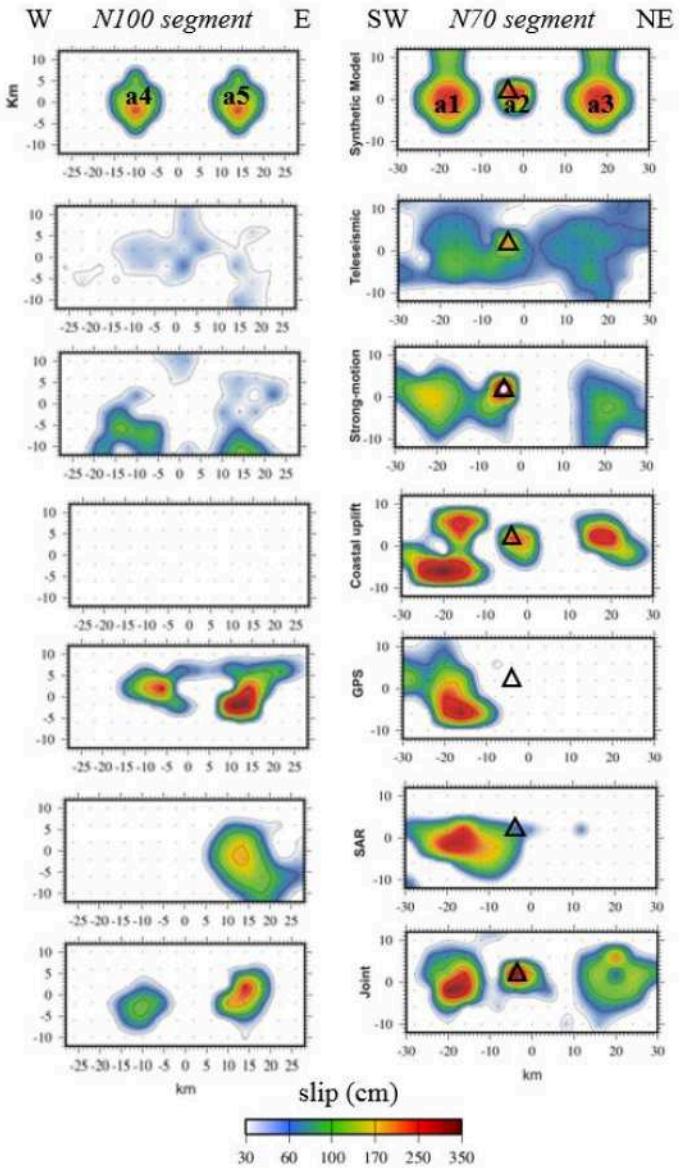


Figure 24: *Resolution tests with synthetic data for the 2seg model. Left column: N70 segment; right column: N100 segment. The first row displays the synthetic model. Rows 2 to 6 show the results of the inversions of the individual datasets, teleseismic, strong motion, coastal uplift, GPS, and InSAR*

respectively. The last row displays the result of the joint inversion of all the datasets together. For other elements, same caption as Figure 4.

4seg model

We show the result of the joint inversion only because individual inversions have too many similarities with the ones shown in the case of the 2seg model. The synthetic slip model is characterized now by nine slip patches, a1 to a5 being the same as for the 2seg model, and a6 to a9 being located on the offshore flat and ramp segments (**Figures 21** (geographical map) **and 23** (slip map)). Slip amplitude on a6 to a9 is taken relatively smaller because we do not expect high slip value on these offshore segments.

As expected, for slip patches a1 to a5 the joint inversion provides a result similar to that obtained with the 2seg model (**Figure 25**).

Slip zones a6 and a7 on the flat segment are hardly retrieved, weaker that the synthetic ones. Nothing is found on the farthest ramp segment (a8 and a9). In fact, low values of slip (< 10cm) can be found on the ramp segment but it cannot be considered as significant.

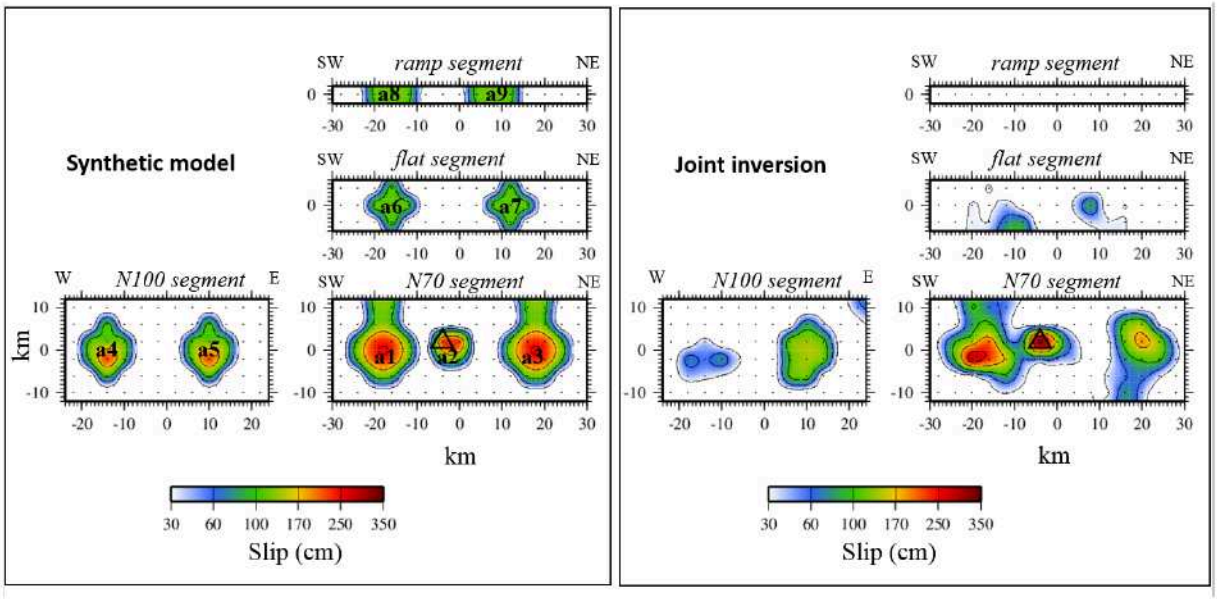


Figure 25. *Resolution tests with synthetic data for the 4seg model. Left frame: synthetic model; right frame: result of the joint inversion of all datasets. For other elements, same caption as Figure 4.*

Discussion - Conclusion

We started by using a simple model with uniform geometry (strike N70, dip 40), identical to the one used in Delouis et al. (2004) except for the dip which was formerly 45°, but incorporating additional data, namely the InSAR (SAR) and strong motion (SM) data. This simple model provided a slip distribution similar to the one of Delouis et al. (2004) but we identified some incompatibility among the more complete datasets, with a difficulty to model the records of the SM station HDEY and the GPS data. This led us to explore more complexity in the geometry of the rupture model.

A change in strike of the rupture plane in the West had been already proposed by Belabbes et al. (2009) and the activation of shallow offshore ramp and flat faults had been suggested by Deverchère et al. (2005). Based on these elements, we constructed two alternative models, named 2seg and 4seg, incorporating a second segment oriented N100° (2seg model) in the West and two additional segments offshore (4seg model).

With the 2seg model, the apparent incompatibility among the datasets could be solved, with a slip distribution modelling correctly all the datasets, in particular the GPS and SM station HDEY which were problematic with the 1seg model. However, a new slip area appears offshore Algiers (**Figure 9**), not identified in previous studies. We showed that this slip zone is required by the SM station HDEY only. We also emphasize that the other data (TELE, GPS, INSAR, CU), although globally little impacted by this slip zone, are compatible with it (RMS values in **Table 3**). For example, we observe that computed InSAR fringes (**Figure 14**) are effectively nearly closing at, or just east of, Cap Matifou, as do the observed ones. Regarding coastal uplift, nothing was documented in the Algiers area. In order to assess whether this westernmost slip zone should have produced coastal uplift which should have been observed, we calculated the related vertical uplift along the coast near Algiers (**Figure 26**). We find predicted uplift to be equal or lower than 20 cm in the Algiers area, a value which may be effectively below the observational threshold.

From the synthetic tests, we learn that we must be careful in not over interpreting the precise shape of this westernmost slip zone, which is not very well resolved.

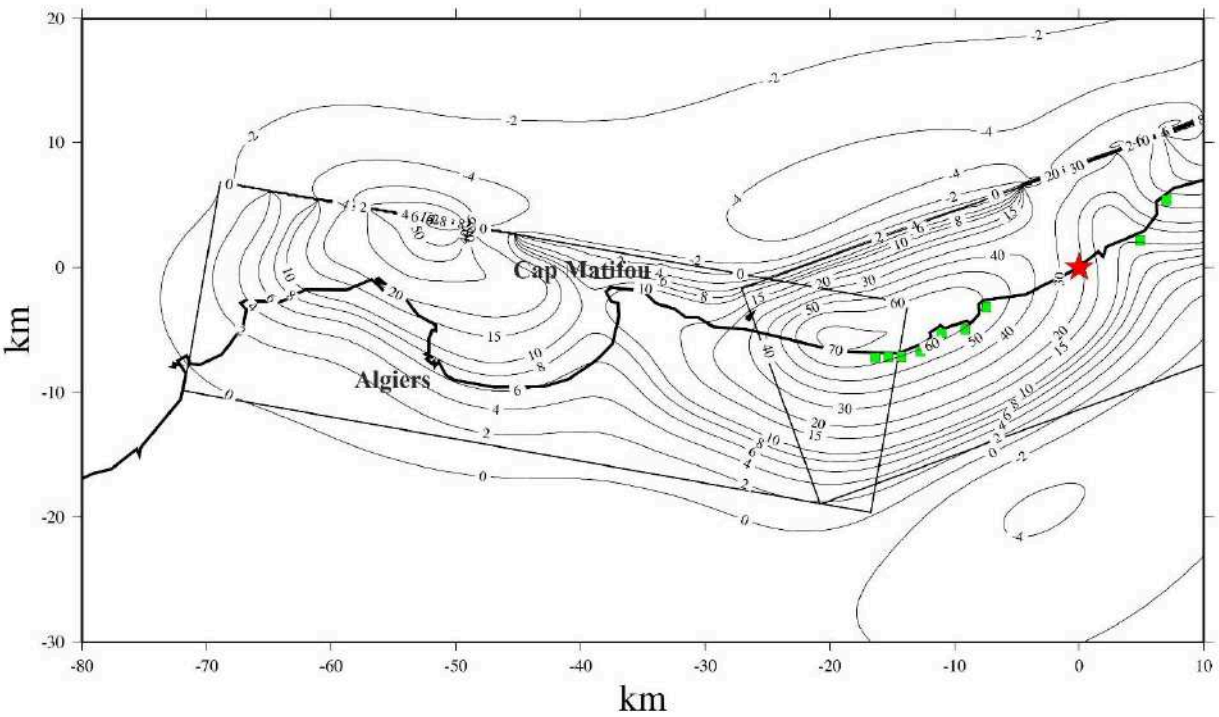


Figure 26. *Vertical displacement at the surface produced by the slip distribution found with the 2seg model (Figure 9), from the joint inversion of all the datasets. Heavy black line: coastline; thin black curves: vertical displacement with values indicated in cm; red star: mainshock epicenter; rectangular frames: surface projection of the contours of the fault segments of the 2seg model; green squares: points where the coastal uplift was measured. Vertical axis positive towards the North, horizontal axis positive towards the East.*

Studies on the aftershocks of the 2003 Boumerdes earthquake display distributions of aftershocks that are mainly located in the vicinity of our N70 fault segment, with very few events west of Cap Matifou. At first glance, aftershocks do not support the existence of the westernmost slip area that we find offshore Algiers. However, the temporary seismic network which was installed to record the aftershocks was deployed essentially east of Cap Matifou, and it did not cover sufficiently the area of Algiers, especially offshore, to record well small (M < 4) aftershocks there. From the international catalogs (e.g. USGS), we can nonetheless state that no aftershocks with M > 4 occurred in this area.

Nonetheless, a few aftershocks located by Ouyed et al. (2011), and to a lesser extent by Kherroubi et al. (2017), may be correlated with the westernmost slip zone, as shown by **Figure 27**.

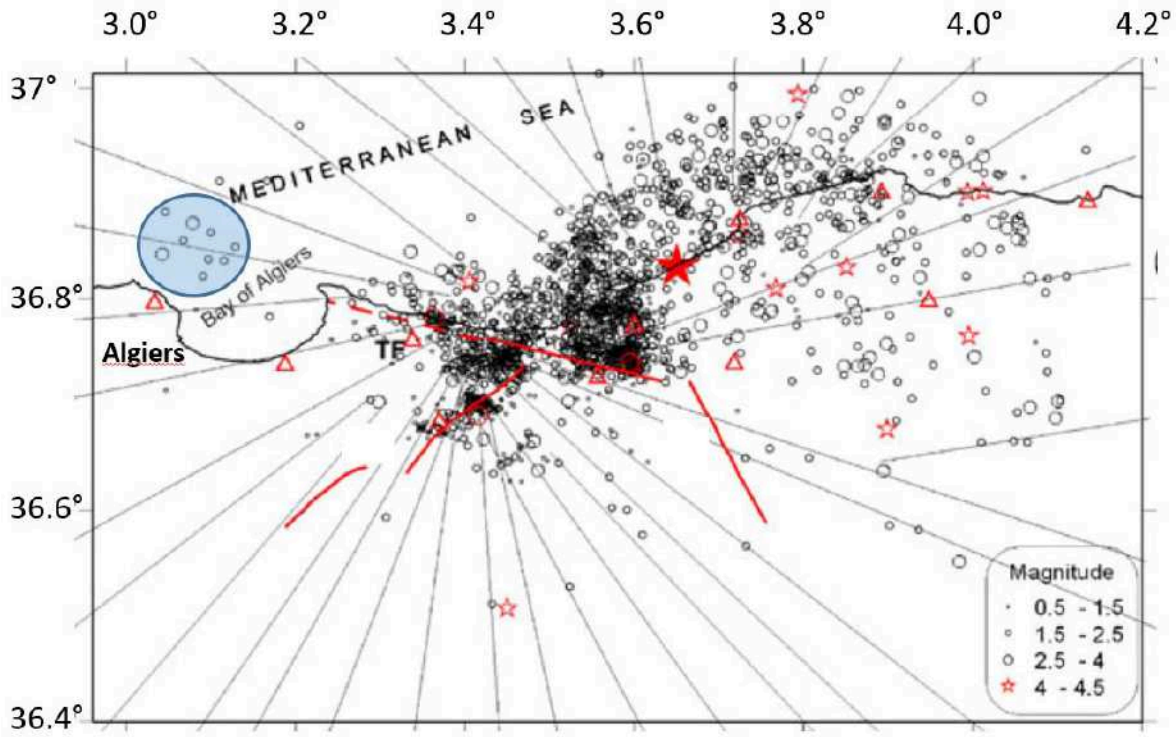


Figure 27. *Aftershocks of the 2003 Boumerdes earthquake from May 23 to June 30, 2003, modified from the Figure 4 of Ouyed et al. (2011). Blue transparent zone: area of aftershocks in the immediate vicinity of the westernmost slip zone of our preferred 2seg model. Red filled star: mainshock epicenter. Red lines; mapped faults (TF: Thenia Fault). Open red triangles: seismic stations.*

Intensity maps of the 2003 mainshock do not show any localized increase of the intensity in the western part of Algiers that may be related to our westernmost slip zone. One possibility to explain this, but unproved at this stage, would be that the coseismic rupture of the fault offshore Algiers, corresponding to the western part of the N100 segment of our 2seg model, generated few destructive high frequencies, i.e. was characterized by the emission of low frequency seismic waves.

Regarding the coseismic activation of the offshore faults derived from the observations of Deverchère et al. (2005), our inversion with model 4seg suggests that some slip with a multi-decimeter amplitude may have occurred there, especially on the flat segment. However, from the synthetic tests we confirm that resolution on the offshore segments is low, and we conclude that inland data, even completed by the teleseismic records, are insufficient to resolve moderate amplitude slip there. Hence, the question of the coseismic activation of shallow offshore secondary faults remains open.

Acknowledgments

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4.2. Effet de la distribution de glissement sur le changement de contrainte de Coulomb sur quelques failles cibles où se sont produits des séismes récents de M > 4-5

Dans cette partie de la thèse, nous allons nous intéresser à l'interaction des failles sismiques en terme de la fonction de rupture de Coulomb (CFF : Coulomb Failure Function). Selon King et al., (1994) et Lin et al., (2004, 2011) lorsqu'un séisme se produit, l'état de contrainte au voisinage de cette faille est perturbé, cela va influencer le champ de contraintes des failles voisines, ce qui va avancer ou retarder le jeu de la faille. En cas d'avancement, il peut en résulter l'occurrence du futur séisme.

Sur la base du modèle de rupture 2seg (segments N70 et N100) du séisme de Boumerdes-Zemmouri trouvé dans cette thèse, nous allons présenter les résultats de la CFF calculée sur des failles cibles qui ont joué après le séisme du 21 mai 2003 (Mw 6.8) considéré comme la faille source. Les failles cibles considérées dans ce travail (**Table 4-1** et **Figure 4-1)** sont des séismes modérés localisés dans le voisinage de la faille source et qui sont :

- Le séisme de *Hammam Melouane* du 17/07/2013, Mw 5.0 situé au S du segment N100
- Le séisme de *Bordj Ménaïel* du 22/02/2014, Mw 4.1, localisé au SE du segment N70
- Le séisme de la *Baie d'Alger* du 01/08/2015, Mw 5.7, localisé au Nord des deux segments N100 et N70.

Séismes	*Date*	*Mw*	*Strike1*	*Dip1*	*Rake1*	*Strike2*	*Dip2*	*Rake2*	*Depth (km)*	*Références*
Hammam Melouane	*17/07/2013*	*5.0*	*5°*	*85°*	*33.8*	*272°*	*56.3°*	*174°*	*7*	*Yelles-Chaouche et al., 2017*
Bordj Ménaïel	*22/02/2014*	*4.1*	*84°*	*83°*	*170°*	*175°*	*80°*	*7°*	*7*	*Semmane et al., 2015*
Baie d'Alger	*01/08/2015*	*5.7*	*45°*	*45°*	*68.7°*	*254°*	*49°*	*110°*	*6*	*Beldjoudi, 2014*

Table 4-1 : *paramètres de la source sismique de séismes qui se sont produits après le séisme de Boumerdes-Zemmouri, localisés dans la région épicentrale du séisme du 21 mai 2003.*

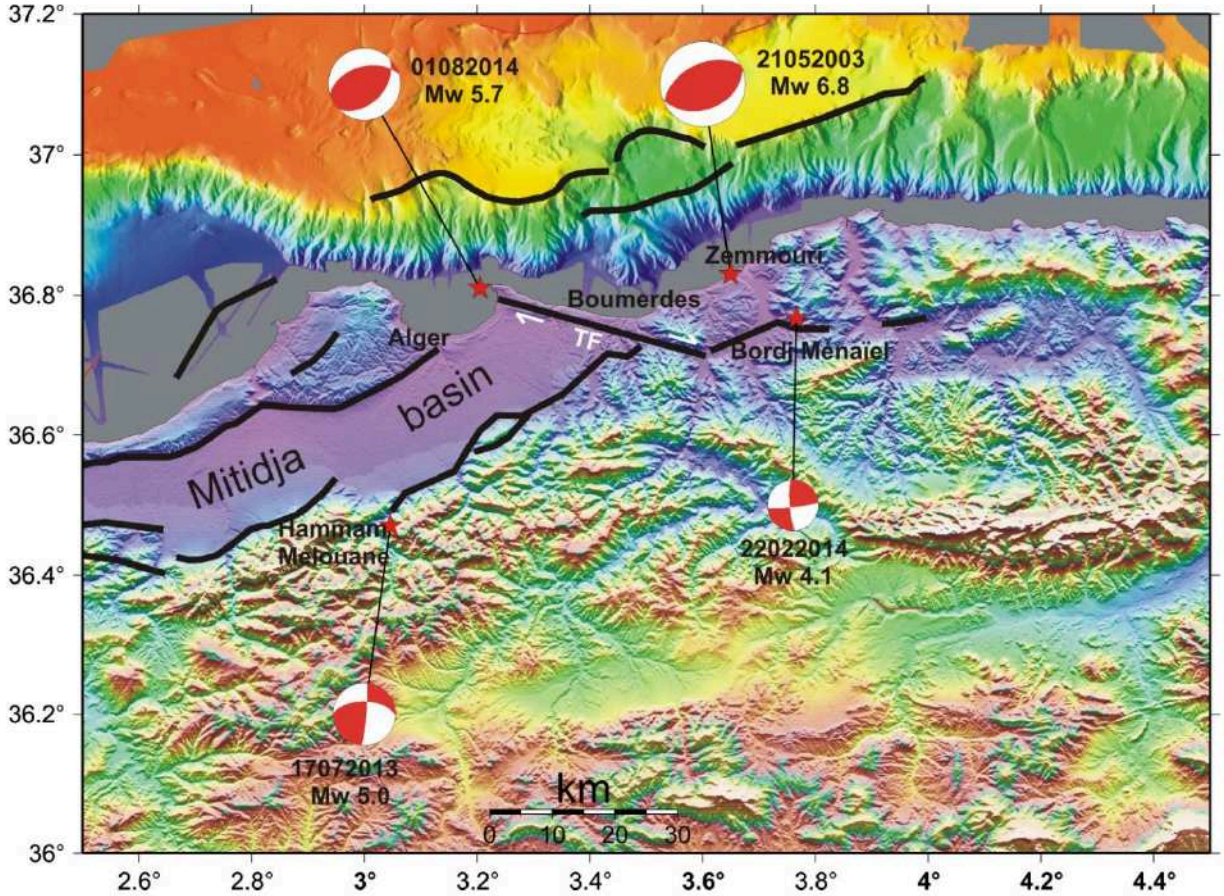


Figure 4-1 : *Mécanismes aux foyers du choc principal (source) et des séismes cibles dans la région de la Mitidja. Les traits en noir montrent les failles actives de la région. TF est la faille de Thénia, les flèches en blancs montrent le sens du mouvement de TF.*

4.2.1. Principe du calcul de la CFF

Le changement de la fonction de rupture de Coulomb (ΔCFF) est calculé par la relation

$$\Delta CFF = \Delta\tau + \mu\,(\Delta\sigma + \Delta p) \qquad (1)$$

Où $\Delta\tau$ est le changement de contrainte de cisaillement sur le plan de faille cible (positif dans le sens du glissement de la faille) généré par le glissement sur la faille source. $\Delta\sigma$ est le changement associé de contrainte normale sur le plan de faille cible, μ est le coefficient de friction et Δp représente la variation de la pression de pores (Stein et al., 1999).

Δp est généralement très mal connue, mais elle reste proportionnelle au changement de la contrainte normale (Lin et al., 2011), et la pluspart des études utilisent l'expression simplifiée de la CFF et définissent un coefficient de friction apparent μ', où Δp est implicitement considérée et l'équation devient :

$$\Delta CFF = \Delta\tau + \mu' \Delta\sigma \qquad (2)$$

Le tenseur de contrainte est calculé par la loi de Hooke, et la déformation cosismique dans le milieu environnant la faille source est calculée à partir de la distribution de glissement sur la faille source en utilisant le formalisme d'Okada (1992) pour un demi-espace élastique et homogène, dans lequel la constante de cisaillement est égale à 3.2 x 10^5 bars et le coefficient de Poisson est pris égal à 0.25. Le changement de contrainte de cisaillement cosismique causée par la faille ''source'' ($\Delta\tau$) est calculée sur un plan spécifique de la faille cible définie par les paramètres strike, dip et rake. Par contre le changement de contrainte normale ($\Delta\sigma$) est indépendante du glissement de la faille cible. Le coefficient de friction apparent μ' est compris entre 0.0 et 0.8. Dans ce travail, nous avons pris une valeur moyenne de μ' de 0.4 (King et al., 1994, Lin et al., 2011).

4.2.2. Le modèle de rupture

Le modèle de rupture que nous allons utiliser dans cette étude est celui présenté dans ce chapitre dans le paragraphe 4.1. dédié à l'étude du glissement du séisme de Boumerdes-Zemmouri obtenu à partir d'une inversion conjointe des données télésismiques, accélérométriques, GPS, SAR et CU. Le modèle est appelé ici Beldjoudi2017 par commodité, composé de deux segments rectangulaires, le premier orienté N70 avec un plongement de 40° vers le S, de longueur 60 km et de largeur 24 km et le deuxième segment orienté N100 avec un plongement vers le SO de 45°, de longueur 56 km et de largeur de 24 km. Le modèle est divisé en 174 sous-failles rectangulaires de (4 km × 4 km) représentée chacune par un glissement individuel. Le moment sismique total de ce model est de 3.27 × 10^{+26} dyne cm, soit une Mw = 6.98 (**Figure 4-2**).

4.2.3 Calcul de la ΔCFF

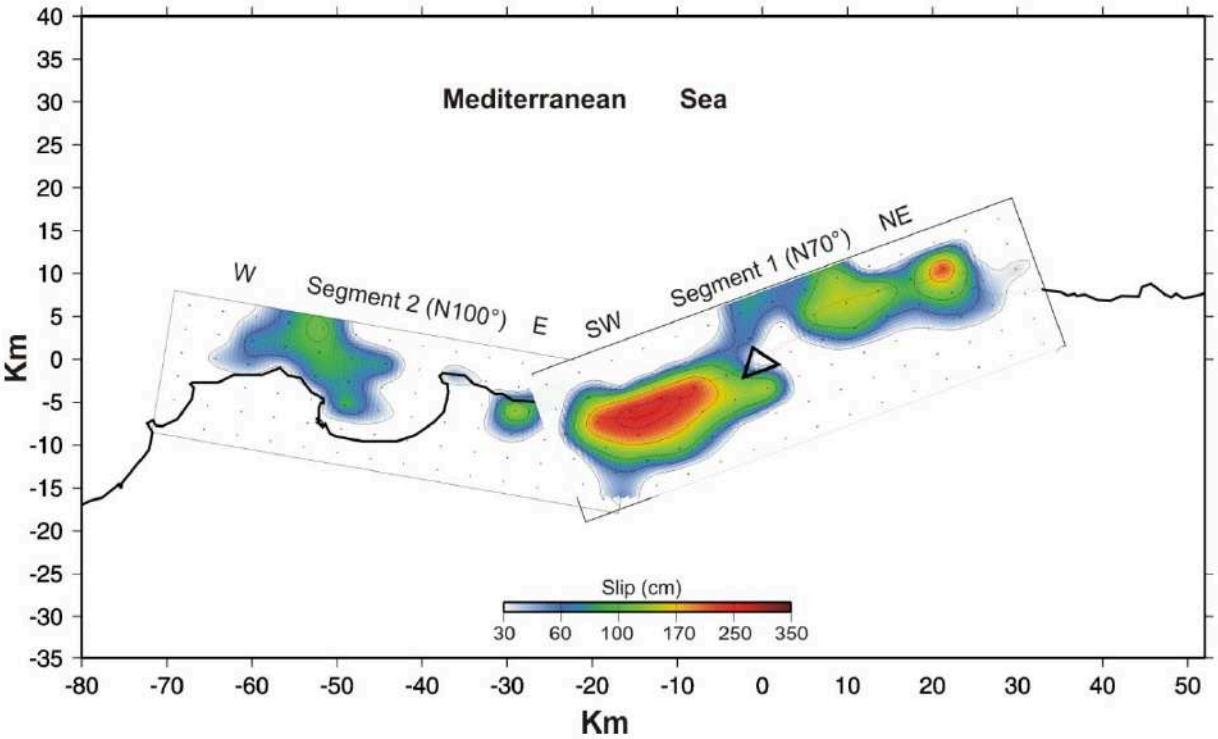


Figure 4-2 : *Modèle de glissement cosismique, Beldjoudi2017 utilisé comme modèle de source pour le calcul de la ΔCFF sur des failles cibles des séismes cités ci-dessus.*

Le séisme de Hammam-Melouane (17/07/2013), Mw 5.0

Ce séisme s'est produit sur la bordure sud de la Mitidja, vers 04h00 du matin (heure locale) au mois du Ramadhan, personnellement, je l'ai bien ressenti, il était à quelques 30 km de chez-moi. C'était, la panique dans mon quartier, heureusement il n'a pas fait de victimes.

Nous avons calculé, la ΔCFF sur les deux plans nodaux donnés par le mécanisme au foyer que j'ai calculé (Yelles-Chaouche et al., 2017) c'est un décrochement senestre pour le plan NS et dextre pour le plan orienté EO (**Figure 4-1** et **Table 4-1**). L'analyse de la ΔCFF sur ce séisme (cible), montre :

Le Plan1 (strike,dip,rake) = (5°/85°/33.8°)

Pour ce plan1, le séisme (cible) se trouve dans une région où la ΔCFF est négative (**Figure 4-3a** et **4-3b**) et à cette endroit la valeur de la contrainte de cisaillement $\Delta\tau$ = -0.042 bar, la contrainte normale $\Delta\sigma$ = -0.153 bar et donc la valeur de la ΔCFF = -0.103 bar, ce qui montre que ce plan n'était pas favorable à la rupture. Les structures situées à l'ouest du segment 2 ;

l'anticlinale du sahel (bordure nord du bassin de la Mitidja), et la bordure sud de la Mitidja se trouvent dans une région de ΔCFF > 0 (chargée). La même remarque peut être faite pour les structures offshore situées au large du segment 2 et au large du segment 1 dans sa partie Est. la localisation d'un séisme est entachée toujours d'erreurs sur sa localisation et selon (Yelles-Chaouche et al., 2017), l'erreur est de ±3 km sur l'horizontale et la verticale, si nous déplaçons notre épicentre à 3 km dans tout les sens, il reste toujours situé dans une zone ou la ΔCFF est négative.

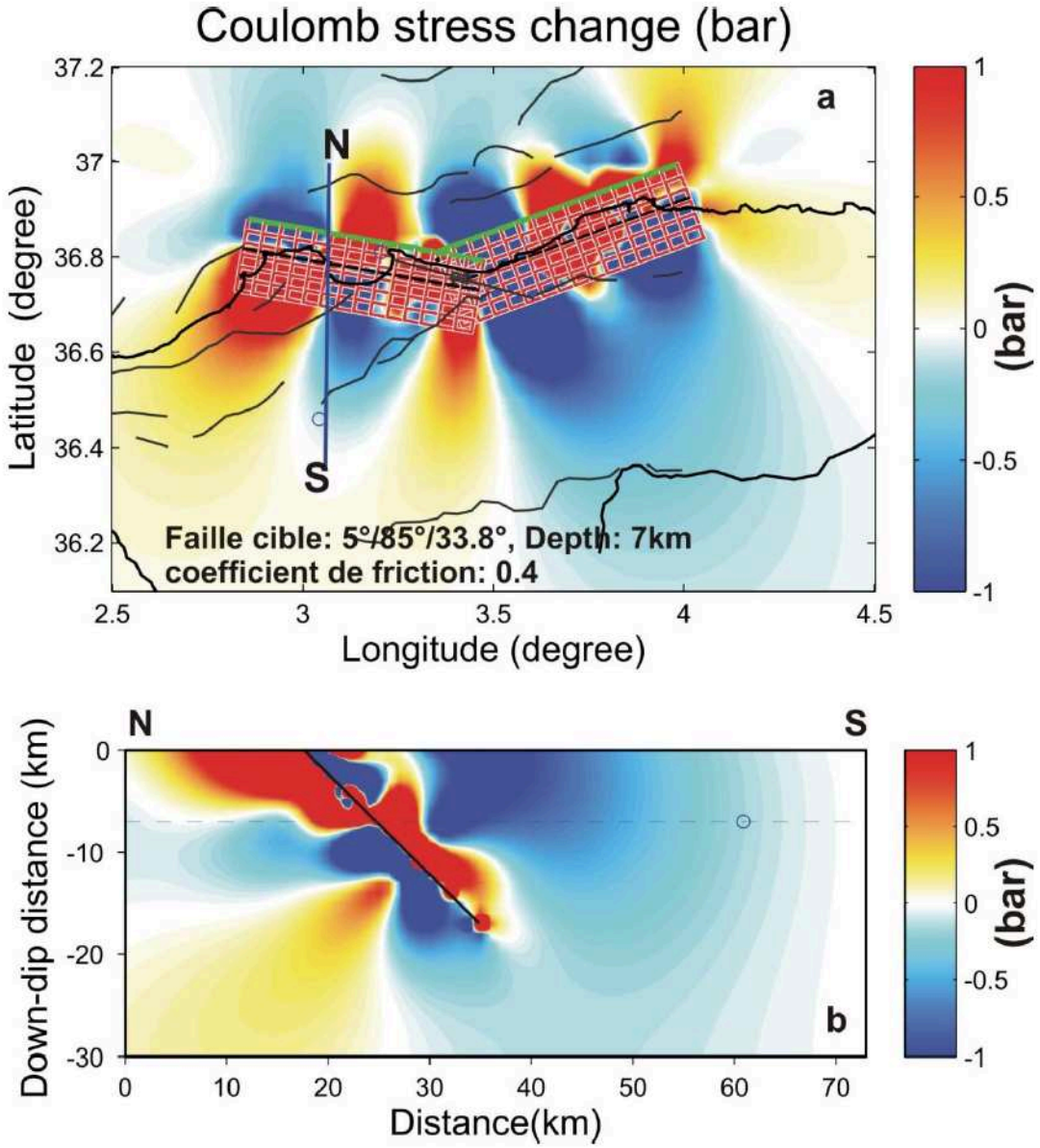


Figure 4-3 : *Carte de la chute de contrainte pour la cible (5°/85°/33.8°). a) la ΔCFF calculée à une profondeur de 7 km. La couleur rouge montre une augmentation de la ΔCFF (zone de chargement) et les failles ayant les mêmes paramètres que la cible sont favorables à la rupture, par contre si celles-ci*

se trouvent dans les zones de couleur bleu, elles ne sont pas favorables à la rupture (déchargement de la contrainte de Coulomb), le trait en bleu avec la mention NS, montre la direction de la coupe, les traits en gris sont les structures actives de la région. Les segments 1 et 2 sont présentés par les rectangles rouges. b) la Coupe NS montre que la cible est localisée dans une région de déchargement. Le trait noir est la projection en profondeur de la faille source.

Le Plan2 (strike,dip,rake) = (272°/56.3°/174°)

L'épicentre se situe dans une région où la ΔCFF est positive et a une valeur de 0.134 bar (**Figure 4-4a** et **4-4b**). Ce plan2 représente donc le plan le plus favorable à la rupture si nous considérons que son activation a été déclenchée par le séisme de Boumerdes. Même remarque, concernant l'augmentation de la contrainte de Coulomb sur les structures du bassin de la Mitidja et les structures au large d'Alger. En profondeur, la région Nord de la faille cible est chargée entre 0-10 km. Ce chargement est repris sur la partie sud de la cible entre 2-20 km. Si nous considérons l'erreur de la localisation et si nous déplaçons l'épicentre de 3 km dans les trois sens, il reste toujours dans une région de ΔCFF >0.

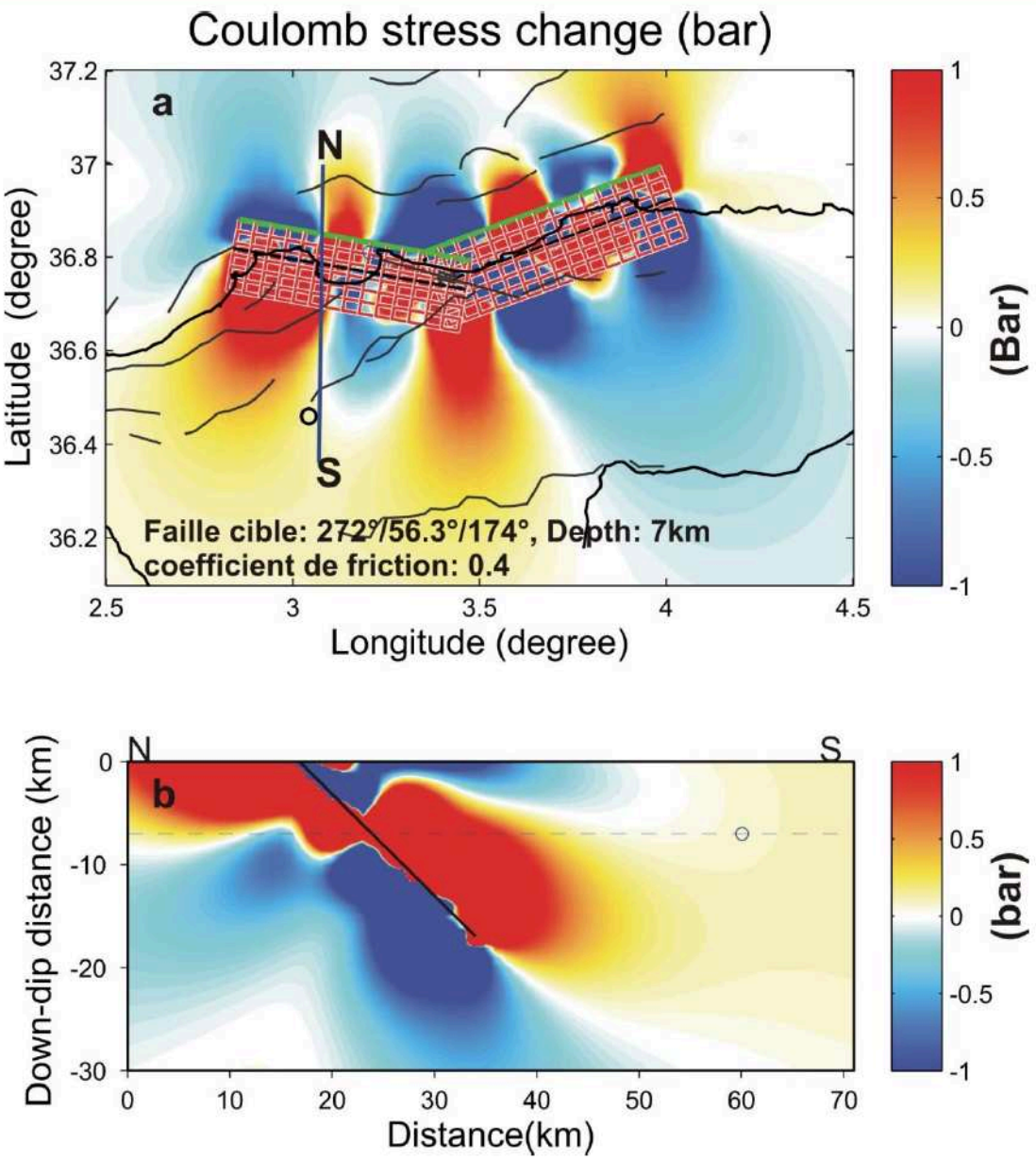


Figure 4-4 : *Même légende que dans la (Figure 3-4)*

Le séisme de Bordj Ménaïel (22/02/2014) Mw 4.1

Ce séisme est le plus faible en termes de magnitude (Mw 4.1) comparé aux séismes de Hammam Melouane (Mw 5.0) et de la Baie d'Alger (Mw 5.7). Ce séisme malgré qu'il soit localisé tout prés de deux failles (**Figure 4-1**), une orientée NE-SO, cartée dans Yelles-Chaouche et al. (2009) comme une faille inverse avec un plongement vers le Nord et une autre structure orienté ~NE inverse avec un plongement vers le Nord. Semmane et al., (2017) n'ont pas pu l'affecter à l'une de ces deux failles et cela par absence de carte

sismoctectonique détaillée dans cette région. Le mécanisme calculé, est celui d'un décrochement, et de fait la question reste posée quand à la faille ayant jouée.

Plan1 (strike,dip,rake) = (84°/83°/170°)

En se référant aux **Figures** (**4-5a**) et (**4-5b**), l'épicentre se trouve dans une région où la ΔCFF est positive. La valeur–du changement de contrainte de cisaillement $\Delta\tau$ = -0.849 bar, la contrainte normale $\Delta\sigma$ = 3.337 bars, ce qui donne une ΔCFF = 0.485 bar. Les structures de la partie Est de la bordure sud et nord du bassin de la Mitidja ainsi que les structures offshore sont situées dans des régions chargées en termes de perturbation de contraintes (ΔCFF > 0). En profondeur, au Nord de la faille cible, nous trouvons une ΔCFF > 0 entre 0-30 km, par contre au Sud de la cible, la zone de chargement touche les structures situées entre 0-10 km.

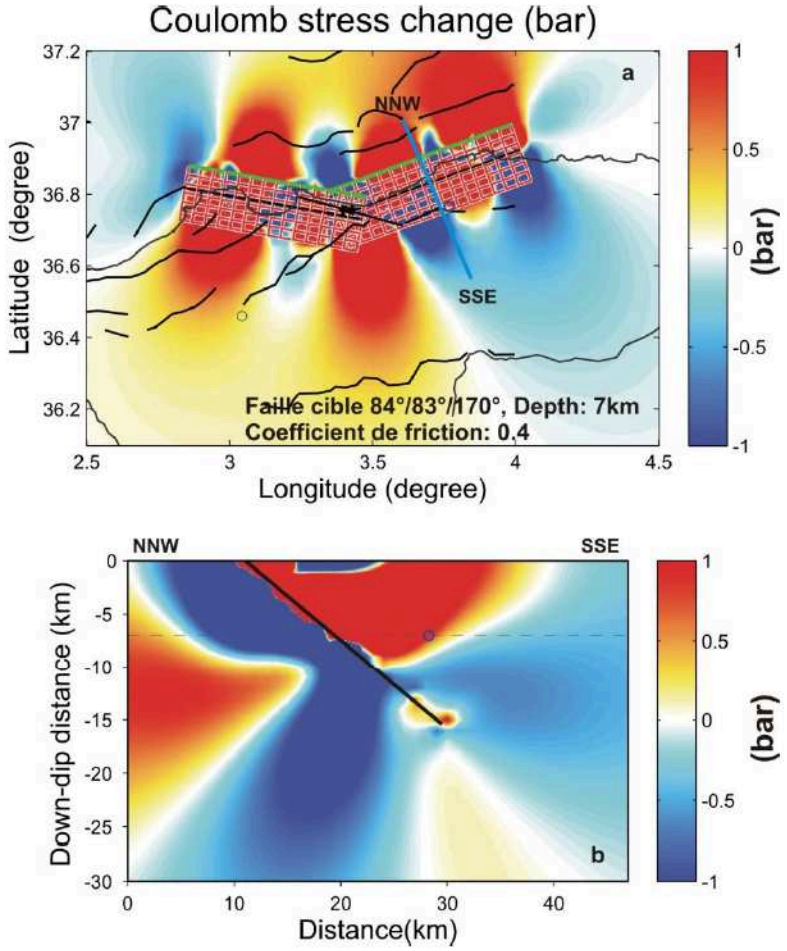


Figure 4-5 : *Même légende que dans la (Figure 4-3), sauf que la coupe est orientée NNO-SSE*

Plan2 (strike,dip,rake) = (175°/80°/17°)

Le calcul de la ΔCFF pour ce deuxième plan, indique que l'épicentre est localisé dans une région où la ΔCFF est positive (**Figure 4-6a** et **4-6b**), le changement de contrainte de cisaillement $\Delta\tau$ = -0.855 bar, la normale $\Delta\sigma$ = 4.028 bar et la ΔCFF = 0.756 bar. Ce que nous pouvons conclure à propos de ce séisme est que les deux plans plan1 et plan2 sont favorables à la rupture.

Même en terme d'erreur de localisation horizontale et verticale (±3 km), l'épicentre demeure toujours dans une région de ΔCFF > 0.

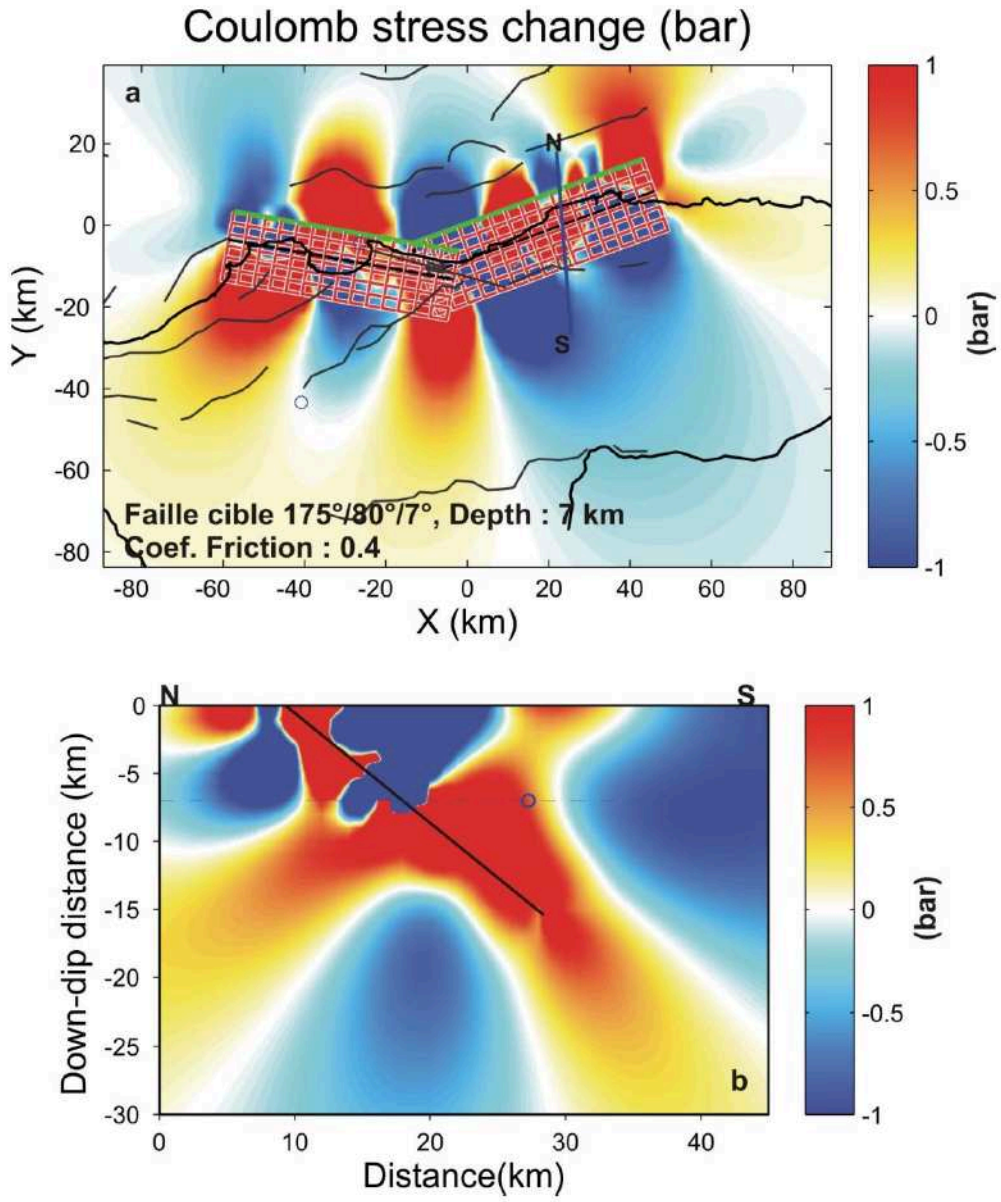


Figure 4-6 : *même légende que dans la Figure (4-3)*

Le séisme de la Baie d'Alger (01/08/2014) Mw 5.7

Ce séisme vient s'ajouter à la série de la sismicité qui a concernée la région d'Alger, Boumerdes, depuis le séisme du 21 mai 2003. Il a la plus forte magnitude parmi les trois. Il s'est produit à 05h00 du matin (heure locale). Je me suis réveillé quelques instants (~10 mn) avant que ce séisme se produise, je l'ai bien ressenti, moi, et ma famille. Dès que la secousse s'est arrêtée, nous sommes sortis vers l'extérieur et j'ai commencé à rassurer ma famille et mes voisins en leurs donnant quelques conseils, puis j'ai rejoint le CRAAG pour suivre et prendre en charge avec mes collègues cette crise sismique. Le mécanisme au foyer (**Figure 4-1**, **Table 4-1**) de ce séisme a été déterminé automatiquement par inversion des formes d'ondes (FMNEAR), puis je l'ai révisé.

Plan1 (strike,dip,rake) = (45°/45°/68°)

Ce plan est localisé dans une zone à ΔCFF > 0, avec un $\Delta\tau$ = 2.573 bar, un $\Delta\sigma$ = -0.175 bar et une ΔCFF=2.503 bar. Il correspond à une faille inverse qui plonge vers le SE, avec une composante de décrochement senestre (**Figure 4-1**). La région Ouest de la Mitidja est relativement moins chargée (**Figure 4-7a**) comparée à celles trouvées pour les séismes de Hammam Melouane et Bordj Ménaïel. La partie Est de la faille source est aussi chargée, les failles qui se trouvent dans la partie offshore ne sont pas chargées. En profondeur (**Figure 4-7b**) et au dessous de la faille source, la ΔCFF est positive et s'étend jusqu'à 30 km. Le dessus de la faille cible est aussi chargé.

Plan2 (strike,dip,rake) = (254°/49°/110°)

Ce plan est représenté par un plan de faille inverse qui plonge vers le NO, avec une composante de décrochement dextre. La ΔCFF calculée sur ce plan est aussi positive, ΔCFF=2.306 bar avec une $\Delta\tau$ = 2.548 bar et une $\Delta\sigma$ = -0.605 bar. La répartition de la ΔCFF (**Figure 4-8a** et **4-8b**) est presque la même que celle trouvée dans le cas précédent. Les deux plans de failles plan1 et plan2 ont les mêmes chances d'avoir été déclenchés. Si nous faisons bouger l'épicentre vers le Nord ou bien vers le Sud en considérant une erreur sur la

localisation de ± 3 km en verticale et en horizontale, l'épicentre va se retrouver dans une zone de ΔCFF < 0.

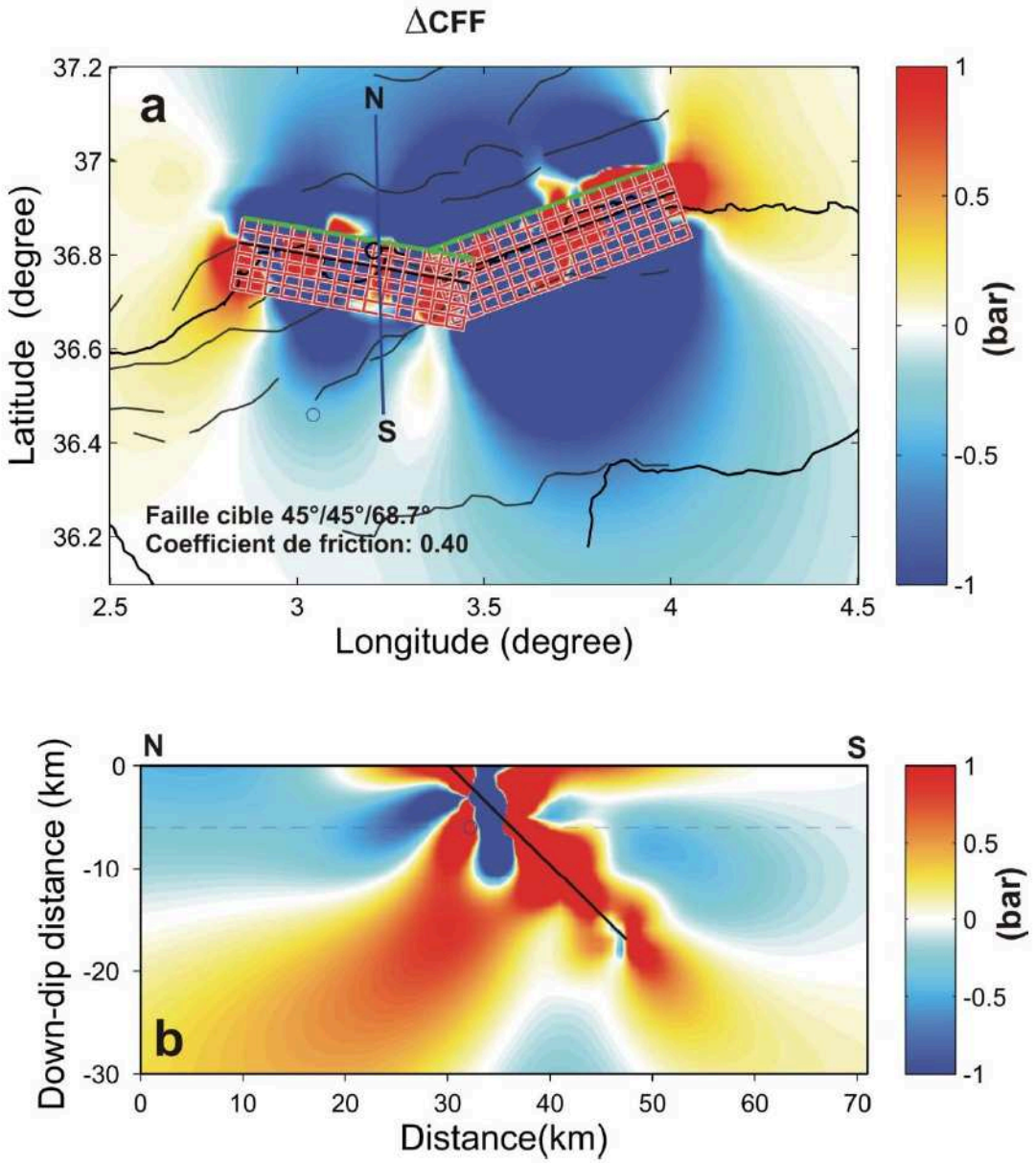


Figure 4-7 : *même légende que dans la Figure (4-3), la* ΔCFF calculée en (a) est à une profondeur de 6 km.

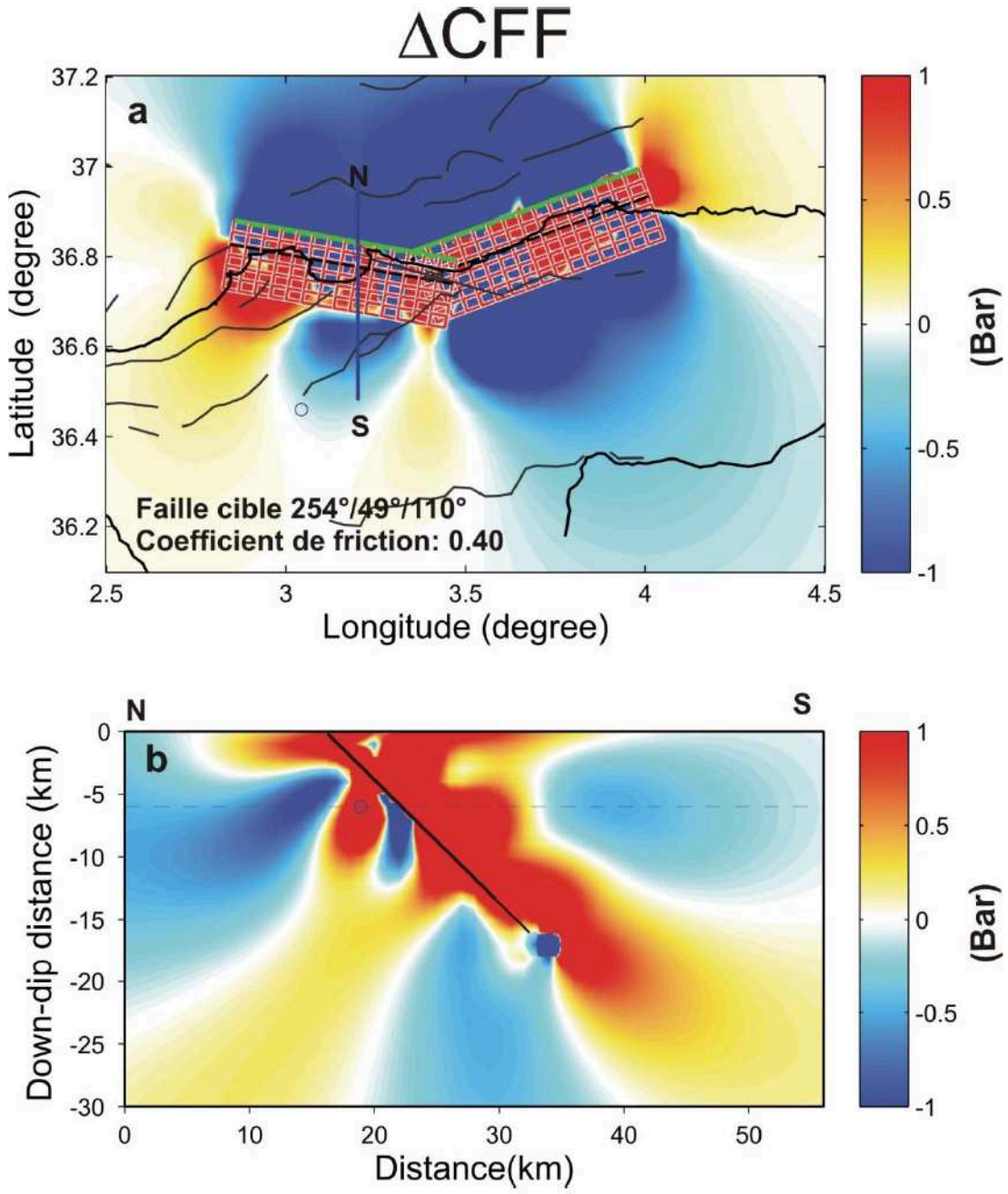


Figure 4-8 : *même légende que dans la Figure (4-3), la* ΔCFF calculée en (a) est à une profondeur de 6 km.

Conclusion

Sur les six plans de failles cibles, nous avons montré que cinq d'entre eux sont favorablement orientés pour avoir été déclenchés par le séisme de Boumerdes de 2003 (notre faille source), comprenant les deux plans du séisme de Bordj Ménaïel, et les deux plans du séisme du séisme de la Baie d'Alger. Pour le séisme de Hammam Melouane, il y a qu'un seul plan qui présente une $\Delta CFF > 0$, mais avec une valeur faible (0.134 bar) comparativement aux deux autres séismes. Nous pouvons dire, (sous-réserve de la validation du modèle de rupture) que ces trois séismes enregistrés entre 2013 et 2014 ont bien été déclenchés par le séisme du 21 mai 2003.

Chapitre 5 : Inversion du tenseur des contraintes en Algérie

5.1. Méthodologie

L'inversion du tenseur des contraintes permet à partir des mécanismes au foyer de déterminer, par une méthode de calcul numérique, les directions des trois contraintes principales, σ1, σ2 et σ3 ; avec σ1 l'axe de la compression, σ3 l'axe de l'extension, σ2 l'axe intermédiaire, dans un espace à trois dimensions. Un facteur de forme R, qui donne la valeur relative de la contrainte intermédiaire σ2 par rapport aux deux autres, est également calculé. La méthode utilisée ici est l'algorithme GRIDSIM_STRESS développé par Delouis et al. (2002).

Lorsqu'un tenseur des contraintes est appliqué sur un plan, il en résulte un vecteur contrainte élastique, égal au produit du tenseur par le vecteur normal au plan. Ce vecteur contrainte peut être décomposé en une contrainte de cisaillement τ et une contrainte normale sur le plan. L'hypothèse émise est que le mouvement de la faille se fait suivant la contrainte de cisaillement sur le plan de la faille. Nous pouvons donc définir un vecteur glissement unitaire :

$$\vec{\gamma} = \frac{\vec{\tau}}{\|\vec{\tau}\|} \qquad (1)$$

Les mécanismes au foyer étant constitués de deux plans, chacun pouvant être le plan de faille, nous devons considérer le vecteur glissement associé à chacun des deux plans nodaux. Le programme GRIDSIM_STRESS va tester différents tenseurs de contraintes, avec pour chacun d'eux, un vecteur glissement calculé sur chaque plan nodal. Pour un mécanisme au foyer donné, le plan nodal associé au plus petit écart angulaire entre le vecteur glissement calculé et le vecteur glissement observé sera sélectionné. Le vecteur glissement observé sur un plan est donné par le mécanisme au foyer, il correspond à la normale de l'autre plan nodal. L'écart angulaire entre les deux vecteurs glissement unitaires (**Figure 5-1**) est quantifié en calculant leur produit scalaire. Un écart moyen, appelé score, est calculé sur l'ensemble des mécanismes au foyer incorporés dans l'inversion. Ce score est utilisé pour évaluer la qualité du tenseur des contraintes testé, le meilleur score possible étant 1.

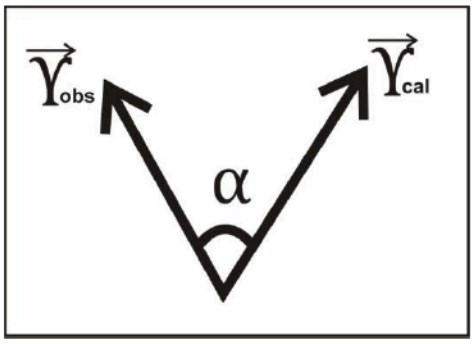


Figure 5-1 : *vecteurs glissement observé γ_{obs} et calculé γ_{cal}, α etant l'écart angulaire entre le glissement observé et calculé.*

$$\overrightarrow{\gamma_{obs}} \cdot \overrightarrow{\gamma_{cal}} = \|\vec{\gamma}_{obs}\| . \|\overrightarrow{\gamma_{cal}}\| . cos\ (\alpha) = cos(\alpha) \quad (2)$$

car les vecteurs glissements sont des vecteurs unitaires.

L'écart moyen sur une population de mécanismes au foyer, qui est le Score, est donné par la relation :

$$Score = \frac{\sum_{i=1}^{n} \vec{\gamma}_{i\,obs} \cdot \vec{\gamma}_{i\,cal}}{n} \quad (3)$$

où n est le nombre de mécanismes incorporés dans l'inversion.

On a :

$\overrightarrow{\gamma_{obs}} \cdot \overrightarrow{\gamma_{cal}} =$ 1 si les vecteurs sont colinéaire et dans le même sens

$\overrightarrow{\gamma_{obs}} \cdot \overrightarrow{\gamma_{cal}} =$ 0 si les vecteurs sont perpendiculaires

$\overrightarrow{\gamma_{obs}} \cdot \overrightarrow{\gamma_{cal}} =$ -1 si les vecteurs sont colinéaires mais de sens opposé.

Le facteur de forme est celui défini par Rivera et Cisternas (1990), R = $(\sigma_z - \sigma_x)/(\sigma_y - \sigma_x)$ avec comme conditions $\sigma_y \geq \sigma_x$ et σ_z l'axe de contrainte le plus proche de la verticale. R varie entre -∞ et +∞ et donne l'information concernant le régime de contraintes. La relation entre σ_x, σ_y , σ_z et les axes principaux σ_1 , σ_2 , σ_3 peut être établie pour n'importe quelle valeur R (Delouis et al., 1993). Par exemple si R Є]0, 1[, alors $\sigma_y > \sigma_z > \sigma_x$ et donc $\sigma_y = \sigma_1$, $\sigma_z = \sigma_2$, $\sigma_x = \sigma_3$, ce qui correspond à un régime décrochant (contrainte σ_2 la plus proche de la verticale, voir **Figure 5-2**. On identifie trois grands domaines de valeurs pour le facteur de forme R : (R < 0), (0 < R < 1) et (R > 1) correspondant respectivement aux régimes de contraintes compressif, décrochant, et extensif.

Les valeurs particulières R = -∞, 0, 1, +∞ correspondent à des cas où deux contraintes sont égales entre elles (cas de contrainte « radiale » ou « uni-axiale » de la **Figure 5-2**).

Dans GRIDSIM_STRESS l'orientation des trois axes de contrainte et le facteur de forme R sont explorés par une combinaison de recherche sur grille et d'algorithme de recuit simulé. L'orientation dans l'espace des trois axes de contrainte est représentée par trois rotations qui correspondent aux angles d'Euler (Φ, θ, Ψ) (Goldstein, 1959), n'importe quelle position dans l'espace de trois axes orthogonaux pouvant être obtenue à partir d'une position initiale (dans notre cas, nous démarrons à partir des positions géographiques, N, E et Verticale). L'algorithme teste des milliers de solutions tenseur, qui sont triées en fonction de leur qualité, donc de leur score. Nous représenterons graphiquement, en projection stéréographique, les meilleures solutions trouvées ainsi que les valeurs du facteur R associées.

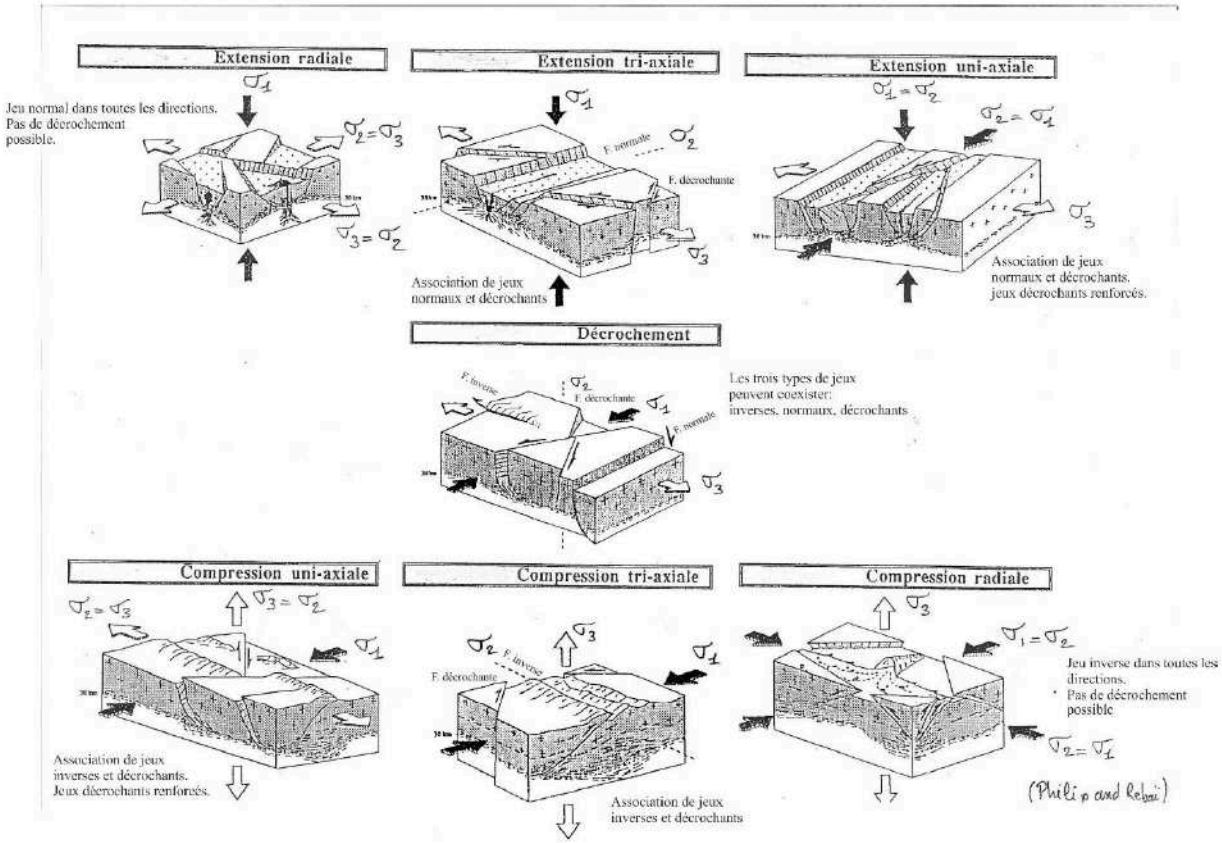


Figure 5-2: *Les différents régimes de contraintes possibles en fonction de l'orientation des trois contraintes principales* σ_1, σ_2, σ_3, *avec la représentation des jeux de faille possibles dans chaque cas (bloc diagrammes). Figure construite à partir de documents non publiés de H. Philip et d'autres éléments dans Rebai (1992).*

5.2. Résultats

Dans ce chapitre, nous allons présenter les résultats de quelques inversions du tenseur de contrainte pour des régions où se sont produits des séismes modérés. Pour cela, et d'une manière générale, nous avons utilisé les mécanismes au foyer disponibles dans chaque région d'étude, en plus des mécanismes au foyer du séisme, de la séquence sismique concernée, ou bien des répliques. Il s'agit de :

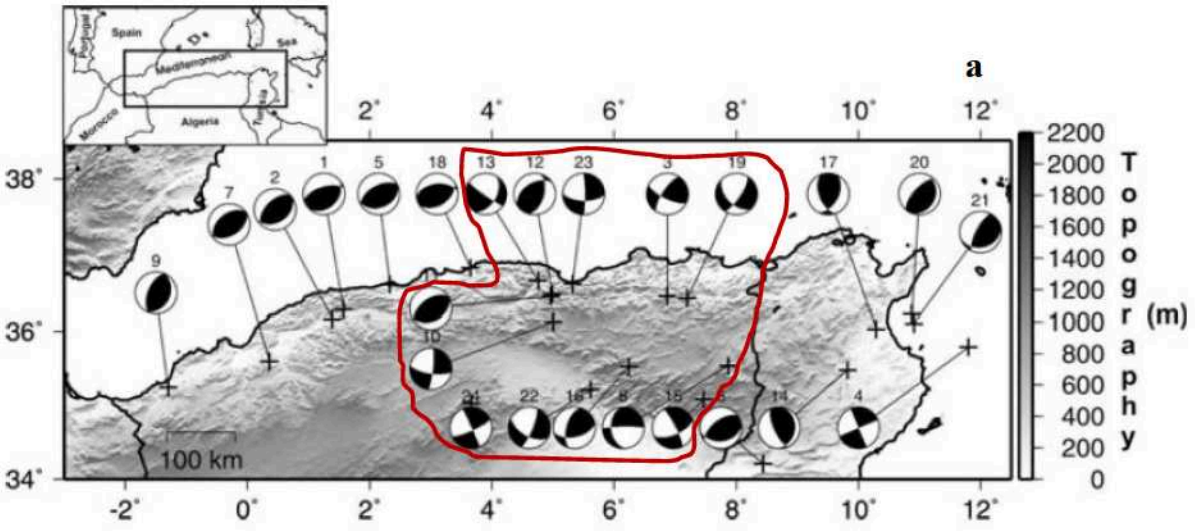
- ✓ Calcul du tenseur de contraintes dans la région des Babors (Est de l'Algérie) à la suite du séisme de Lâalam (Papier en Annexe).
- ✓ Calcul du tenseur de contraintes dans la région du Chélif (Ouest de l'Algérie) à la suite du séisme de Tadjena (papier intégral dans le chapitre 3).
- ✓ Calcul du tenseur de contraintes dans la région de Béni-Ilmane à la suite de la séquence sismique en mai 2010 (papier intégral dans le Chapitre 3).

- ✓ Tenseur de contraintes dans la région de Beni Haroun (Mila, Est de l'Algérie) à la suite de l'activité sismique induite dans cette région causée par un transfert d'eau d'un barrage vers un autre (Papier en Annexe).
- ✓ Tenseur de contraintes calculé dans la région de Boumerdes à la suite du séisme du 21 mai, 2003 (papier en révision).

5.2.1. Inversion du tenseur des contraintes dans la région des Babors en relation avec le séisme de Lâalam

Dans cette inversion du tenseur de contraintes, nous avons sélectionné 13 mécanismes au foyer disponibles dans la région d'étude localisée entre les longitudes 3.7° et 8° et les latitudes 35° et 36.5° **Figure 5-3a**. Les résultats sont illustrés sur la **Figure 5-3b**.

Le régime de contraintes est essentiellement décrochant, avec la contrainte intermédiaire σ2 la plus proche de la verticale, la compression σ1 orientée N325 et l'extension σ3 N232. Les valeurs de σ2 et σ3 peuvent devenir égale dans le cas où R=0. Ce régime décrochant est bien reflété par les mécanismes au foyer, majoritairement décrochants avec des composantes inverses ou normales, et quelques mécanismes nettement plus inverses. C'est l'une des caractéristiques de la sismicité de la région Est de l'Algérie, lente et modérée avec principalement des failles en décrochement, avec présence de deux structures actives importantes, la faille décrochante de Sigus orientée EW et la faille d'Ain Smara d'orientation NE-SW avec un jeu décrochant senestre (Bounif et al., 1987).



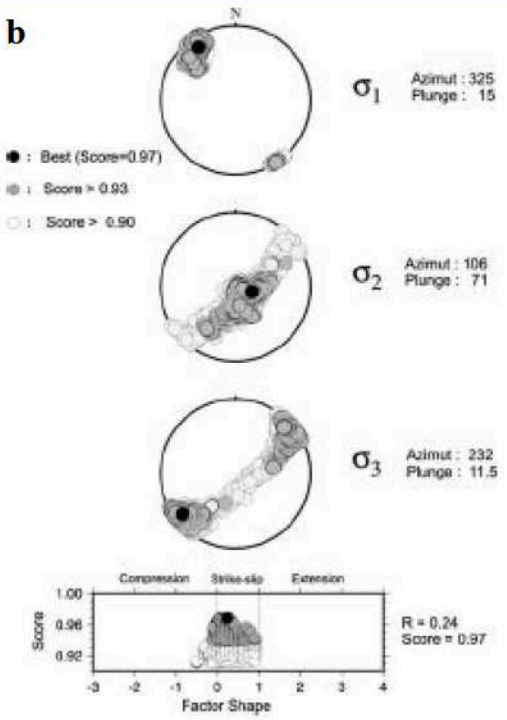


Figure 5-3 : a) *Mécanismes au foyer des principaux séismes en Algérie et en Tunisie. Les Mécanismes au foyer à l'intérieur de l'encadré rouge sont ceux utilisés dans l'inversion du tenseur de contraintes à la suite du séisme de Lâalam (Beldjoudi et al., 2009). b) Résultats de l'inversion du tenseur de contraintes.*

5.2.2. Séisme de Tadjena dans la région du Chelif (El Asnam)

Dans cette inversion du tenseur de contrainte, la population de mécanismes au foyer est présentée dans la **Figure 5-4a**, au nombre de 15. La meilleure solution (score = 0.99) est pour une contrainte compressive σ1 horizontale orientée N325°, σ2 est presque verticale et la contrainte extensive σ3 est horizontale, orientée N°55 (**Figure 5-4b**). Le régime de contrainte est donc préférentiellement de type décrochant (σ2 proche de la verticale), même si des solutions correspondant à un régime compressif (σ3 plus proche de la verticale) existent également avec des scores légèrement plus faibles (**Figure 5-4b**). La préférence pour le régime décrochant est associée au fait que plusieurs mécanismes présentent des composantes normales significatives, ce qui pose des difficultés pour un régime compressif avec σ3 verticale. En fait, les solutions dans le régime compressif (R < 0) correspondent à des σ3 obliques, plus proches de la verticale que σ2, mais en en restant assez éloignés tout de même (**Figure 5-4b**). Un tel régime compressif à axes obliques est moins plausible.

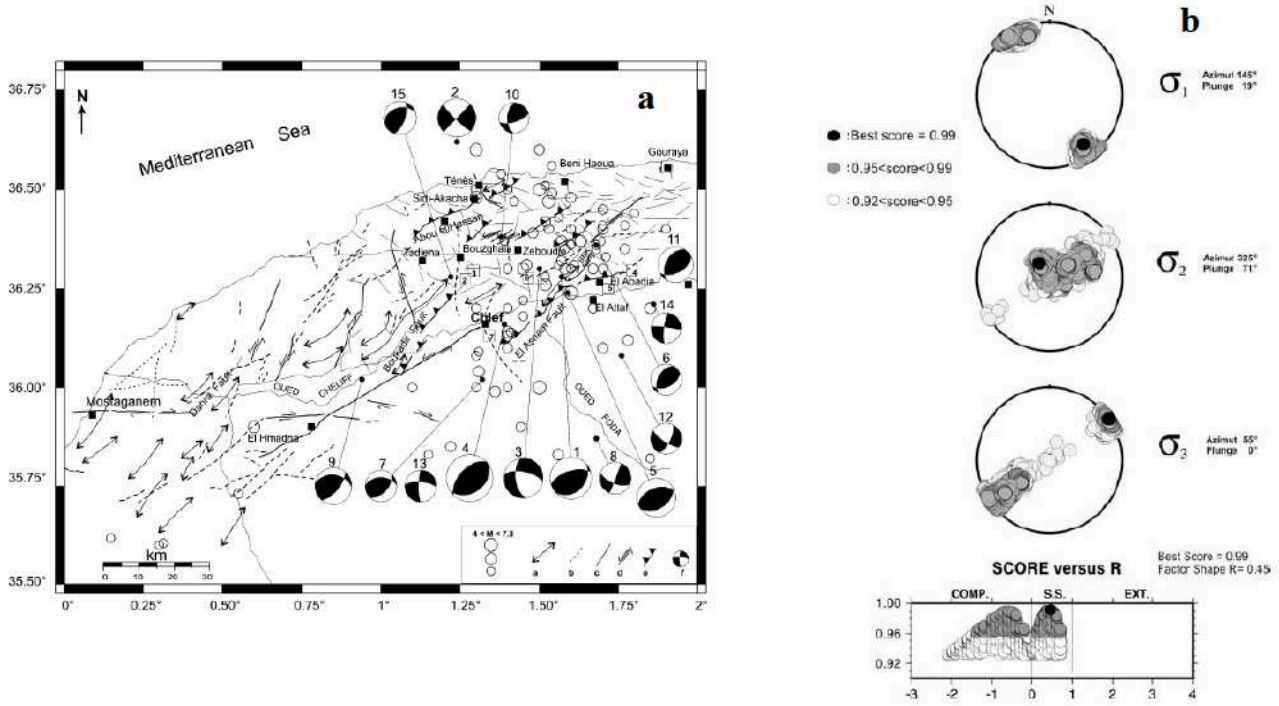


Figure 5-4 : *a) mécanismes au foyer utilisés dans l'inversion du tenseur de contraintes pour la région du Chélif, à la suite du séisme de Tadjena, 2006, Mw 5.0. b) Résultats de l'inversion du tenseur de contrainte.*

5.2.3. Séisme de Béni-Ilmane dans la région des Biban-Hodna

C'est une séquence sismique qui s'est produite près de la localité de Béni-Ilmane dans la région de transition des Bibans-Hodna. Un premier séisme de Mw = 5.5, un deuxième l'a suivi quelques heures après avec Mw 5.1, et un troisième choc de Mw 5.2 a terminé cette séquence laissant la place à des répliques de magnitudes inférieures. Pour cette séquence, nous avons calculé un tenseur de contrainte avec les trois chocs obtenus par inversion des formes d'ondes, les mécanismes au foyer disponibles dans la région et les mécanismes des répliques calculées dans Yelles-Chaouche et al., (2014) (**Figure 5-5a et b**). La meilleure solution obtenue pour le tenseur de contrainte a un score de 0.96 et un facteur de forme R = 0, pour une contrainte de compression σ1 horizontale, orientée N340°, une contrainte intermédiaire σ2 presque verticale et une contrainte en extension, σ3 horizontale orientée N77° (Figure 5-5c). La valeur R = 0 signifie que les contraintes σ2 et σ3 sont égales (compression uniaxiale), ce qui est confirmé par l'étalement des solutions σ2 et σ3 sur le plan perpendiculaire à σ1 (**Figure 5-5c**). Cette situation correspond bien à l'égale abondance de

mécanismes inverses et décrochants, avec peu de composante normale exprimée dans les mécanismes.

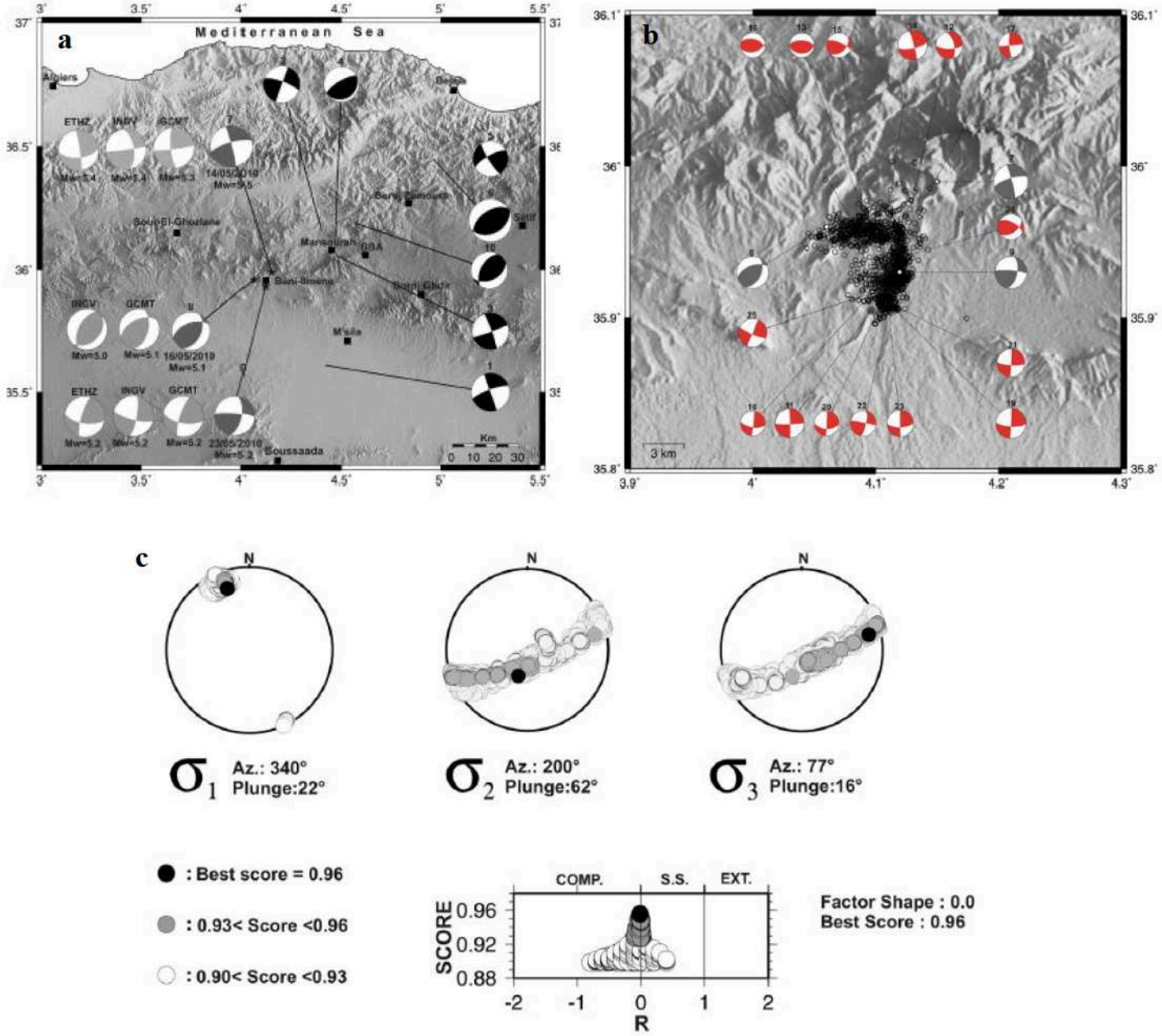
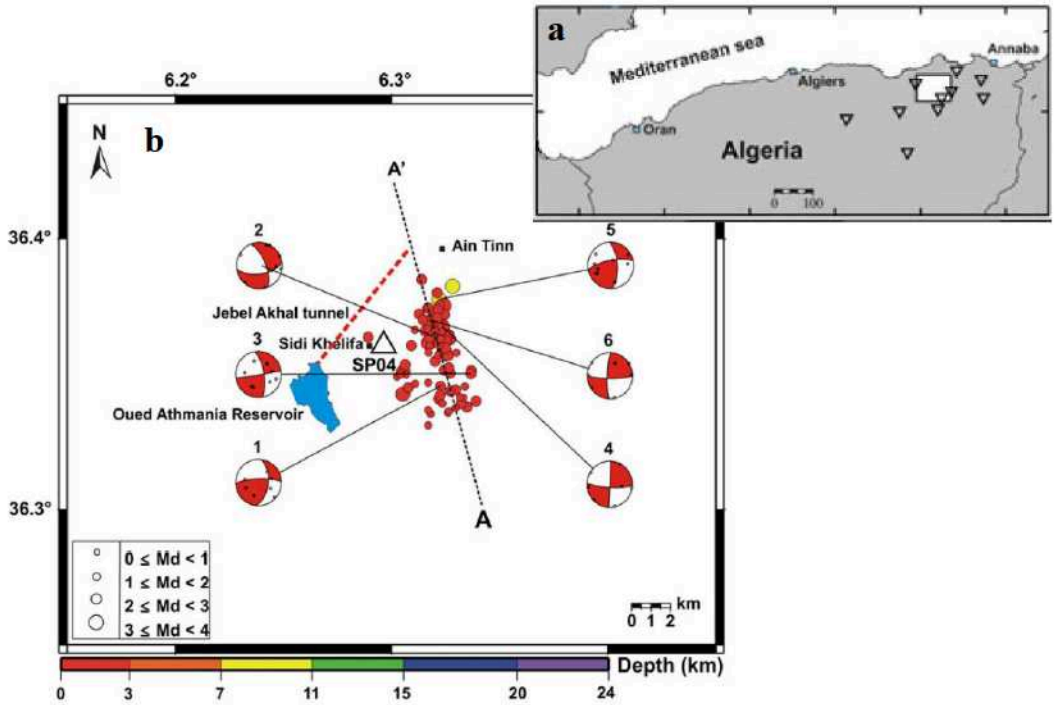


Figure 5-5 : *Mécanismes au foyer utilisés et calcul du tenseur de contrainte. a) Mécanismes des séismes survenus dans la région (noirs). Mécanismes des trois chocs de la séquence de 2010 (gris foncé). Mécanismes des trois chocs calculés par les centres sismologiques mondiaux (gris clair). b) Zoom sur la séquence sismique de 2010, avec les mécanismes au foyer des principales répliques (Rouge) et les mécanismes des trois chocs (gris foncé).* Nous observ*ons aussi deux grandes zones de répliques (cercles ouverts), une orientée EO et l'autre NS, localisées dans Yelles-Chaouche et al., (2014). c) Résultats de l'inversion du tenseur de contraintes.*

5.2.4. Sismicité induite dans la région de Beni Haroun (Mila, Est de l'Algérie)

Ce tenseur de contraintes est calculé pour des mécanismes au foyer de séismes induits, causés par le transfert d'eau à travers des galeries souterraines naturelles entre le barrage de Béni Haroun (1milliars de m^3) et le réservoir de Oued El Athmania dans la région de Constantine dans l'Est algérien (**Figure 5-6a**). Les mécanismes sont au nombre de six, principalement en décrochement (**Figure 5-6b**). La meilleure solution du tenseur de contrainte est obtenue pour un score de 0.99 et un facteur de forme R = 0.5, qui correspond à un régime de contrainte en décrochement (σ2 la plus proche de la verticale). Néanmoins, des solutions valables existent également pour les domaines en compression (R < 0) et en extension (R > 1) (cercles en gris) (**Figure 5-6c**). L'indétermination du tenseur des contraintes vient du fait que tous les mécanismes se ressemblent et sont presque assimilables à un mécanisme décrochant unique. Dans le cas d'un mécanisme unique, la méthode prévoit une distribution de solutions pour σ1 qui remplisse entièrement les cadrans en dilatation du mécanisme, et une distribution de solutions pour σ2 qui remplisse entièrement les cadrans en compression. Nous ne sommes pas très loin de cette situation. On note cependant la tendance relativement marquée à avoir l'axe σ1 orienté NW-SE et σ3 NE-SW.



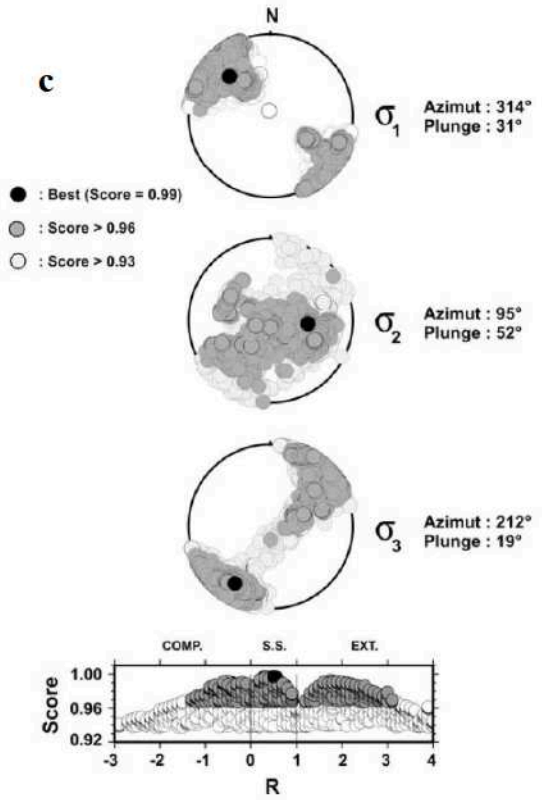


Figure 5-6 : *a) Figure montrant la localisation géographique de la zone d'études (carré blanc) et les triangles montrent la localisation des stations courtes périodes utilisées dans l'étude de cette sismicité induite. b) Mécanismes au foyer utilisés dans l'inversion du tenseur de contraintes avec les polarités des premières arrivées qui ont permis le tracé des mécanismes. Les points rouges montrent l'activité sismique (épicentres).* c) *Résultats de l'inversion du tenseur de contrainte.*

5.2.5. Inversion du tenseur des contraintes pour les répliques du séisme de Boumerdes-Zemmouri de 2003

Dans ce travail, nous avons calculé le tenseur de contrainte dans la région de Boumerdes-Zemmouri, à la suite du séisme du 21 mai 2003. 78 mécanismes au foyer des répliques de ce séisme ont été utilisés pour le calcul de tenseurs. La région d'étude a été divisée en 5 zones suite à l'interprétation de la répartition des répliques dans Kherroubi et al., (2017). Pour chaque zone nous considérés ses mécanismes au foyer, présentés sur la **Figure 5-7**, et nous avons lancé une inversion du tenseur de contrainte. A la fin, nous avons obtenu quatre tenseurs pour chacune des zones, 2, 3, 4 et 5, un tenseur global a été calculé qui incorpore l'ensemble des 78 mécanismes.

Nous n'avons pas calculé de tenseur de contraintes pour la zone 1, car elle comportait trop peu de mécanismes au foyer.

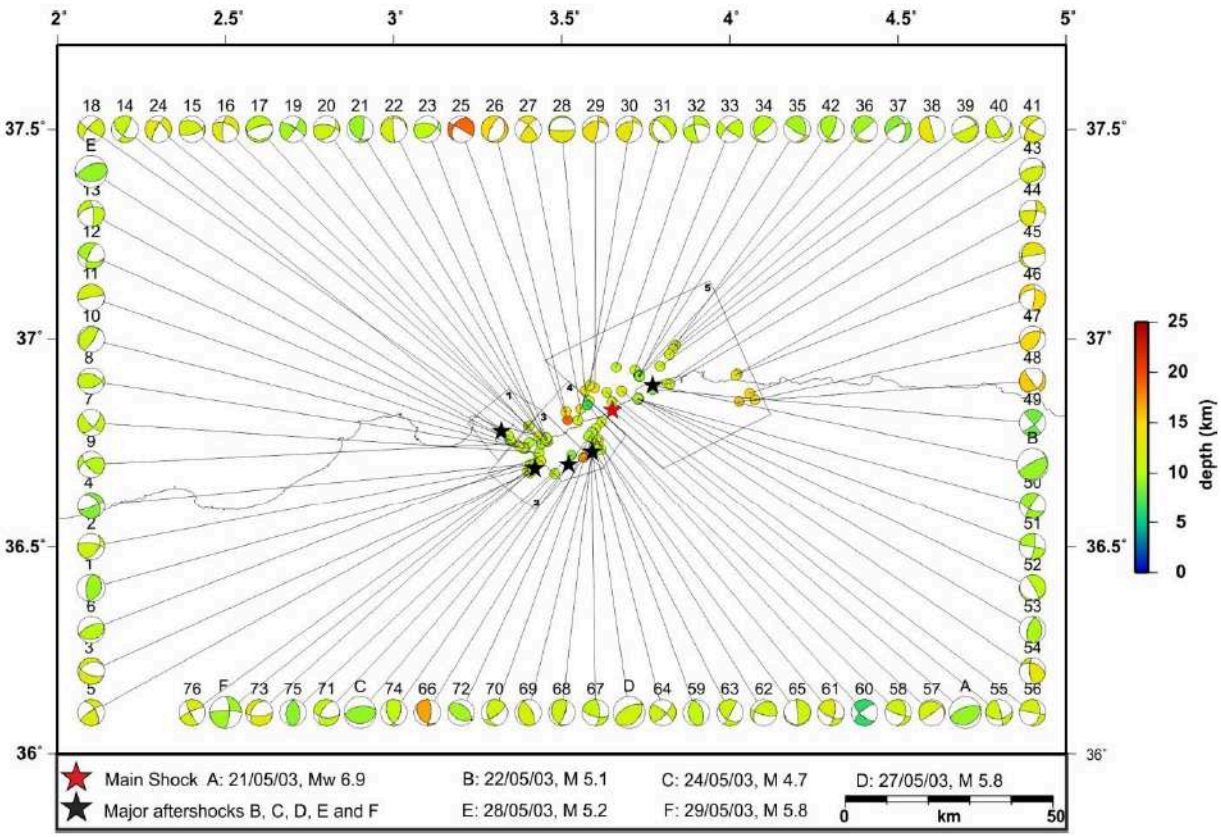


Figure 5-7 : *78 mécanismes au foyer utilisés dans l'inversion des tenseurs de contrainte à la suite du séisme de Boumerdes du 21 mai 2003. Les 5 zones considérées sont encadrées, cahcune avec son numéro. (Figure dans Kherroubi et al., 2017).*

Zone 2

Dans cette Zone 2 la contrainte principale σ1 est horizontale orientée ~EO (N98°), la contrainte σ3 est proche de la verticale (plongement =55°) et orientée (N227°), la contrainte σ2 est horizontale. La combinaison de ces trois contraintes a donnée comme meilleure solution un Score de 0.67, pour un facteur de forme R de -3, correspondant à un régime compressif (σ1 horizontale et σ3 verticale) comme présenté sur la **Figure 5-8** (partie gauche). Ce résultat peut paraitre surprenant, mais il reflète l'existence de mécanismes inverses dont les plans nodaux sont proches de NS, et de mécanismes décrochants avec des plans NE-SW dextres et des plans NW-SE senestres. Il s'agit d'orientations inhabituelles pour ces jeux de faille pour le nord de l'Algérie, que nous pouvons supposer reliés aux perturbations de contraintes générées par le choc principal de Boumerdes. Le score global assez faible (0.67) traduit probablement des variations locales du tenseur des contraintes qui resteraient à analyser.

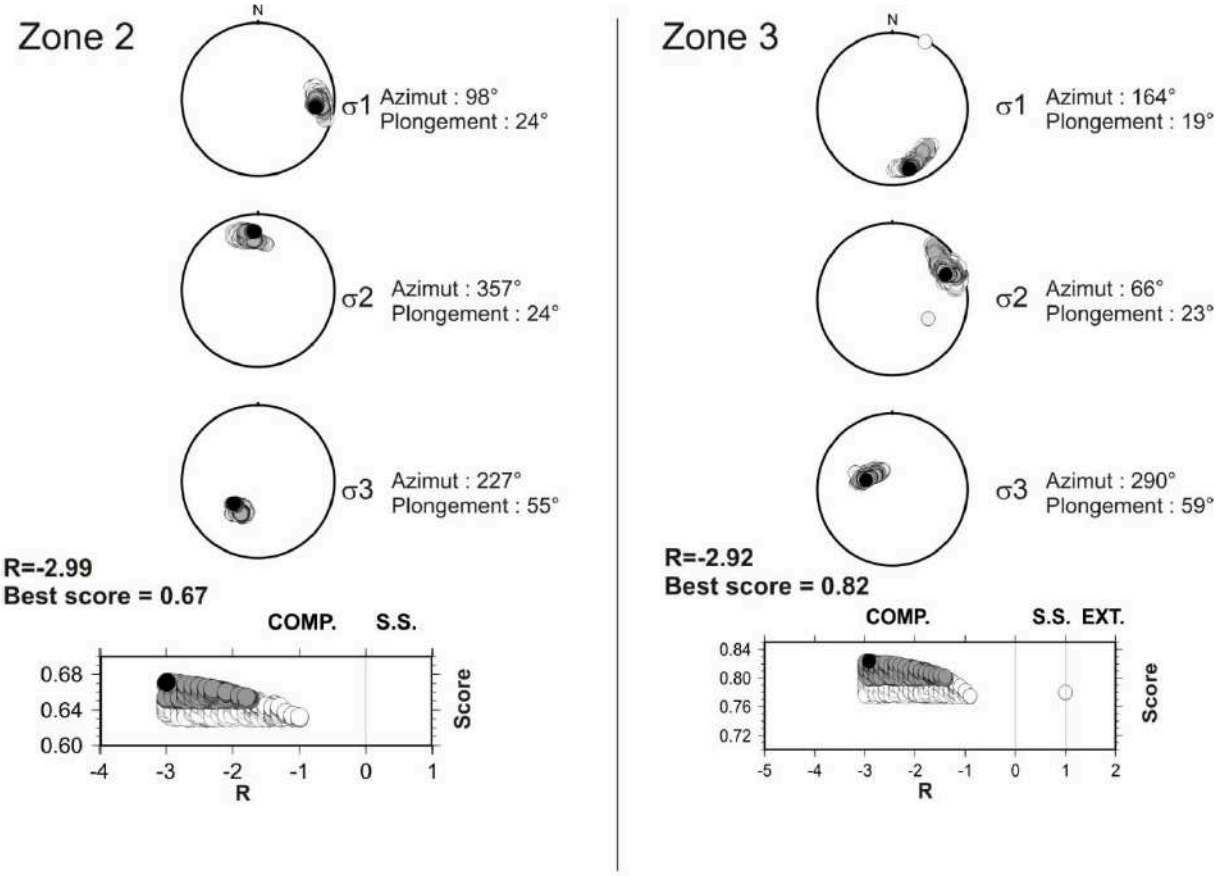


Figure 5-8 : *Résultats de l'inversion du tenseur de contrainte. A gauche tenseur de contraintes pour la Zone 2 ; à droite tenseur de contrainte pour la Zone 3.*

Zone 3

Pour cette Zone, nous retrouvons un tenseur plus classique pour le nord de l'Algérie, avec σ1 horizontal orienté N164° et σ3 l'axe le plus proche de la verticale (**Figure 5-8**). Le facteur de forme est celui d'un régime compressif (R = -2.9). Les mécanismes au foyer inverses et décrochant ont tendance à reprendre leurs orientations classiques pour la région, c'est-à-dire des plans inverses NE-SW et des plans décrochants plus proches de NS et EW (**Figure 5-7**).

Zone 4

Le tenseur trouvé dans cette Zone est celui d'un régime compressif, avec un facteur de forme R =-2.2, la contrainte σ1 horizontale orientée N140°, une contrainte σ3 presque verticale, et σ2 horizontale (**Figure 5-9**). Cette Zone est mitoyenne de la Zone 3, le tenseur y garde une direction proche, avec toutefois une légère rotation antihoraire de 24°.

Zone 5

Dans cette Zone, on retrouve un tenseur de contrainte du type de celui de la Zone 2, avec une contrainte σ1 horizontale orientée EO et σ3 verticale, pour un régime compressif (**Figure 5-9**). Mêmes remarques que pour la Zone 2 concernant la relation probable avec des perturbations de contraintes du choc principal et le score bas manifestant une certaine hétérogénéité.

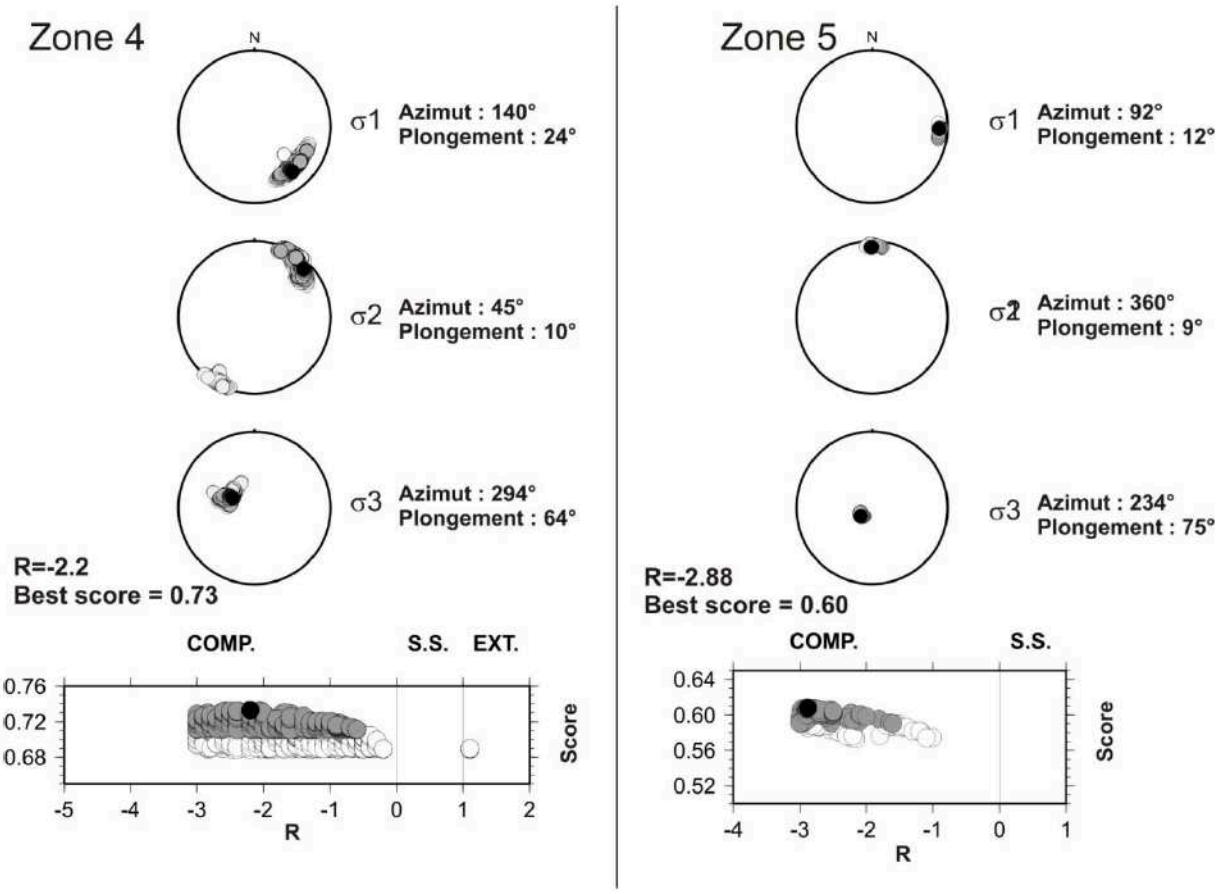


Figure 5-9 : *Résultats de l'inversion du tenseur de contraintes. A gauche tenseur de contraintes pour la Zone 4, à droite tenseur de contraintes pour la Zone 5.*

Tenseur global

Le résultat de l'inversion avec l'ensemble des 78 mécanismes a donné comme meilleure solution un facteur de forme R = -3.0 correspondant à un régime compressif, avec une contrainte compressive σ1 horizontale et orientée N319°, une contrainte intermédiaire σ2 horizontale orientée N45°, et une σ3 proche de verticale (**Figure 5-10**). Le tenseur global (moyen) reflète donc la tendance générale du rapprochement des deux plaques tectoniques Afrique-Eurasie. Toutefois, le score très bas de 0.54, qui correspond à une déviation angulaire

moyenne de 57° ($\cos^{-1}$ 0.54) entre les vecteurs glissement calculés et observés, reflète de fortes hétérogénéités, comme le montre les résultats divers trouvés pour les zones 2, 3, 4, 5 individuellement. Il n'est donc pas totalement judicieux de prendre en considération ce tenseur moyen.

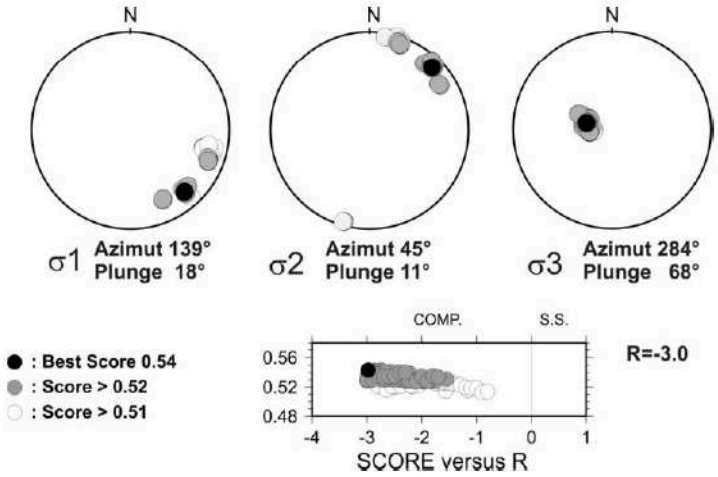


Figure 5-10 : T*enseur de contraintes obtenu pour une inversion globale des 78 mécanismes au foyer.*

5.2.6. Discussion-comparaison avec autres résultats publiés sur l'ensemble de l'Algérie

Dans cette partie de la thèse, nous avons présenté les différents résultats de calcul de tenseur de contraintes dans différentes régions d'Algérie, (1) La région des Babor, des Biban et le constantinois (est-algérien), (2) la région du Chélif (ouest-algérien) et (3) la région d'Alger-Boumerdes

Les tenseurs de contrainte que nous avons calculés pour des régions bien à l'intérieur des terres correspondent à un régime de contrainte qui oscille entre compressif ($R < 0$) et décrochant ($0 < R < 1$), voire exactement intermédiaire dans le cas R=0, avec comme constante une direction de σ1 proche de l'horizontale et orientée NW-SE à NNE-SSE, en accord avec la direction de convergence Afrique – Eurasie. Les axes σ2 et σ3 peuvent alternativement être horizontal ou vertical, en s'échangeant.

Pour la zone côtière nord, nous nous basons sur les répliques du séisme majeur de Boumerdes, ce qui complique les choses. Toutefois, il en ressort un tenseur plus nettement compressif ($R < 0$), avec σ3 vertical. L'axe σ1 est horizontal avec une tendance globale NW-SE en accord

avec la convergence des plaques, mais il peut être localement perturbé (Zones 2 et 5) ce qui est attendu au voisinage des zones de fort glissement d'un séisme majeur.

L'inversion du tenseur de contraintes dans ces différentes régions montre que la contrainte principale σ1 a une direction NW-SE à NNW-SSE et que sa valeur varie entre N314° et N340°. Cette direction est comparable avec celle des autres études menées dans différentes régions du Nord algérien (Figure 5-11). Ousadou et al. (2012) en exploitant les mécanismes au foyer des répliques du séisme de Constantine de 1985 ont trouvé une contrainte σ1 horizontale orientée N348°, et σ3 horizontale orientée N257° favorisant les failles en décrochement. Ayadi et al. (2008), lors de l'inversion de 30 mécanismes au foyer des répliques du séisme de Boumerdes du 21 mai 2003, ont trouvé une contrainte principale σ1 horizontale, orientée N350°, et une contrainte σ3 verticale, ce régime de contrainte compressif favorable pour des failles inverses. Bounif et al. (2003) lors de leur étude du séisme du Mont Chenoua (Ouest d'Alger) de 1989, ont trouvé une contrainte principale σ1 horizontale orientée NS et une contrainte σ3 verticale, indiquant également un régime compressif dans la région. Des études régionales concernant le calcul des tenseur de contrainte montrent que σ1 est d'une direction moyenne NW-SE (Negredo et al., 2002 ; Henares et al., 2003). Serpelloni et al., (2007) ont trouvé que les vecteurs déplacements au nord de l'Algérie maintiennent une orientation NW-SE, ainsi que les autres études menaient par McClusky et al., (2003), Calais et al., (2003) et les résultats du Modèle Nuvel-1a.

En conclusion nos résultats d'inversion du tenseur de contrainte pour des régions bien précises adhèrent bien avec la cinématique globale du rapprochement Afrique-Europe, et les résultats obtenus donnent des précisions supplémentaires par rapport aux modèles globaux.

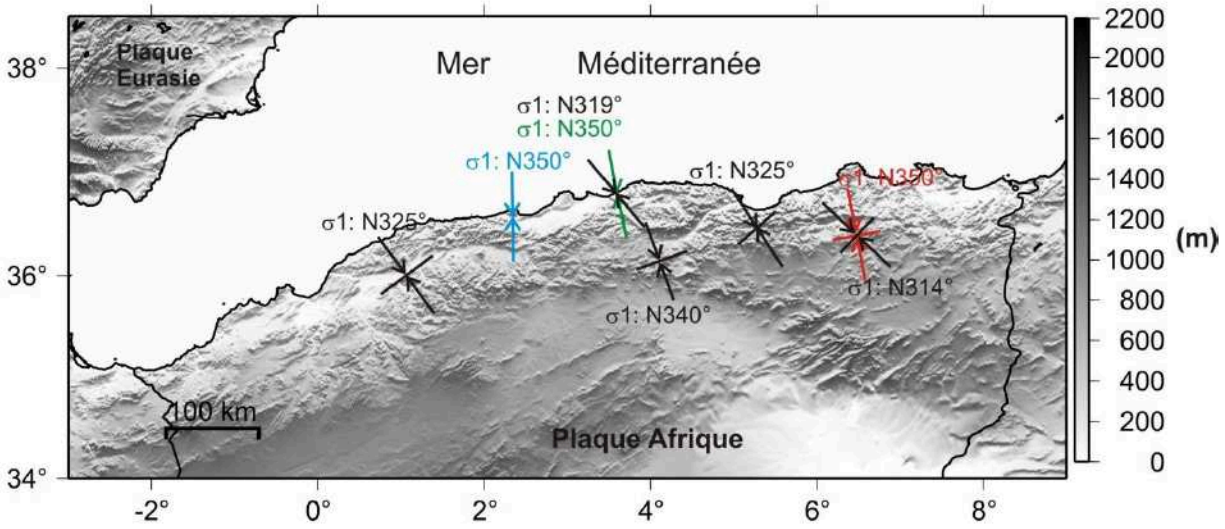


Figure 5-11 : *Carte qui montre les orientations des différents tenseurs de contraintes obtenus dans différentes régions du nord algérien, menées par plusieurs auteurs. En noir, les résultats présentés dans cette thèse (Beldjoudi et al.,2009, 2012 et 2016 ; Semmane et al., 2012, Kherroubi et al., 2017). En rouge, tenseur obtenu par Ousadou et al., (2012). En vert, tenseur obtenu par Ayadi et al., (2008). En bleu, tenseur obtenu par Bounif et al., (2003).*

Conclusion et perspectives

Cette thèse a eu pour objet l'étude de la modélisation de la source sismique des séismes du Nord de l'Algérie en exploitant les données sismologiques et géodésiques. Plusieurs résultats peuvent être retracés dans cette conclusion.

Depuis l'installation du réseau de surveillance sismologique algérien (ADSN), des calculs automatiques de mécanismes aux foyers de séismes de magnitude M > 3.8 sont devenus réalisables, permettant de construire petit à petit un catalogue plus complet en termes de mécanismes. Les mécanismes au foyer de séismes modérés calculés dan cette thèse sont principalement de type 1) inverses orientés généralement NE-SO ou 2) de décrochement senestre pour le plan NS et dextre pour le plan EO. Le séisme de Tadjena (Mw 5.0, 2006) dans la région du Chélif, s'est produit sur la faille de Boukadir située à quelques 30 km de la faille d'El Asnam (Oued Fodda) lieu du séisme de 1980. La faille de Boukadir présente les mêmes caractéristiques que celle d'El Asnam (longueur, largeur, orientation, plongement) et donc nous pouvons dire que potentiellement elle peut générer un séisme de même magnitude que celui de 1980.

Nous avons vu aussi que lors de la séquence sismique de Béni-Ilmane de 2010 (centre – nord de l'Algérie), des mécanismes au foyer en faille inverse et en décrochement sont déterminés. La séquence s'est produite au niveau d'un pli-faille orienté NE-SO à NE. Bien que cette séquence sismique fût modérée, ses effets se sont répercutés sur les constructions et sur la surface où nous avons observé des fissures sur les maisons de la région ainsi que des chutes de pierres, des fissures sur la surface du sol, cela dans la région épicentrale. Au travers de la carte géologique présentée dans le papier du séisme de Béni-Ilmane, nous avons montré que le cœur du pli-faille est très fracturé (tectonisé) et est bordé par des failles actives pouvant générer des séismes modérés. Si de futurs séismes se produisent sur ce pli-faille, on s'attend à ce qu'ils se manifestent sous forme de séquences sismiques, étant donné la proximité des failles formant un réseau dense.

Nous avons aussi présenté les résultats d'inversion des formes d'ondes de quelques séismes modérés survenus entre 2013 et 2016. Il s'agit des séismes de Hammam Mélouane (2013, au sud d'Alger), Bordj Ménaïél (2014, au Sud de Boumerdes) et M'ziraa (sud algérien). Ce dernier de magnitude 5.1 s'est produit à la zone de transition entre l'Atlas saharien et la

plateforme saharienne, avec un mécanisme en faille normale orientée Est-Ouest. Au travers des cartes de sismicité réalisées depuis l'installation du réseau (ADSN), nous observons que cette région de M'ziraa, présente une sismicité modérée avec un taux plus faible que celui du Tell Atlas, mais à l'avenir il faudra prévoir une densification du réseau de surveillance sismologique dans ce secteur, afin de suivre plus en détail la sismicité de cette région.

Une autre partie majeure de cette thèse concerne l'inversion conjointe des données télésismiques, accélérométriques, et géodésiques pour le calcul du glissement cosismique du séisme de Boumerdes-Zemmouri de 2003. Nous notons que c'est le premier travail où l'ensemble des données disponibles est utilisé conjointement pour le calcul du glissement. Un modèle de source composé de deux segments de failles a été trouvé, avec un fort glissement réparti suivant deux aspérités sur le segment orienté N70°, et deux aspérités de plus faible glissement sur le segment orienté N100°. L'activation de segments de faille superficiels plus éloignés en mer est possible, mais difficile à prouver en raison du manque de données proches. Nous avons aussi calculé, le changement de la contrainte de Coulomb sur des failles cibles (séismes de Bordj Ménaïel, Hammam Melouane et de la Baie D'Alger), produit par le glissement cosismique associé au séisme de Boumerdes-Zemmouri. Ce calcul a montré que parmi les six plans nodaux (deux pour chaque séisme), cinq se trouvaient dans des régions chargées positivement, c'est-à-dire favorables à leur déclenchement.

Nous avons calculé également des tenseurs de contraintes dans différentes régions du nord algérien, cela suite aux séismes de Laâlam (2006), Tadjena (2006), Béni-Ilmane (2010), Boumerdes (2003) et Béni-Haroune (2007). L'étude a montré que le tenseur au nord de l'Algérie peut être divisé en deux grandes familles : (1) un premier tenseur purement compressif avec une contrainte $\sigma 1$ horizontale et $\sigma 3$ verticale, présent dans la région centre (Alger, Boumerdes, Tipaza). Il produit principalement des failles inverses. Et, (2) un tenseur à tendance uniaxiale, avec une contrainte $\sigma 1$ horizontale et $\sigma 2 = \sigma 3$, présent dans la région Est algérienne et dans la région du Chélif, qui favorise non seulement des failles inverses mais aussi le jeu des failles en décrochement. Ces résultats de tenseurs de contraintes se corrèlent bien avec la convergence Afrique-Eurasie.

L'ensemble des approches et méthodes mises en œuvre durant la thèse, me permettront d'analyser en détail les futurs séismes qui se produiront en Algérie, qu'ils soient modérés ou plus forts.

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Annexe 1: The 22 February 2014 Mw 4.1 Bordj-Menaïel Earthquake, Near Boumerdes-Zemmouri, North-Central Algeria

by F. Semmane, B. Y. N. Benabdeloued, H. Beldjoudi, and A. K. Yelles-Chaouche

Dans ce papier, j'ai calculé la source sismique du séisme modéré en inversant les formes d'ondes sismiques.

The 22 February 2014 M_w 4.1 Bordj-Menaïel Earthquake, Near Boumerdes-Zemmouri, North-Central Algeria

by F. Semmane, B. Y. N. Benabdeloued, H. Beldjoudi, and A. K. Yelles-Chaouche

INTRODUCTION

A number of destructive earthquakes have occurred in the region of Algiers, the capital of Algeria. Two well-known historical earthquakes were recorded in 1365 and 1716 (Harbi *et al.*, 2004). A more recent one, the 2003 M_w 6.9 Boumerdes earthquake (also called the Zemmouri earthquake), took place in the Boumerdes district (Fig. 1), about 50 km east of Algiers (Yelles-Chaouche *et al.*, 2003). The Boumerdes earthquake occurred on a blind-thrust fault offshore from the Algerian coast. The rupture was along a northeast–southwest-striking fault plane, although the precise fault trace was unknown. During this earthquake, an average value of 55 cm of uplift of the shoreline was observed (Meghraoui *et al.*, 2004), consistent with the thrust focal mechanism, at least in the high-slip region on the fault plane (Semmane *et al.*, 2005).

More recently, in the same district (Boumerdes), an earthquake of moment magnitude M_w 4.1 (M_d 4.3) occurred on 22 February 2014 at 20:30 local time, 12 km southeast of the epicenter of the 2003 Boumerdes earthquake, and at a depth of 7–11 km. The epicenter was located 5 km northeast of the central part of the town of Bordj-Menaïel and 5 km northwest of the town of Naciria (Fig. 2).

From macroseismic investigations, the maximum intensity of the Bordj-Menaïel earthquake was estimated as V on the 1998 European Macroseismic Scale (EMS-98) near the towns of Bordj-Menaïel and Naciria, and the earthquake was felt up to 90 km from the epicenter. However, no significant damage occurred. A search of the Algerian earthquake catalog yielded six previous earthquakes of $2.0 < M_d < 4.0$ recorded since 2002 that took place within 10 km of the epicenter of the 22 February 2014 earthquake. No active faults have been identified in this area.

The Bordj-Menaïel earthquake is of interest because it was located very close to the epicenter of the 2003 Boumerdes earthquake, close to the strike-slip Thénia fault (Fig. 2). This fault is a near-vertical east–west-trending structure that is believed to have played a role in the arrest of the rupture to the west during the 2003 Boumerdes earthquake (Semmane *et al.*, 2005). Furthermore, the 2014 Bordj-Menaïel event may provide a better constraint on the distribution of crustal deformation and the local seismic-hazard map in this region. In this study, we provide the precise location of the Bordj-Menaïel earthquake, the focal mechanism, and an estimate of the source parameters. These data will improve our understanding of this tectonically complex area.

GEOLOGICAL SETTING AND SEISMOTECTONICS

The region of interest is in the eastern part of the Mitidja basin, a Neogene basin in the central Tellian Atlas. The Tellian Atlas comprises a series of east–west to northeast–southwest-trending folds, associated active faults, and Quaternary basins, all oriented parallel to the shoreline. The region is tectonically active due to convergence between the African and Eurasian plates. Déverchère *et al.* (2005) provided the first detailed tectonic framework of the area offshore from the Boumerdes region using high-quality data acquired during the Maradja cruise in September 2003. Two cumulative scarps identified on the slope and at the foot of the margin coincide with the locations of the two main slip patches of the 2003 Boumerdes earthquake (e.g., Delouis *et al.*, 2004).

Before the occurrence of the 2003 Boumerdes earthquake, the seismicity in the region was linked to the vertical Thénia fault (Glangeaud, 1952; Boudiaf, 1996). This fault represents the eastern limit of the Mitidja basin, and it extends offshore from Algiers (e.g., Mauffret, 2007). Previous earthquakes likely occurred on this fault, including the 1953 Boudouaou event (intensity V, Medvedev-Sponheuer-Karnik [MSK] intensity scale; Boudiaf *et al.*, 1998), the 23 May 1982 Isser event (**M** 4.9; Mokrane *et al.*, 1994), and the 1987 Thénia event (**M** 4.5; Boudiaf *et al.*, 1998). Since the development of the permanent Algerian Digital Seismic Network (ADSN), which involved the installation of a number of new seismic stations starting in 2007 (Yelles-Chaouche *et al.*, 2013), the number of seismic events recorded in Algeria has increased significantly. At least one moderate event of magnitude $M_d > 4$ is recorded per year in addition to the permanent background seismic activity (Fig. 1).

Our region of interest comprises post-Miocene alluvium deposits and marine terraces (Glangeaud, 1952). The bedrock east and west of the epicenter of the 2003 Boumerdes earth-

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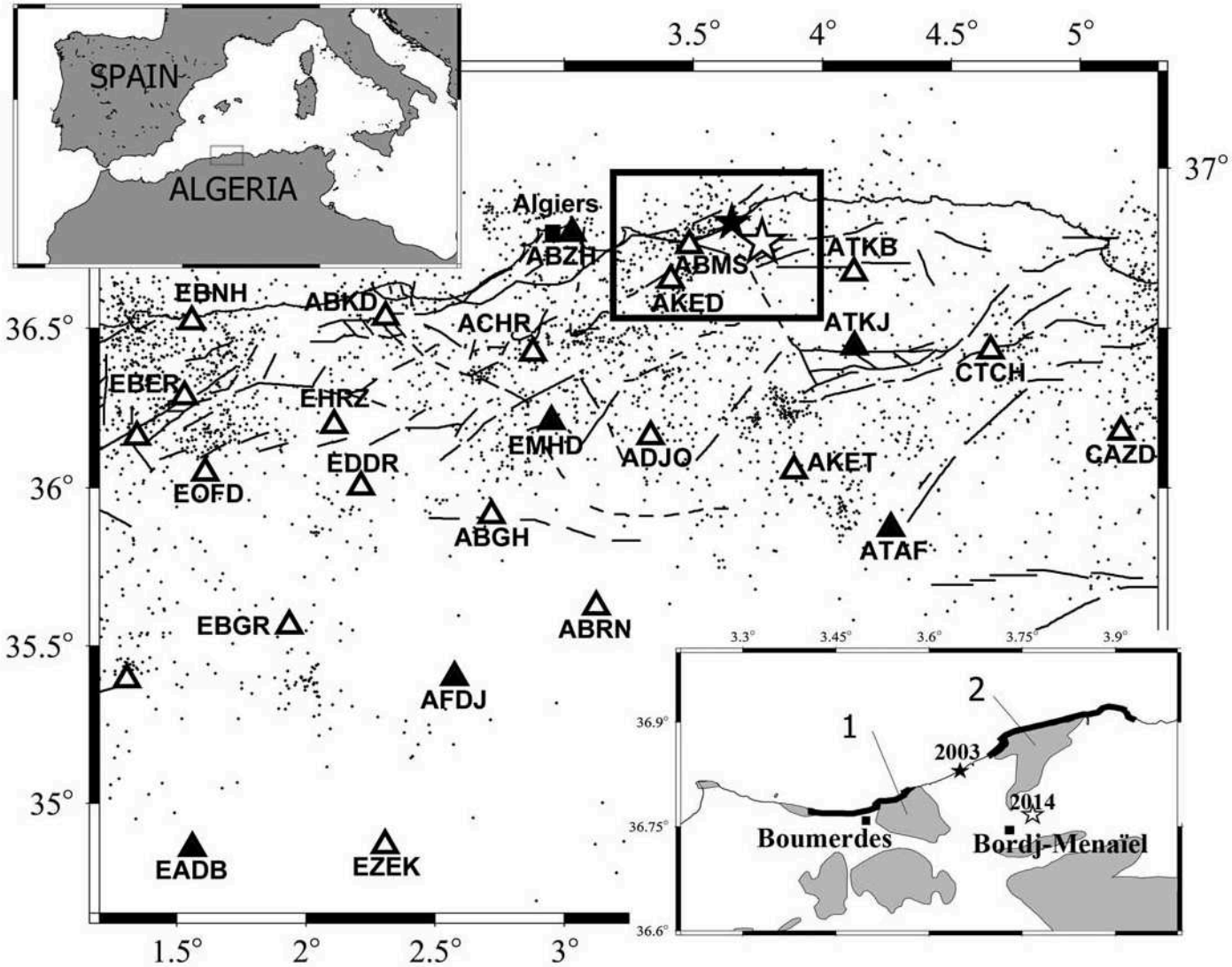


▲ **Figure 1.** The studied area and location of stations that recorded the earthquake used in this study (closed triangles are broadband stations and open triangles are one-component short-period stations). Dots show the seismic epicenters from 1992 to March 2014; the 2003 Boumerdes epicenter is represented by the black star, and the Bordj-Menaïel earthquake is represented by the white star. The black lines are active faults mapped by Guemache (2010). The inset shows the basement in the region in gray. The locations of the numbered blocks 1 and 2 coincide with the two asperities found in the slip distribution models and also with the uplift zones (two thick lines on the coast).

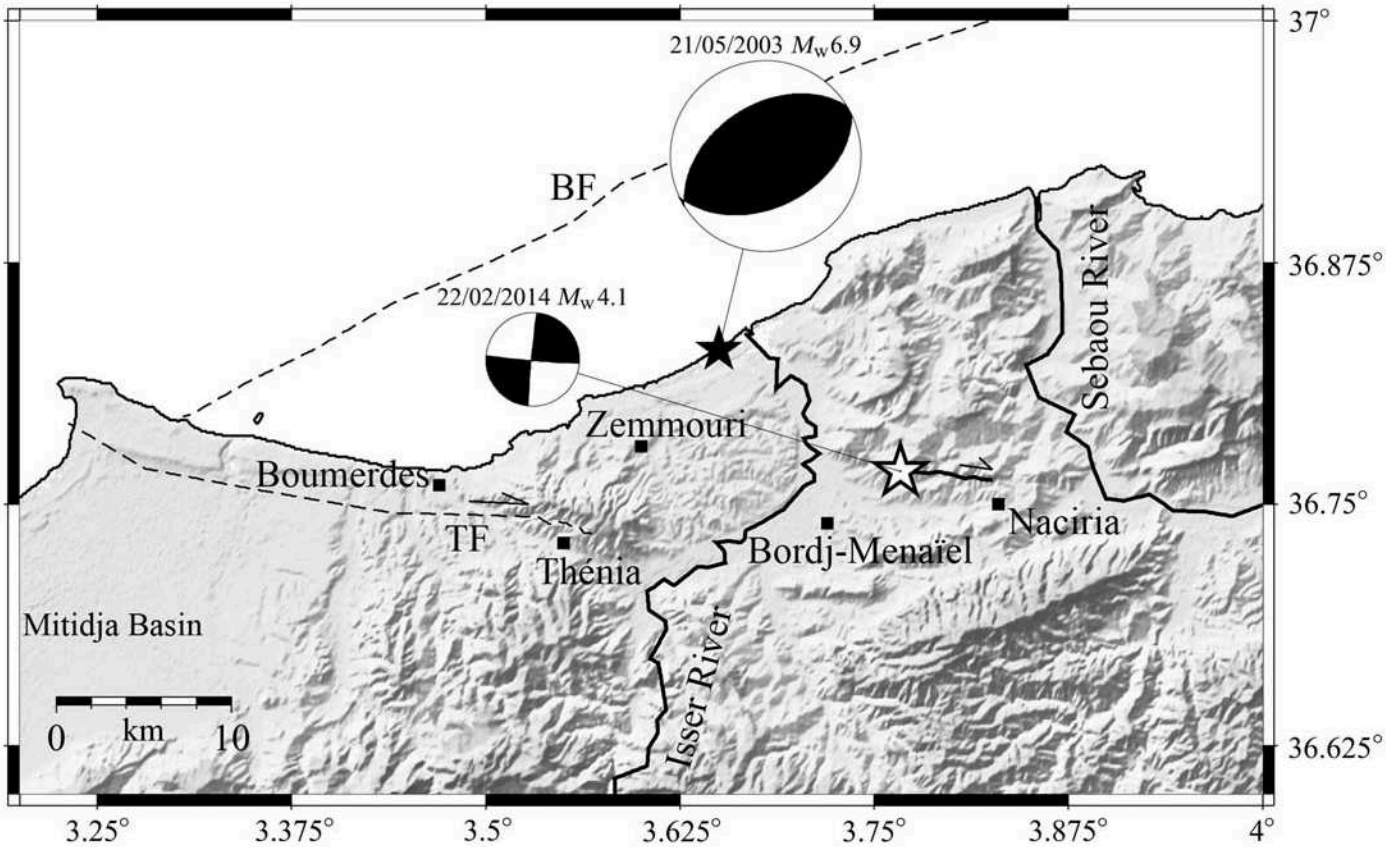
quake, comprises two blocks of igneous rocks (Fig. 1, inset), the locations of which coincide with two asperities identified in slip distribution models (e.g., Delouis *et al.*, 2004; Semmane *et al.*, 2005) and zones of uplift (Meghraoui *et al.*, 2004).

SEISMIC NETWORK AND DATA

Almost all the seismological stations installed north of Algeria recorded the Bordj-Menaïel earthquake. The nearest station of the permanent network is located ~25 km from the epicenter. This event was well recorded by 28 seismological stations of the ADSN. In addition, one station (MAHO) from the Spanish Seismological Network was used to cover the gap northeast of the epicenter. All of these stations are located at distances of 25–350 km from the epicenter, and they comprise 9 broadband stations (STS-2, Q330) and 20 one-component short-period stations (Ranger SS-1 seismometers and Q330 digitizers), among which 5 were collocated with strong-motion sensors (Episensor). The stations with acceptable signal-to-noise ratios (SNRs) were used to determine the epicenter and the focal mechanism of the earthquake. Six of the nine broadband stations (20 Hz sampling) were used to calculate the source parameters.

EARTHQUAKE LOCATION AND FOCAL MECHANISM

We tested different earthquake location programs and different velocity models using SEISAN software (Havskov and



▲ **Figure 2.** The white star is the epicenter of the 2014 Bordj-Menaïel earthquake and its focal solution. The black star is the 2003 Boumerdes earthquake epicenter and its focal solution. The continuous thick black line near the city of Naciria is the probable trace of the fault responsible of the 2014 earthquake. The two dashed lines are the supposed Boumerdes fault (BF) and the Thénia fault (TF). The Isser and Sebaou Rivers are also shown.

Ottemoller, 1999). We show only the best solutions obtained using the HYPOINVERSE location program (Klein, 2002). The velocity model (Table 1) is used for routine earthquake locations in Algeria. The best solution for the hypocenter of the Bordj-Menaïel event, obtained using data from the 29 stations (28 *P*-wave and 2 *S*-wave arrival times), locates the mainshock at 5 km from center of the town of Bordj-Menaïel (Fig. 2), at 36.767° N, 03.767° E, and 11 km depth. The root mean square (rms) and the horizontal and vertical errors of the location estimate are 0.12 s, 0.90 km, and 1.1 km, respectively. The epicenter is also located 5 km northwest of the town of Naciria, east of Bordj-Menaïel.

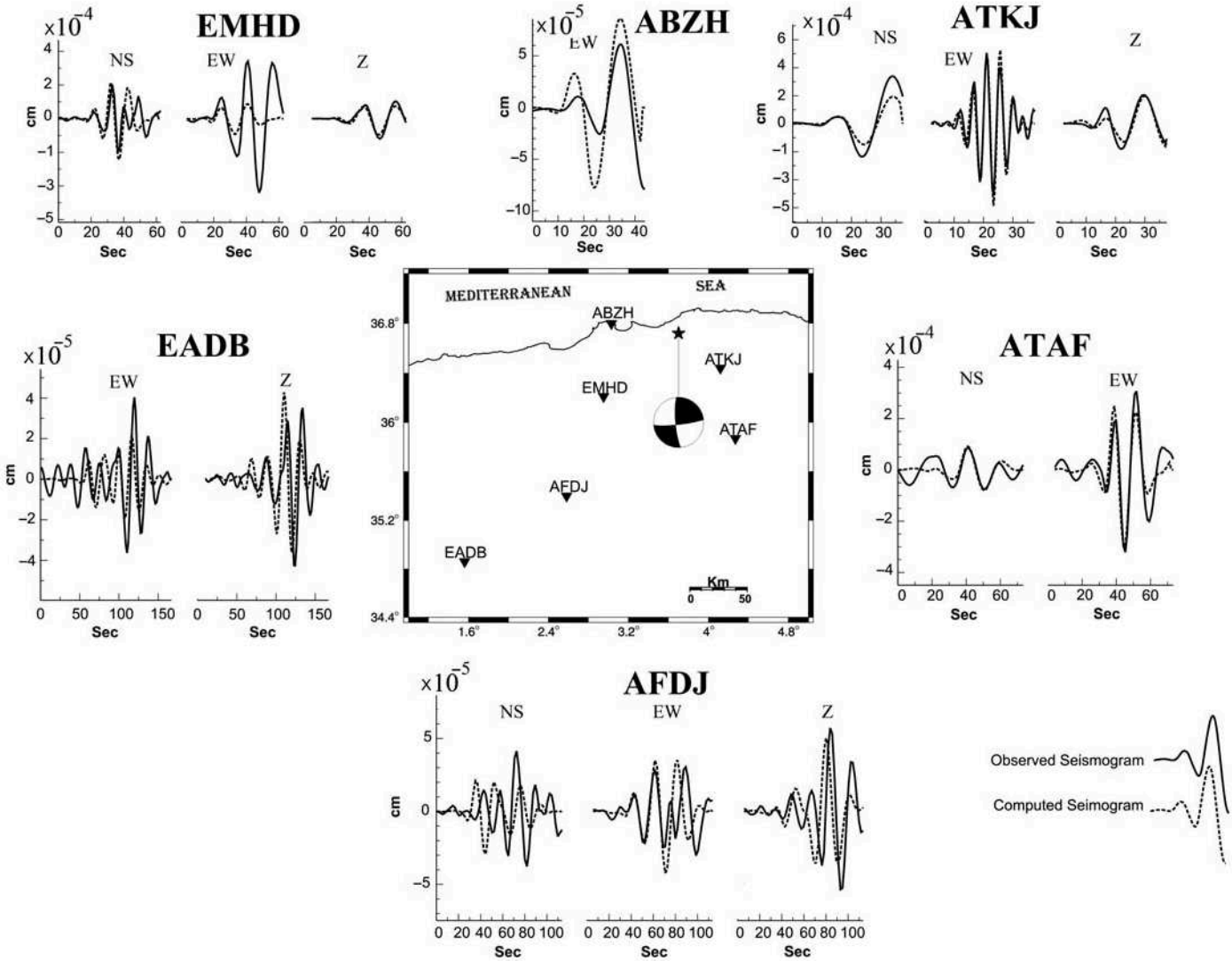
Table 1
1D Velocity Model Used in This Study

V_P (km/s)	V_S (km/s)	Z (km)
4.5	2.6	0
5.5	3.2	9
6.5	3.8	20

V_P, *P*-wave velocity; V_S, *S*-wave velocity; Z, depth to top of layer.

The focal mechanism of the Bordj-Menaïel earthquake was obtained using two approaches: the recorded *P*-wave first motions from the 29 stations using FPFIT software (Reasenberg and Oppenheimer, 1985) and broadband full waveform inversion using the FMNEAR method (Delouis, 2014). One of the two nodal planes (Fig. 2) suggested by the *P*-wave first-motion polarity is near vertical and strikes east–west (N95°, 88°, 178°), corresponding to a right-lateral strike-slip fault under regional northwest–southeast tectonic compression. The estimated directions of compression (P axis) and extension (T axis) are N140°, 75° and N45°, 70°, respectively, very similar to the regional average (Beldjoudi *et al.*, 2009). The second nodal plane is oriented north–south (N185°) and corresponds to a left-lateral strike-slip fault.

The primary source parameters determined using the FMNEAR (Delouis, 2014) include the moment magnitude and the double-couple focal mechanism (strike, dip, and rake). The corresponding focal mechanism solution consists of a grid search on the strike, dip, and rake parameters with a point source assumption (Fig. 3). Six near-source broadband records were integrated to displacement and band-pass filtered between 0.03 and 0.3 Hz. Synthetic seismograms were computed using the discrete wavenumber method of Bouchon (1981) and



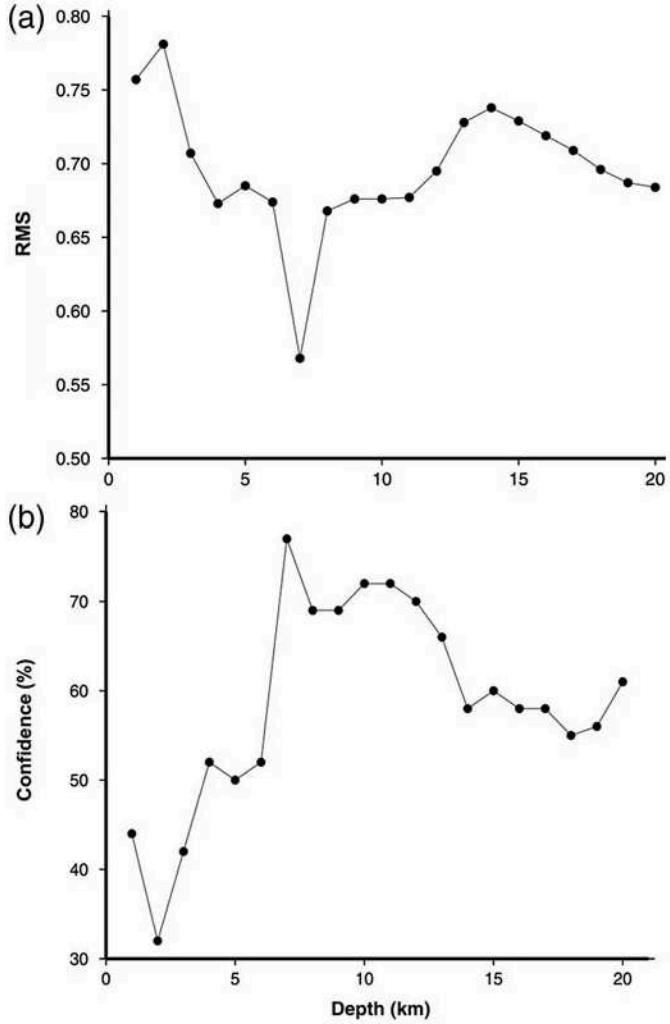
▲ **Figure 3.** Focal mechanism of the event that occurred on 22 February 2014, obtained using waveform modeling (the FMNEAR method in Delouis, 2014). The observed and computed seismograms are represented in displacement, in continuous and dashed black lines respectively, and are band-pass filtered between 0.03 and 0.3 Hz. The star shows the epicenter location, and the triangles represent the broadband stations used in the inversion. There are missed components, ABZH (NS and Z), EADB (NS), and ATAF (Z), which presented a high misfit and were discarded automatically by the program. In addition to some possible problems in the waveform data (which was not the case in this study) that could cause outliers in the results, the velocity model, the source model, the *P*-wave picking, and the epicenter location could also be at the source of the erroneous results.

filtered in the same way. The best-fitting parameters of the first plane are [strike, dip, rake] = [84°, 83°, 170°] and those of the second plane are [175°, 80°, 7°]. The focal depth was considered within the 1–20 km depth range by assuming the epicenter was well located. The best solution for depth is 7 km, and the quality of the solution for the focal depth is shown in Figure 4. The confidence index (CI), which is scaled between 0% (no confidence) and 100% (total confidence), is an empirical indicator of the quality and uniqueness of the solution. The CI is 78% for the best solution on the focal depth, and the corresponding rms is 0.56.

The absence of well-located aftershocks does not allow the true fault plane to be distinguished from the auxiliary plane. However, geological (and geomorphological) evidence enables the fault plane to be identified. The east–west fault-plane solution is consistent with the topographic lineament observed north of Bordj-Menaïel and Naciria on a digital elevation model (DEM), aerial photographs, and high-resolution satellite images (Fig. 2). The geological structure related to the topographic lineament is almost parallel to the Thénia fault and perpendicular to the Boumerdes fault.

EARLY AFTERSHOCKS

To identify potential early aftershocks (Peng *et al.*, 2007), we examine a 30 Hz high-pass-filtered seismogram recorded by the



▲ **Figure 4.** Quality of the depth solution. (a) Root mean squares (rms) versus depth. (b) Confidence index (CI) versus depth. These two parameters are detailed in Delouis (2014). The best scores (CI = 78%, rms = 0.56) are found at 7 km depth.

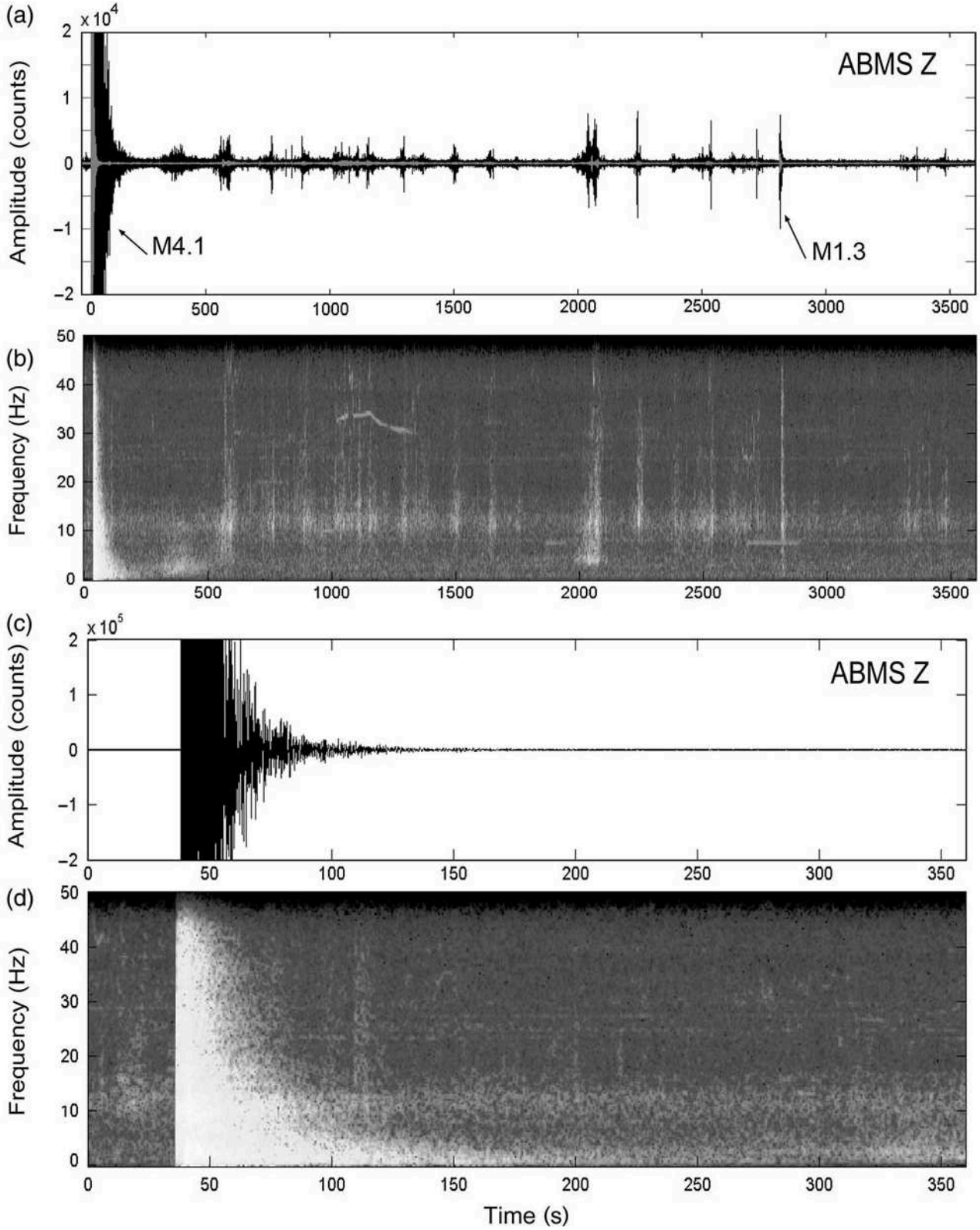
closest station ABMS in the permanent seismic network (Fig. 5a). Located at about 25 km from the epicenter, this station includes a vertical-component SS-1 short-period seismometer coupled with a Q330 digitizer. The sampling rate is 100 samples/s. The first event (magnitude M_d 1.3) occurred about 46 min (~2800 s) after the mainshock. This event is not listed in the standard catalog because it was only recorded with a good SNR by station ABMS. Because this station is installed in an urbanized area, it often records different types of noise that are well observed on both the seismogram and the spectrogram (Fig. 5a,b). An enlarged plot (Fig. 5c,d) does not show any additional events in the first 5–6 min after the mainshock. About 21 hrs after the mainshock, another small event of magnitude M_d 1.8 is observed (Fig. 6). These two events (46 min and 21 hrs after the mainshock) may be considered as aftershocks of the Bordj-Menaïel earthquake, because the *S-P* times of the two events are very close to that of the mainshock. In addition, the waveforms of the *P*-wavetrains are quite similar (Fig. 6), indicating that the hypocenters and the focal mechanisms probably are very close.

SOURCE PARAMETERS

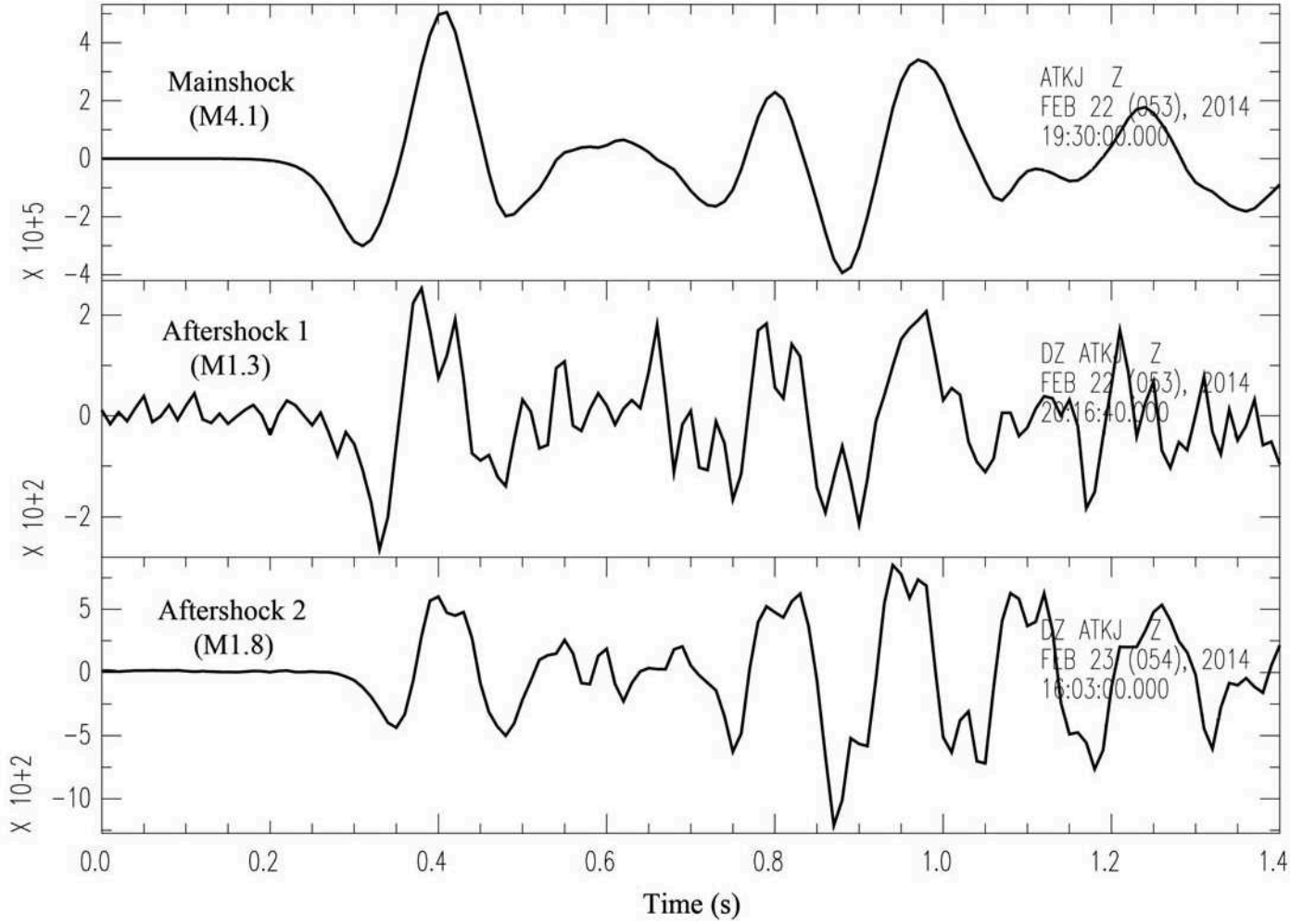
The source parameters of the Bordj-Menaïel earthquake (Table 2) were estimated from the *P*-wave displacement spectra following a simple circular rupture model (Brune, 1970, 1971; Hanks and Wyss, 1972). We use seismic data of six broadband stations with the best SNRs. We selected the vertical component in order to minimize the influence of site effects. The seismograms were first baseline corrected and detrended. The onset times of *P*-wave arrival were visually identified. We extracted the window (*P*-wavetrains, about 5 s) for all the components with a Hanning taper window (Fourier transformed using FFT), and instrumentally corrected; the spectra were then smoothed by applying the mean (average) smoothing algorithm from the Seismic Analysis Code (SAC; Goldstein *et al.*, 2003). The usable frequency range extends from ~0.02 to 10 Hz. At higher frequencies, we are limited by the sampling rate. The corner frequencies (f_c), the source dimension (radius r), the seismic moment (M_0), the stress drop ($\Delta\sigma$), and the average displacement on the fault plane ($\bar{u}$) were calculated. The moment magnitude was calculated at all six broadband stations following Hanks and Kanamori (1979). The other source parameters are as follows: corner frequencies range from 1.3 to 3.5 Hz; the source dimension (radius) ranges from 655 to 1912 m; the seismic moment from 0.45×10^{15} to 4.1×10^{15} N·m, which corresponds to $M_w = 3.7$–4.4; the stress drop is 0.35–64.3 bar; and the average displacement is slightly more than 3 cm.

The seismic moment and the moment magnitude were also obtained by waveform inversion (Fig. 6). The results ($M_0 = 1.58 \times 10^{+15}$ N·m, $M_w = 4.07$) are similar to those obtained using the *P*-wave displacement spectra. Furthermore, the moment magnitude is consistent with the magnitude computed by the German GeoForschungsZentrum (GFZ), using data from regional seismic networks.

The *P*-wave displacement spectra approach is somewhat qualitative because it often underestimates the corner frequency if the spectrum is not corrected for attenuation (e.g., Lanza *et al.*, 1999). One of the effective methods that more accurately estimate the source parameters is the empirical Green's function (EGF) technique (e.g., Lanza *et al.*, 1999). These two aftershocks, as shown in Figure 6, would have been good candidates for the EGF method if they have been well recorded with good SNR by the broadband stations considered in this study. The best SNR (SNR > 3) is observed at the ATKJ, the nearest broadband station (48 km epicentral distance) in the 3–13 Hz frequency range. The small difference between the waveforms in Figure 6 is the long-period content of the mainshock. Figure 7a shows the displacement spectra for the mainshock and the largest aftershock observed at the ATKJ station. The spectral ratio of the two events shows a possible corner frequency of the mainshock between 2 and 5 Hz



▲ **Figure 5.** (a) 30 Hz (gray) and 0.5 Hz (black) high-pass-filtered vertical-component velocity seismograms recorded at the ABMS station within 1 h of the Bordj-Menaïel mainshock. The mainshock signal was clipped for a better view. (b) The spectrogram computed from the 0.5 Hz high-pass-filtered data (Peng *et al.*, 2011). The first event (magnitude M_d 1.3) occurred about 46 min (~2800 s) after the mainshock. Because the ABMS station is installed in an urbanized area, it often records different types of noise that are well observed on both the seismogram and the spectrogram. (c) An enlarged view of the first 5–6 min of raw data after the mainshock. (d) The spectrogram of (c). No early aftershock is observed.



▲ **Figure 6.** The raw velocity *P* waveforms of the Bordj-Menaïel mainshock and the two aftershocks recorded at the ATKJ station. The amplitude is in counts.

(Fig. 7b), consistent with the direct estimation of 3.5 Hz with the mainshock displacement spectra alone (Table 2).

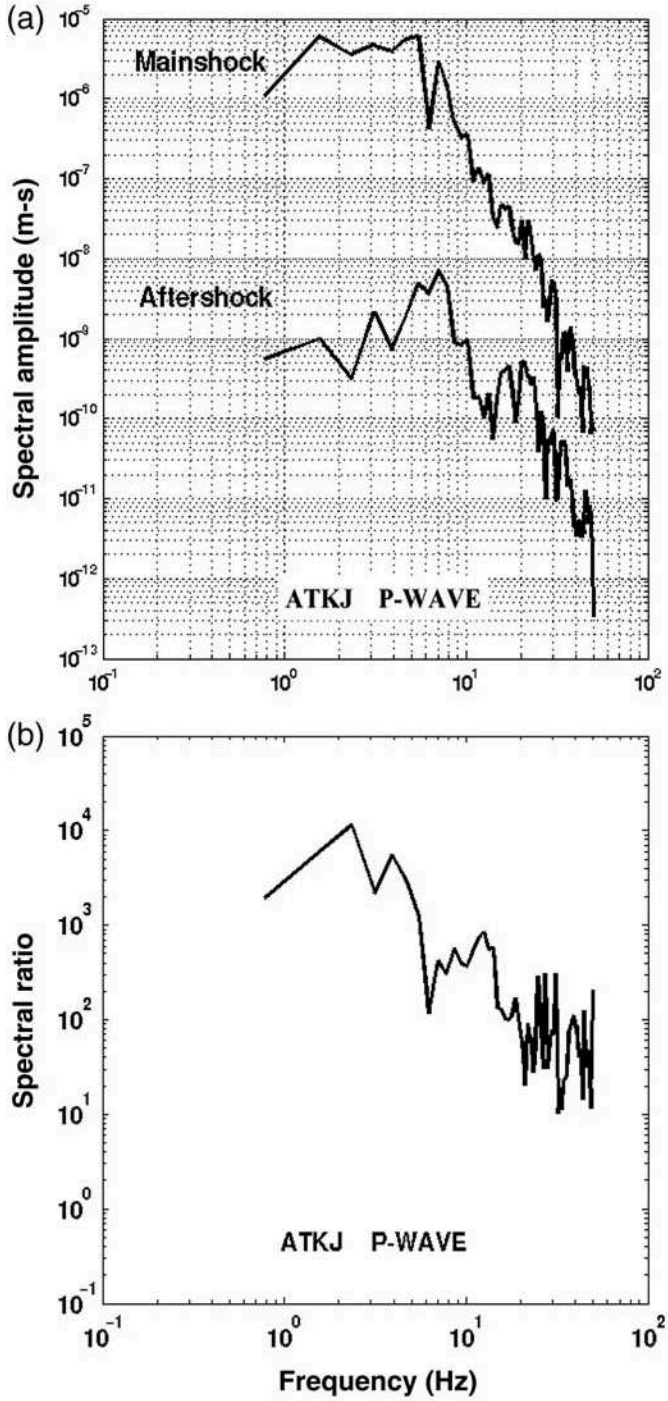
DISCUSSION AND CONCLUSION

Our study of the moderate-size Bordj-Menaïel earthquake showed that useful information can be extracted from small and moderate events if they are well recorded. Until recently, studying small-magnitude earthquakes (M ≤ 4) in Algeria was difficult or even impossible due to a lack of accurate data. The development of the Algerian Digital Seismological Network allows more accurate recording of moderate earthquakes. Consequently, more active faults can be identified, improving the assessment of the seismogenic potential in this region.

Table 2
Estimated Source Parameters of the Bordj-Menaïel Earthquake

Station	ΔEp (km)	Hyp (km)	f_c (Hz)	r (m)	A_0 (ms)	M_0 (N·m)	$\Delta\sigma$ (bar)	$\bar{u}$ (cm)	M_w
ABZH	66	66.9	2.2	1043	3.0×10^{-6}	3.09×10^{15}	12.0	3.0	4.3
AFDJ	187	187.3	1.5	1529	$0.6e \times 10^{-6}$	1.70×10^{15}	2.13	0.78	4.1
ATAF	109	109.6	1.3	1912	1.0×10^{-6}	1.69×10^{15}	1.06	0.49	4.1
ATKJ	48	49.2	3.5	655	4.0×10^{-6}	4.10×10^{15}	64.3	10.0	4.4
EADB	290	290.2	1.3	1765	0.1×10^{-6}	0.45×10^{15}	0.35	0.15	3.7
EMHD	97	97.6	3.2	717	1.5×10^{-6}	2.30×10^{15}	26.9	4.6	4.2

ΔEp, epicentral distance; Hyp, hypocentral distance ; f_c, corner frequency; r, source radius; A_0, spectral plateau level; M_0, seismic moment; $\Delta\sigma$, stress drop; $\bar{u}$, average displacement; M_w, moment magnitude.



▲ **Figure 7.** (a) *P*-wave displacement spectra at the ATKJ station for the mainshock and the largest aftershock (**M** 1.8). The spectral peaks and troughs match quite well between the mainshock and the aftershock. (b) The spectral ratio.

The epicenter of the Bordj-Menaïel earthquake was located northeast of the town of Bordj-Menaïel, between the Neogene Mitidja and Tizi-Ouzou basins. The very few detailed geological maps of the epicenter area do not show much information on the locations of any active faults. Consequently, we used a DEM and other aerial imagery to locate the likely fault responsible for the earthquake. The epicenter region, as observed on the DEM, contains an east–west-trending lineament that is also observed on aerial photographs and satellite images. This lineament is several kilometers long and is subparallel to the Thénia active fault. Saadi *et al.* (2005) considered that this fault influences the drainage diversion of the Sebaou River to the east. Similarly, Boudiaf *et al.* (1998) attributed the drainage diversion of the Isser River, near the Sebaou River, to movement on the Thénia fault. Moreover, the region contains several other lineaments of 4–8 km in length that strike either northeast–southwest or northwest–southeast. The lineament interpreted in the epicenter region could extend to offshore of the Zemmouri area and be > 40 km long.

Guemache (2010) produced a seismotectonic map showing faults with similar strike (60°–90° N) to those suggested by the focal mechanism in this study. Based on the 3D aftershock distribution and gravity data, Ouyed *et al.* (2011) also produced a seismotectonic model of the Boumerdes region, suggesting the existence of an inland fault segment oriented perpendicular to the Boumerdes fault that could be an extension of the Bordj-Menaïel fault. The location of the proposed fault segment coincides with an abrupt change in the depth of aftershocks of the Boumerdes earthquake, and Ouyed *et al.* (2011) suggested that this fault segment (F2 in their paper) is a reverse fault.

Because there is no updated seismotectonic map of the region, we could not attribute the Bordj-Menaïel event to a specific fault. The 2014 Bordj-Menaïel earthquake may have reactivated a fault corresponding to the lineament observed at the surface in the epicenter region. As mentioned before, the 2014 event was very close to a lineament observed on a DEM and on aerial photographs. The focal mechanism of Bordj-Menaïel earthquake indicates a strike-slip movement. The fault segment responsible for the earthquake may be linked to the Thénia fault, which shows a similar strike.

The 2014 epicenter is located in a region that showed a positive static coulomb stress change of 0.1–0.2 bar following the 2003 Boumerdes earthquake (Semmane, 2005), suggesting a possible triggering relationship. The moment magnitude of this latest event is M_w 4.1. The maximum intensity was estimated as V on the EMS-98 scale in the epicenter region near the towns of Bordj-Menaïel and Naciria. The earthquake resulted in little damage, likely because it occurred at a depth of 7–11 km.

DATA AND RESOURCES

Seismograms used in this study were recorded by the Algerian Digital Seismic Network (ADSN), which are not yet open to public. However, at Centre de Recherche en Astronomie Astrophysique et Géophysique (CRAAG), we are open for collaborations. Interested persons may contact the center's director A. K. Yelles-Chaouche (a.yelles@craag.dz). Some plots were made using Generic Mapping Tools v.4.5.3 (www.soest.hawaii.edu/gmt; last accessed April 2015; Wessel and Smith, 1998). The spectrograms were calculated using Z. Peng's scripts (http://geophysics.eas.gatech.edu/people/zpeng/Teaching/SAC_Tutorial/; last accessed April 2015).

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We thank all those people who participated in the installation and maintenance of the ADSN network and those who are involved in its management and data quality control. Also, we thank the people who carried out the macroseismic survey despite the very moderate magnitude of the event. The article benefited from useful comments and suggestions from Editor-in-Chief Z. Peng, the Associate Editor, and an anonymous reviewer.

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Annexe 2: The *M*w 5.0 Hammam Melouane Earthquake (North Central Algeria) of 17 July 2013 in the Context of the Tellian Atlas Seismicity

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Dans ce papier, j'ai calculé la source sismique du séisme modéré en inversant les formes d'ondes sismiques.

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The M_w 5.0 Hammam Melouane Earthquake (North Central Algeria) of 17 July 2013 in the Context of the Tellian Atlas Seismicity

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Abstract—On 17 July 2013 (03:00 GMT) a new moderate earthquake of magnitude 5.0 happened in the Tell belts of northern Algeria, more precisely near the village of Hammam Melouane (Algeria), 30 km south of Algiers, the Capital of Algeria. The main shock parameters and aftershocks activity analysis, reveal that the earthquake occurred on a 5 km long dextral strike-slip fault oriented N114°E, a conjugate strike-slip fault of the major NE–SW reverse fault system of the neogene Mitidja basin. This event caused damage to houses and social infrastructures but no fatalities were reported. Onland the earthquake triggered rock falls and minor landslides along the Hammam Melouane river. The occurrence of the Hammam Melouane in the northern limit of the Tell belts is representative of the African-Eurasiatic interplate seismicity of northern Algeria where major seismic events could occurred as the previous Boumerdes event of May 21st, 2003 (M_w 6.8).

1. Introduction

On 17 July 2013 at 03:00:53GMT a moderate earthquake of magnitude M_w 5.0 occurred 30 km south of Algiers, the capital of Algeria. The epicenter was located at 36.4623°N, 3.0426°E, 2.4 km southeast of the Hammam Melouane village which is located at the foot of the northern boundary zone of the Tell belt of northern Algeria (Fig. 1). The Hammam Melouane earthquake is the most important earthquake in the region since the Boumerdes earthquake of May 21st, 2003 (M_w 6.8) (Yelles-Chaouche et al. 2003). These two events occurred on the same fault system extending from Dellys east of Boumerdes city to Mouzaia village west of the Blida city. The two events are separated by just 50 km, which could indicate a stress transfer between them, as suggested by Lin et al. (2011).

Northern Algeria is repeatedly shaked by moderate to strong events. Their occurence may represent the instantaneous deformation of the interplate zone between the two main African and Eurasiatic plates. In the last decade, many moderate earthquakes were recorded in several part of the Tell-Atlas belt of northern Algeria. Among them, we cited the Beni Ilmane event of July, 14th, 2010 (Yelles-Chaouche et al. 2013a; Beldjoudi et al. 2016), the Bejaia sequence of 2013–2014 (Kherroubi and Yelles-Chaouche 2015), the Algiers event of August 1st, 2014 (Khelif et al. 2015), the Merouana event of March, 2014 (Chami et al. 2015). Focal mechanisms of all these events are indicator of the compressional stress regime of the regions affected.

The Hammam Melouane earthquake is considered among the most important events which happened along the northern Tell boundary, indicating that the main cities of the Mitidja basin (Algiers, Tipaza, Blida, and Boumerdes) are under exposure to seismic risk.

2. General Tectonic Context

The Hammam Melouane earthquake represents seismicity of the actual plate boundary limit between the African and Eurasiatic plates. It is known that during the Cenozoic evolution of the Tethys domain, this boundary region evolution was marked by the break off of the ALKAPECA block, the roll back of the Tethysian slab and the suture of the Kabylian blocks on the African margin (Bouillin 1986; Faccenna et al. 2004; Schettino and Turco 2006). From

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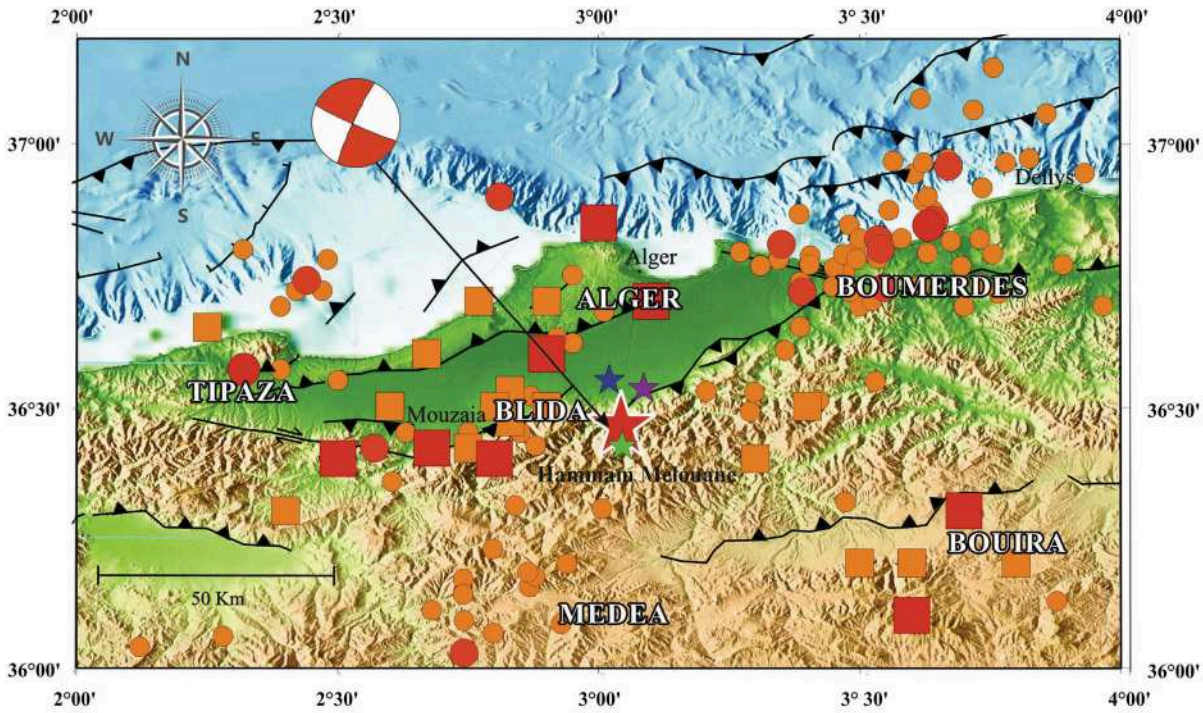


Figure 1
Map of the study area, showing historic seismicity with estimated $M > 4$, from 1365 to the present day [CRAAG; Mokrane et al. (1994)]. *Red squares* correspond to historic earthquakes with intensity IX $\leq I_0 \leq$ X, *brown squares* to historic earthquakes with VII $\leq I_0 \leq$ VIII, *red circles* correspond to instrumental earthquakes with $5 \leq M < 7$, and *brown circles* to instrumental earthquakes with $4 \leq M < 5$. The mainshock is indicated by a *red star* and its focal mechanism (present study). The *green*, *blue* and *magenta stars* represent IGN, EMSC and USGS locations respectively. *Black lines* represent the main active faults (Yelles-Chaouche et al. 2009)

the Pleistocene period, the northern limit of the Tell belt region experienced a tectonic inversion represented by a compressional active tectonics (Strzerzynski et al. 2010) attested by the actual compressive stress field regime and the seismicity (Buforn et al. 2004). The recent Maradja and Spiral projects (Yelles-Chaouche 2010) focusing on the superficial and deep seismic structures of the Algerian margin demonstrate that the compressional deformation led deflection of the oceanic lithosphere of the Algerian basin towards the continent and an incipient subduction zone may initiate close to the bottom slope of the margin (Hamai et al. 2015).

Among the five deep seismic profiles of the Spiral project, the Tipaza profile shows through velocities analysis that the continental crust where the Mitidja lies, thins from 15 km in the upper margin to only 6 km at the margin foot where started the ocean continent transition zone (OCT) (Leprêtre et al. 2013).

Onland, the actual limit is marked by the existence of Neogene basins or internal massifs issued from the break-off of the ALKAPECA domain (Domzig et al. 2006).

Lying on the margin, the Mitidja basin is also subjected to compressive stresses, similar to other Neogene basins in northern Algeria (e.g., Mleta, Chlef, Tizi-Ouzou basins). From these results obtained by the several surveys carried in the region, Yelles-Chaouche et al. (2009) presents a new deformation model of the Algiers region integrating the several main active structures identified all around the Mitidja basin.

This model outlines that the first deformation evidences of the African-Eurasiatic plate convergence touch the foot of the margin which represent northern limit of the Tell belts. Behind this limit, the seismic events are triggered by the several active reverse faults or strike slip fault of the Tell-Atlas belt system.

3. Seismotectonic Context

The Tell-Rif system displays predominantly NE-SW trending folds and reverse faulting defining a right-stepping pattern (Bracene and Frizon De Lamotte 2002).

The Hammam Melouane occurred on the southern border of the Mitidja basin. This later formed after the last post-nappe and extensional phases that affected the Tellian chain (Boudiaf, 1996), It has undergone several compressional phases since the Pleistocene due to convergence of the African and Eurasian plates. Marine surveys of the Algiers margin Maradja (2003, 2005) recently carried out along the margin indicate that compressional deformation starts along the bottom slope of the central margin, as observed in front of the Khayr Eddine bank (Yelles-Chaouche et al. 2009).This could indicate the presence of an incipient subduction zone, thereby explaining the moderate to high rate of seismicity in the region. Among the main events (Table 1), the strong event of January 3rd, 1365 is the first to have been described in the literature (Ambraseys and Vogt 1988). At this epoch, Algiers was a small city called Beni Mezghena. The earthquake produced severe shaking (estimated intensity **X**) and was followed by a small tsunami, suggesting an offshore location. The offshore part of the Mitidja basin has also experienced many other recent events, of which the M 5.7 Ain Benian earthquake of September 4th, 1996 has been well studied (Sebaï 1997).

Onland, along the Algiers coast, the dominant structure is the NE–SW Sahel anticline reverse fault, which is cut by NW–SE conjugate faults (Boudiaf 1996). The total length of the reverse fault, from Tipaza to Algiers is 60 km, and it marks the northern boundary of the Mitidja basin. To the north of the fault (Aymé 1952) reported marine terraces undergoing active uplift of 0.2 mm/year and a shortening rate of 0.4 mm/year (Meghraoui 1991). Several large historic earthquakes have occurred upon this fault, including the great Algiers earthquake of February 1716 (Mokrane et al. 1994). This event has been well documented, as several reports mention with details effects of this event. (Ambraseys and Vogt 1988) suggested an epicentral location near the village of Kolea, 20 km west of Algiers. When Casbah, the old city of Algiers, was reconstructed following the earthquake, the first seismic engineering guidelines measurements were introduced.

The western end of the Mitidja basin is marked by the Mt.Chenoua Paleozoic massif, one of the massifs of the internal domain (Belhai et al. 1990). Along the southern border of this massif, the M_w 6.0 Chenoua earthquake occurred on 29 October 1989 (Meghraoui 1991; Bounif et al. 2003). Farther west, the closure of the basin is marked by a complex fault system where the Oued Djer earthquake occurred on October 31st, 1988. On the opposite side, the eastern part of the Mitidja basin is a complex tectonic zone where the NW–SE striking Thenia fault crosses the NE–SW Boumerdes fault and NNW–SSE Tablat fault (Fig. 1).

The southern margin of the Mitidja basin is marked by the contact between neogene basin deposits and Paleozoic thrust sheets of the Tellian Atlas (Bonneton 1977). This margin is also marked by the major NE–SW-striking Blida fault system, which has a total length of ~ 100 km and is cut by several conjugate faults, similar to those observed in the Bouinan fault system (Guemache et al. 2010).

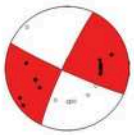
Table 1

Main earthquakes in Hammam Melouane region from 1365 to 2013

Location	Date	Intensity	Magnitude	Casualties
Algiers	03 January 1365	X	–	–
Algiers	– February 1716	X	–	20,000
Blida	02 March 1825	X	–	7000
Mouzaia	02 January 1867	X	–	–
Oued Djer	31 October 1988	VII	$M = 5.4$	–
Chenoua	29 October 1989	VIII	$M_w = 6.0$	22
Ain Benian	04 September 1996	VII	$M_d = 5.7$	–
Boumerdes	21 May 2003	X	$M_w = 6.8$	2300

Table 2

Mainshock locations determined by different seismological observatories and their focal mechanism

Organism	Location	Magnitude	Focal solution				
			Plans	Strike (°)	Dip (°)	Slip (°)	Focal mechanism
CRAAG	36.4705°N 3.0462°E Depth = 9 km	$M_L = 5.1$	–	–	–	–	–
USGS	36.534°N 3.086°E Depth = 9.8 km	$M_B = 4.9$	–	–	–	–	–
EMSC	36.55°N 3.02°E Depth = 10 km	$M_B = 4.9$	–	–	–	–	–
IGN	36.424°N 3.045°E Depth = 3 km	$M_w = 4.9$	Plan1 Plan2	75 178	73 55	144 21	
Present study P_{wave} first motion solution	36.4623°N 3.0426°E Depth = 9.5 km	$M_w = 5.0$	Plan1 Plan2	114.2 204.7	85.7 84.6	174.6 4.3	
Present study *Waveform modelling*	36.4623°N 3.0426°E Depth = 7 km	$M_w = 5.0$	Plan1 Plan2	272 5	56 85	174 34	

Along this southern boundary, the earliest well-documented event was a strong earthquake on March, 2nd 1825, near the city of Blida, south of Algiers, which resulted in ~7000 fatalities (Rothé 1950). This event possibly ruptured the central segment of the Tell-Atlas fault system. Forty years later, on January 2nd 1867 at 7 h 13 min, a new major earthquake was triggered by the western segment, totally destroying the town of Mouzaïa where 48 persons were killed and 100 injured. Ambraseys and Vogt (1988) based on many previous reports, gave details of this event. They indicate that the earthquake affected many villages in the Mitidja area as El Affroun, Chiffa Oued Djer. The main shock was largely felt as far as Djidjelli, 290 km far north east from the epicenter area. Some ground cracks were also observed, as well as large landslides and rockfalls. Aftershocks activity lasted about 6 months. From all the observations, the maximum Intensity evaluated was X with a related magnitude of 6.5.

So we remark that these two historical events have ruptured two main segments of the main Blidean Atlas fault whereas the actual activity is more concentrated in the eastern part.

In the twentieth century, the significant event in the Algiers region was the Boumerdes earthquake of May 21st, 2003. The main fault related to this event extents from the eastern part of the Mitidja basin to the Boumerdes margin. The earthquake also ruptured the eastern part of the Blidean-Atlas Fault, between Khemis el Khechna and Boudouaou (Bounif et al. 2004) as many aftershocks of the mainshock were located in this area.

Finally, it is important to note that since the recent installation of the Algerian Digital Seismic Network (ADSN) (Yelles-Chaouche et al. 2013b), the seismicity of the Algiers region has been well recorded. Recent monitoring indicates that deformation in this region is distributed mainly all around the Mitidja basin.

4. *The Hammam-Melouane Main Shock and its Effects*

The main shock of the Hammam Melouane earthquake was initially located by CRAAG at 36.4705°N, 3.0462°E, near the village of Hammam Melouane, 30 km south of Algiers, at the foot of the

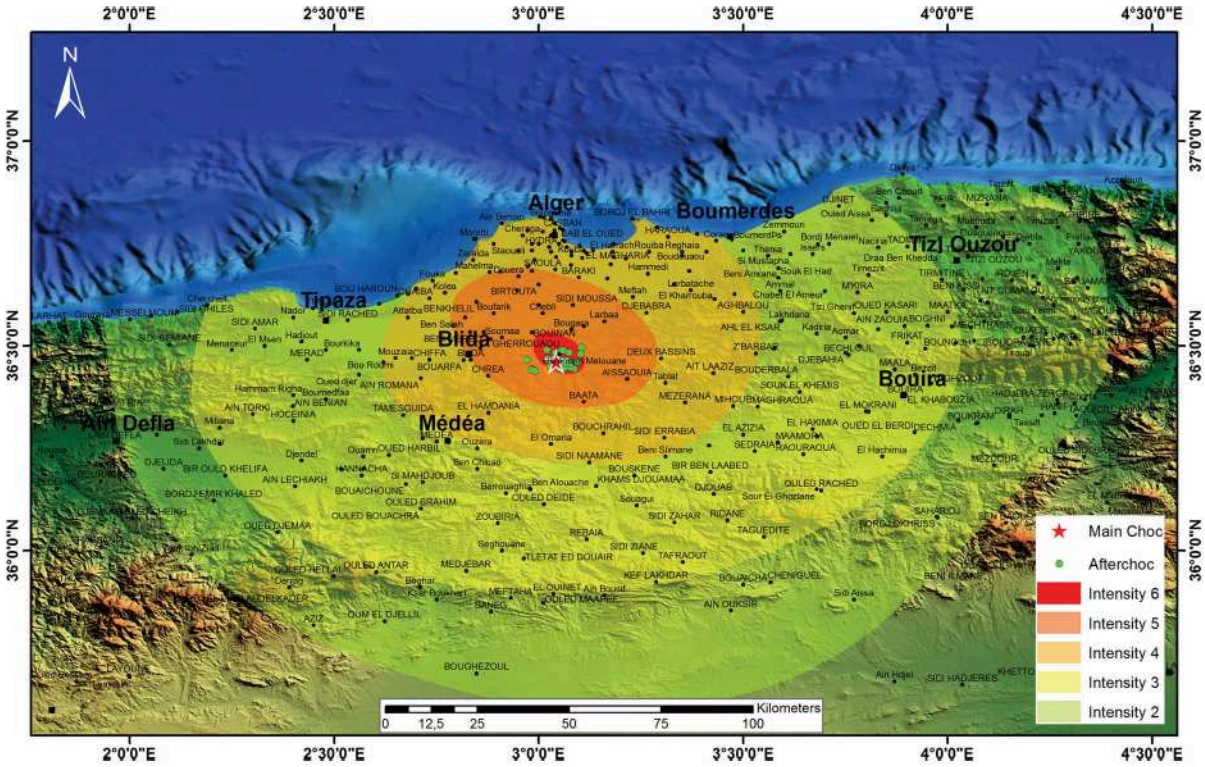


Figure 2
Isoseismal map (EMS-98) for the Hammam Melouane earthquake (CRAAG). The maximum estimated intensity was $I_0 = \text{VI}$. The *red star* represents the epicentral location of the mainshock. *Green circles* correspond to aftershocks

northern flank of the Tell belts. However, relocation with local data suggests a more precise epicenter 2.4 km south of this estimate (36.4623°N, 3.0426°E), at a small village called Magtaa Lazreg. The CRAAG location differs from locations presented by international seismological centers by tens of kilometers (Fig. 1). The seismic moment was estimated at $M_0 = 2.724123 \times 10^{16}$ Nm using the circular model of (Brune 1970), yielding a moment magnitude of M_w 5.0 (Kanamori 1977). The shallow hypocentral depth of 9.5 km is characteristic of crustal seismicity in northern Algeria.

The best-fit focal mechanism, computed from P-wave first motion polarities for 21 vertical-component stations in the ADSN, corresponds to left-lateral strike-slip movement, with nodal planes oriented N114°E and N205°E (Table 2). This is in agreement with CSEM and IGN solutions. In the epicenter location, surface features are displaced by 5 km in a left-lateral sense along a NW–SE orientation (Fig. 1), possibly indicating a buried strike-slip fault connected to the Sidi-Rached Tipaza-strike-slip fault (Boudiaf 1996).

Because the Hammam Melouane earthquake occurred at night, it was strongly felt in the capital, Algiers, and throughout the Mitidja basin; i.e., in the districts of Blida, Algiers, Tipaza, Boumerdes, Medea, Tizi-Ouzou, and Bouira. No fatalities resulted from the event, but injuries occurred during the resulting panic. This event mainly damaged farms and old private houses in the epicentral area; i.e., in several villages located along the Blida massifs (Bougara, Bouinan, and Khemis-El-Khechna). The Housing Ministry reported 1000 buildings seriously damaged by the earthquake. The strongest effects were reported in Magtaa Lazreg, with an estimated EMS-98 Intensity **VI** (Fig. 2). In this village, large cracks formed in the walls of many houses. The post

Figure 3
Photographs of damage resulting from the Hammam Melouane earthquake. **a** Rockfalls along the Hamma Melouane river. **b** The new school of Magtaa Lazreg affected by cracks. **c** Fissures observed in the Post Office of Magtaa Lazreg. **d** Minor landslides in the Magtaa Lazreg Area

office and college (Fig. 3b, c) suffered the most damage and were declared unsafe. Some minor landslides were observed along the slopes of the Hammam Melouane river, mainly in Magtaa Lazreg (Fig. 3d). Significant rockfalls also occurred upon a major outcrop along the road to Hammam Melouane village (Fig. 3a). An isoseismal map, computed from observed impacts in the area, suggests a main NW–SE orientation of the contour-lines, in agreement with one of the nodal plane directions obtained from the focal mechanism.

5. *Waveform Modeling of the Main Shock*

We used the FMNEAR method (Delouis 2014) to calculate seismic source parameters for the Hammam Melouane event. Seven broadband stations (21 total

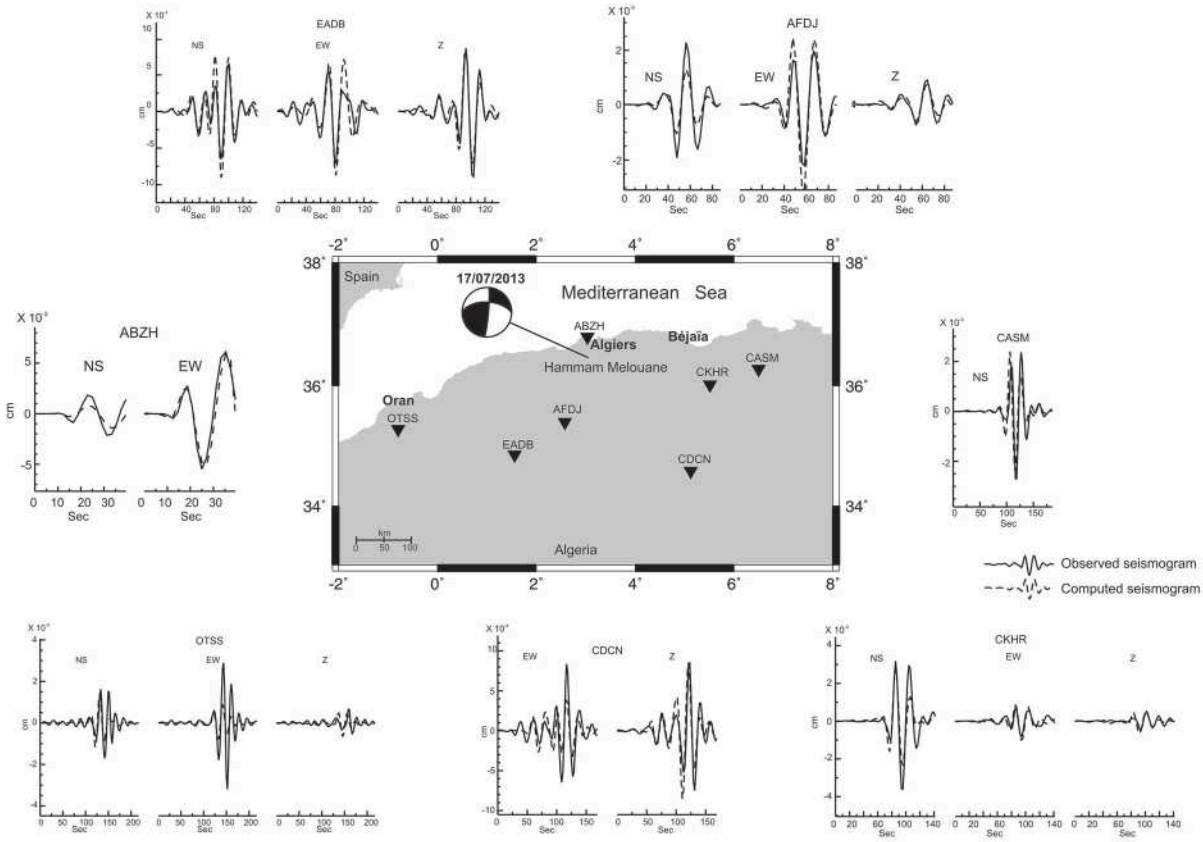


Figure 4
Focal mechanism of the 17 July 2013 mainshock, obtained using waveform modeling with the FMNEAR method of Delouis (2014). Observed and computed displacement seismograms are plotted with *continuous* and *dashed black lines*, respectively. Seismograms are band pass filtered between 0.03 and 0.06 Hz. The *star* indicates the epicenter. *Inverted triangles* represent broadband stations used in the inversion

components) were integrated into the inversion procedure. Seventeen components were selected by the inversion to estimate the focal mechanism and source parameters (Fig. 4); the other four were discarded automatically by the algorithm. The best solution has an RMS of 0.511 and a confidence index of 80%. The optimal depth is 7 km (Fig. 5). The first plane of the double-couple (strike 5°, dip 85°, rake 34°) describes N-S left-lateral movement; the second (strike 272°, dip 56°, rake 174°) describes E-W right-lateral movement. The estimated seismic moment is $M_0 = 3.54 \times 10^{16}$ Nm and the corresponding moment magnitude is M_w 4.97.

The first fault plane, oriented E-W with a dip of 56°, is vertical on the focal mechanism obtained from P-wave first motion polarities. The second plane (N-S) is similar to that calculated using P-wave first motion polarities. The two focal mechanisms obtained using different methods are very similar to each other.

6. *Data acquisition and Processing of Aftershocks Sequence*

Seismic activity in the Algiers region is monitored by several permanent stations belonging to the ADSN (Yelles-Chaouche et al. 2013b). In addition, a few hours after the main shock, a CRAAG team moved to the epicentral area to reinforce the seismological network, and to investigate the earthquake's effects. Seven portable Geostar digitizers with L22-D passive geophone sensors ($f_c = 2$ Hz) were deployed in the epicentral area. One of them was near the relocated

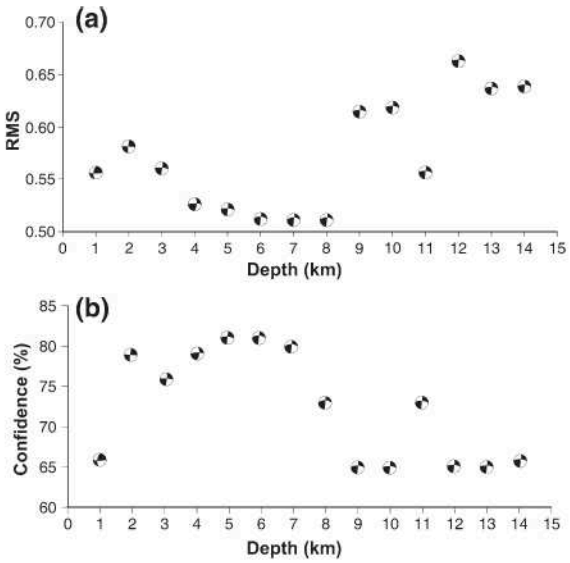


Figure 5
Solution quality as a function of depth. **a** Root mean square (RMS) versus depth (km). **b** Confidence Index (CI) versus depth. The best solution (CI = 80%, RMS = 0.511) corresponds to a depth of 7 km

epicenter at Magtaa Lazreg village (Fig. 6). The southernmost station in the temporary network was the most difficult to install because of rough topography, poor security, and dense vegetation in the Blidean Atlas. During the lifetime of the temporary network, some portable stations were out of service for few days due to technical problems.

The mainshock triggered a significant aftershock sequence that continued for nearly 3 months (until October 2013). A total of 412 events were recorded. Using the duration magnitude (M_d) scale of (Ouyed et al. 2011) for the Boumerdes aftershock sequence, aftershock magnitudes were no greater than M_d 3.1. Nevertheless, events were still recorded frequently in the area up to a year later.

Aftershocks were initially located with the Hypoinverse software package (Klein 1978).To improve relative locations, this was followed by detailed analysis using HypoDD (Waldhauser and Ellsworth 2000). A velocity model was created from one of the deep seismic profiles made during the Spiral program, off Tipaza (Fig. 7a) (Leprêtre et al. 2013) and is summarized in Table (Fig. 7b). Only events with a minimum of five phase picks (P and S) were used; this criterion allowed us to select 99 reliable events. To further improve location accuracy, we performed several tests by varying the V_P/V_S ratio; the best results were obtained for $V_P/V_S = 1.66$. After imposing the additional constraints of an average residual (RMS) <0.20, an azimuthal gap (GAP) <180°, and error on the horizontal (ERH) <2 km and vertical (ERZ) <2 km, only 80 events of the 99 agree with the previous criteria.

Application of the hypoDD program, using conjugate gradient method for least square solutions improves more significantly the aftershocks location. Indeed, relocation errors estimation is as follows: the East–West error (Ex) is between 4.9 and 50.4 m (average = 14.9 m), the North–South error (Ey) is between 5.7 and 59.8 m (average = 16.4 m) and the depth error (Ez) is between 7.7 and 36.5 m (average = 15.5 m).

7. *Analysis of the aftershocks sequence*

The relocated aftershock sequence (Fig. 8a) displays a coherent event cloud mainly oriented NW–SE. This is in agreement with the N114°E fault plane solution obtained from the focal mechanism of the main shock. The total length of the rupture reached about 4 km, in agreement with the calculated moment magnitude. From cross-section A–B in Fig. 8c (azimuth N25°E), the rupture occurred fault plan of 3 km width.

From Fig. 8b, it is evident that the rupture did not reach the surface. Indeed, aftershocks were mainly concentrated at depths of 7–10 km. This shallow seismicity, mainly concentrated in the epicentral area of Hammam Melouane, represents active deformation of the upper part of the crust. It happens where the Mohorovic discontinuity of the continental crust thins and nearby the zone where the contact between the two oceanic and continental lithosphere occurred (Hamai et al. 2015).

In other hand, the two cross-sections in Fig. 8c, d strongly suggest that foci are distributed on a near-vertical plane, consistent with the dip of the main shock focal mechanism (85.7°SW). On these figures, we can see that the rupture deepens to the north from 5 km in the SE tip of the fault to 7 km on its NW extremity. We did not see any aftershock activity extending inside the Mitidja basin.

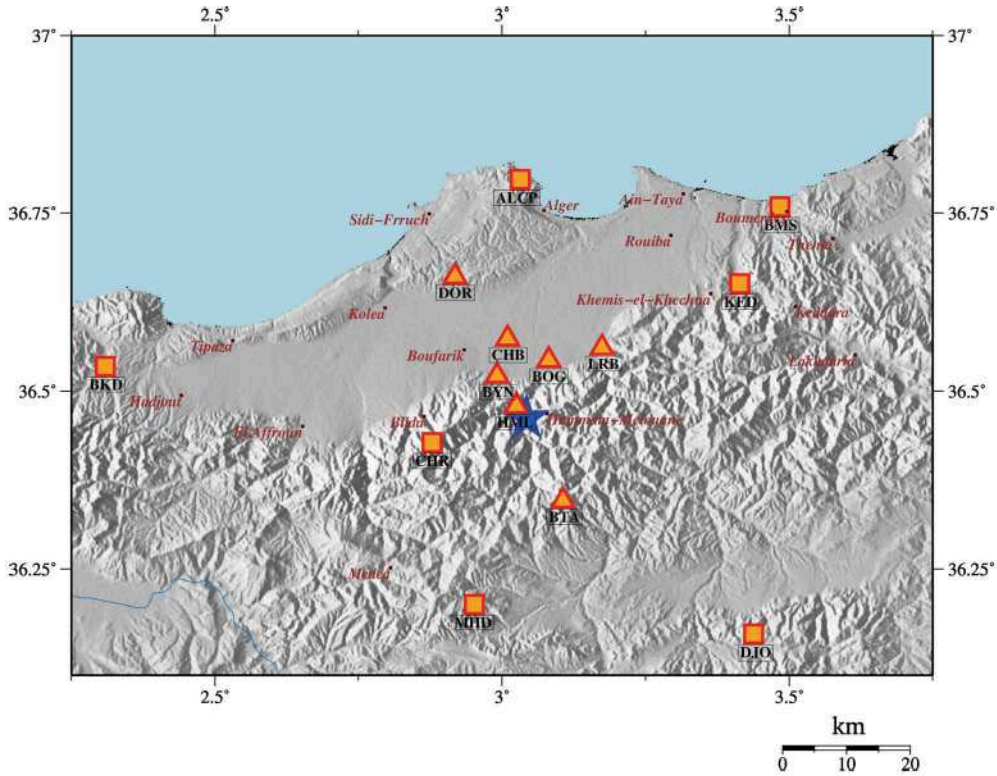
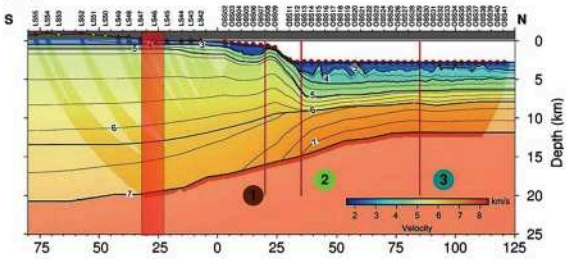


Figure 6
Permanent and temporary seismic stations that monitored the seismic sequence of this study. *Squares* correspond to stations of the permanent Algerian Digital Seismic Network (ADSN). *Triangles* correspond to portable stations deployed in the epicentral area. The *blue star* represents the mainshock of the Hammam Melouane earthquake

(a)



(b)

Thickness (km)	P-wave velocity (km/s)
1.0	3.0
1.5	4.0
12.5	5.0
7.0	6.0
Half space	7.0

Figure 7
a Deep seismic profile along the Tipaza (Spiral project, Leprêtre et al. 2013), the *red rectangle* represent the region of the profile corresponding to southern border of the Mitidja basin. **b** The velocity model used for this study

8. Focal Mechanisms of Large Aftershocks

Focal mechanisms of the largest 11 aftershocks were computed using the SPHERA code (Rivera and Cisternas 1990), based on P-wave first motion polarities detected on vertical-component sensors. To ensure accuracy and reliability, we only computed focal mechanisms for events recorded by at least eight stations. The focal solutions labeled 1–11 and their spatial distribution are shown in (Fig. 9). Compressive quadrants are colored in red.

We observe a predominance of strike-slip movement in Fig. 9, in agreement with the focal mechanism of the mainshock. From these mechanisms, we deduce that the main fault plane is not linear, but comprises several minor segments, each with an orientation slightly different from that of the main fault plane. Some mechanisms have a more (or less) pronounced normal component (events 1–6, 9,

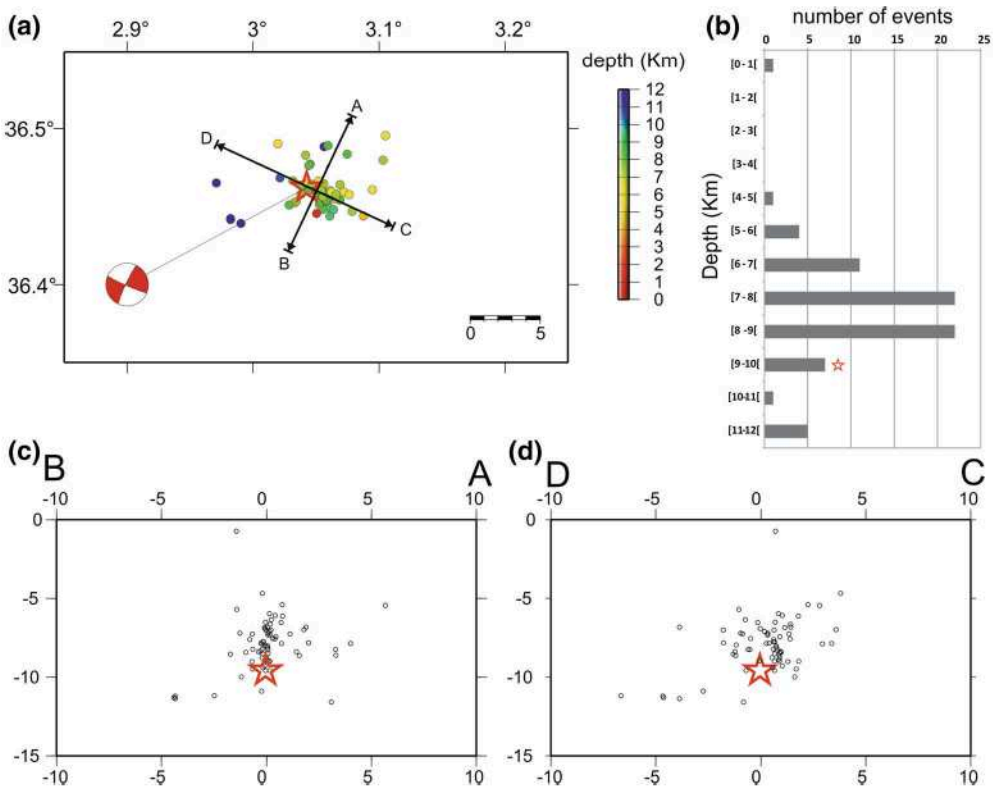


Figure 8
Horizontal and *vertical* distributions of aftershocks

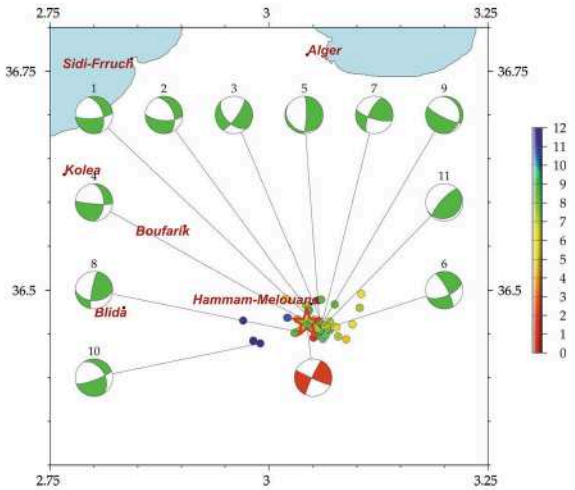


Figure 9
Horizontal distribution of relocated events, showing the focal mechanisms of large aftershocks

and 10) or a reverse component (events 7 and 8). These latter events could express transtensional or transpressional movement along a segment of the fault, in relation to the compressional stress regime. The reverse faulting mechanism of event 11 could indicate movement on a segment related to adjacent major reverse faults in the region.

In general, focal mechanisms obtained from this study are consistent with strike-slip faulting that cuts the main reverse fault systems at the foot of the Tell belt mountains.

9. *Discussion and Conclusions*

The Hammam Melouane earthquake of 17 July 2013 is among the many moderate to strong events that periodically strike northern part of the Tell Atlas belt of

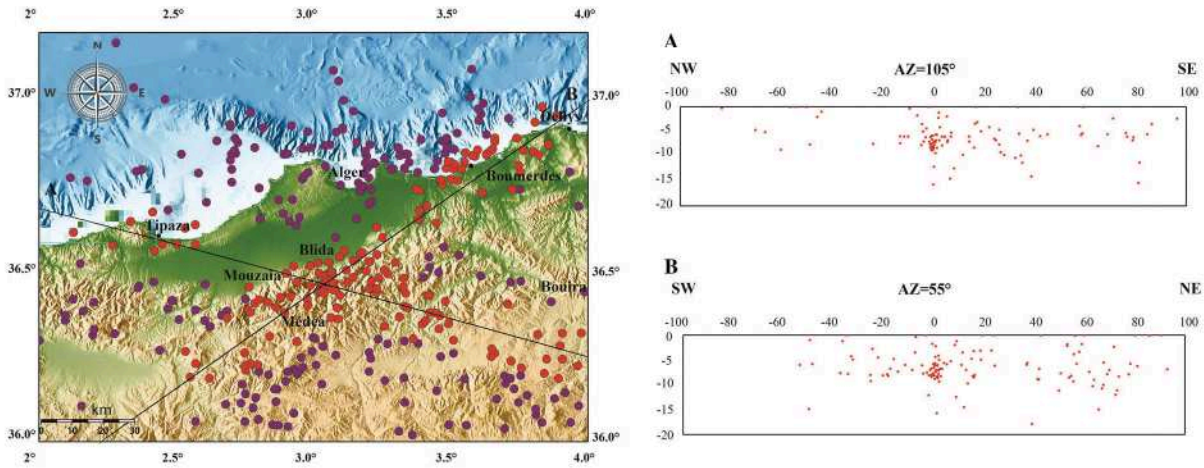


Figure 10
Seismic events map recorded from 1998 to 2015 with magnitude $M > 3$ of the Mitidja basin (CRAAG), *A* NW SE and *B* NE SW corss-sections, the *red circle* plotted on the map correspond the events plotted in the cross-sections

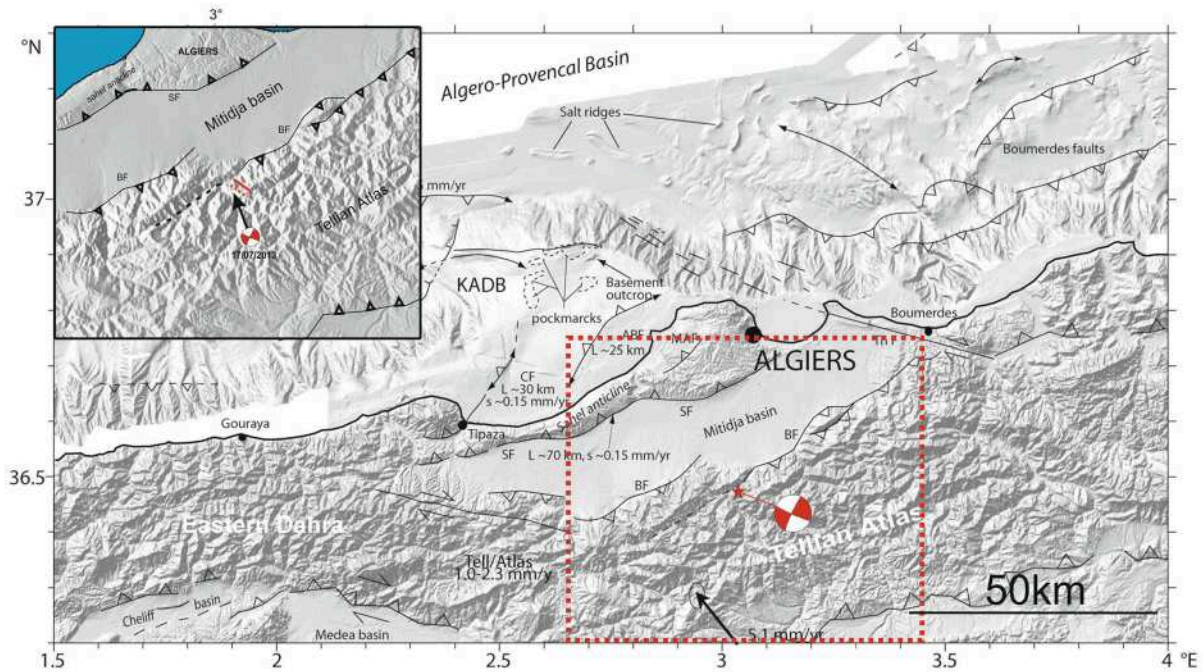


Figure 11
General seismotectonic framework of the Algiers region (Yelles-Chaouche et al. 2009). The location of the Hammam Melouane mainshock is represented by a *red star* and the moment tensor solution calculated in the present study. *Inset* location of the fault segment, as inferred from the aftershock distribution and fault plane solutions. The *red dashed line* corresponds to the surface projection of the fault segment. The *solid red line* indicates a continuous fault at the surface

northern Algeria as the several earthquakes recorded since the El Asnam event of October 10th, 1980 in the Chleff basin. This shallow seismicity characterizes the interplate zone between the two major convergent African-Eurasiatic plates. In this limit the ocean lithosphere of the Algerian basin hit the continental

lithosphere of the African margin where the graduate uplift Tell belts and closure of the several Neogene basins traduce deformation of this collision.

We can assume that the Hammam Melouane event also occurred 10 years after the Boumerdes earthquake of May 2003, which ruptured the margin of the Boumerdes area and the eastern part of the Mitidja basin.

The Hammam Melouane event occurred along more precisely the main fault of the southern boundary of the Mitidja basin, one of the Neogene basin of the Tell belt. This main active fault system of about 100 km appears as an echelon fault system represented by NE–SW oriented segments, as the Menaceur-SidiYahia fault in the west; the Hadjout-Meurad, Oued Djer-Lalla Aicha, and OuledYaich-Bouinan faults in the center; and the Bougara-Larbaa-Khemis El Khechna fault in the east (Boudiaf 1996; Guemache et al. 2010; Meghraoui 1988) Many NW–SE strike-slip faults crosscut this fault system. Many historical seismic events have been also reported along this fault system as the strong event of 1825. Since the recent installation of the ADSN, a number of events have been located in the fault system, evidenced more active tectonics of this region.

A NW–SE and NE-SW cross sections of the seismicity recorded from 1998 to 2015 along the Mitidja basin (Fig. 10) do not evidence any extent of

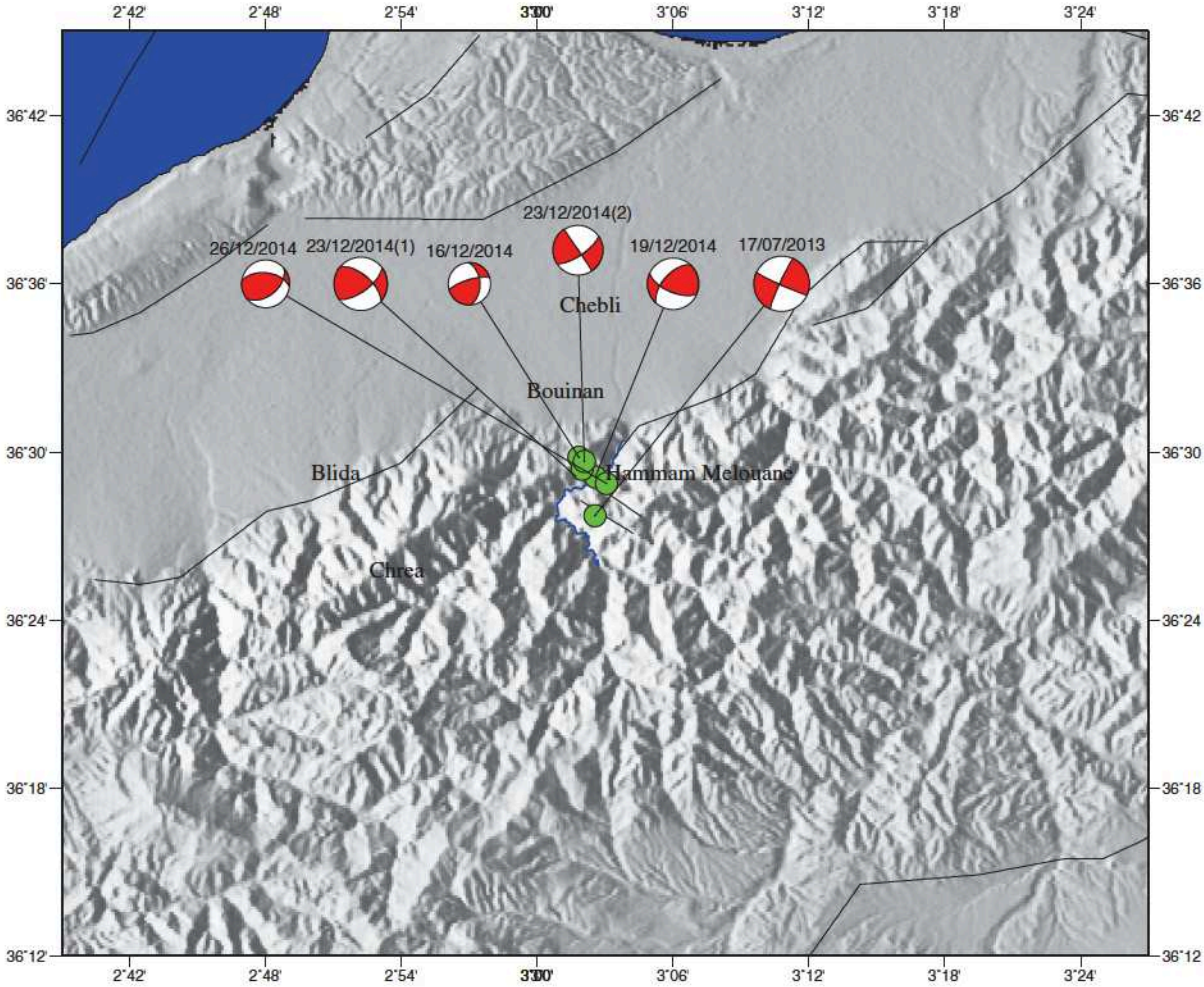


Figure 12
Location and focal solution of the Hammam Melouane earthquake of July, 17th, 2013, and the more recent events in the area

the Hammam Melouane fault towards the inner part of the basin. Indeed, this figure displays two clouds concentrated near the Blida fault in the south and the Sahel fault in the north.

After the Boumerdes earthquake of May 2003, a stress transfer occurred between several faults of the northern flank of the Tell belts, involving the Boumerdes fault offshore, the strike slip NW–SE Thenia Fault and the eastern segments of the southern Mitidja Fault. The Hammam Melouane earthquake did not rupture the main blind reverse fault (from Boudouaou to Larbaa), as predicted by (Lin et al. 2011) but involve only a small NW–SE strike slip around the Bouinan village. In this area, segmentation of the main Blidean fault system is evidenced by a topographic shift of 15 km (Fig. 11). This shift could represent a surface expression of a major NW–SE strike-slip fault cross-cutting the Mitidja basin as described by Bonneton and Truillet (1979), from the Sahel structure in the north to the Tell Atlas Fault in the south.

The Hammam Melouane region has experienced intense seismic activity during the past 6 months, which is ongoing (December 2014 to June 2015). We are still studying the new events to understand the link between the two periods of intense seismicity. From Fig. 12, it appears that the rupture jumped to an adjacent fault segment. However, it is difficult to draw conclusions about a possible relationship between this new seismic crisis and the Hammam Melouane event. More investigations are required to understand the complete seismic structure of the Hammam Melouane region.

If we assume that the seismic activity over the last two centuries is interrelated, then we can infer that the events on the northern limit of the Tell belts correspond to an incremental rupture process of this limit, where different segments of the fault system ruptured step by step at different times. Each segment could rupture a 5–15 km length of the main fault, which corresponds to a moderate to strong earthquake (up to M_w 7.0). However, further investigations are required to understand the maximum possible rupture length, the recurrence interval, and the stress transfer, before we can fully characterize the seismic hazard of one of the most important economic regions of northern Algeria.

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Annexe 3: The earthquake swarm of December 2007 in the Mila region of northeastern Algeria

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Ma contribution dans ce papier était le calcul tu tenseur de contraintes.

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ORIGINAL PAPER

The earthquake swarm of December 2007 in the Mila region of northeastern Algeria

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Abstract In December 2007, the Mila region of northeastern Algeria experienced thousands of microearthquakes ($0.8 \leq Md \leq 3.9$) recorded by eight temporary stations, in addition to permanent stations. Most of the events were too small to be located precisely, but a set of 122 precisely located events shows an alignment of epicenters, extending mainly in a horizontal band at about 1–2 km depth in a NNW–SSE direction and concentrated in a small area, 3 km southeast of Jebel Akhal, a small rocky hill between the Beni Haroun dam/reservoir and the Oued Athmania reservoir. The reservoirs are connected by pipelines, and a pumping station ensures water can be transferred between them at transient pressures of up to 80 bars. During the pumping in 2007, only 45 % of the transferred water ($\sim 600{,}000$ m^3 per day) was recovered at the Oued Athmania reservoir, and a large amount of the slightly pressurized water leaked through defective joints in a tunnel that passes through the Jebel. This water penetrated deeply into the soil with the assistance of preexisting fractures, faults, and karsts. Nine days after the first pumping started, a local increase in pore fluid pressures at shallow depths triggered seismicity southeast of Jebel Akhal, where the faults were probably close to failure. The focal mechanisms show a near vertical N–S strike-slip fault plane under regional NW–SE tectonic compression. One of the fault plane solutions is consistent with the NNW–SSE direction along which the seismic events are aligned. Furthermore, a long-term comparison of the seismic activity in the region versus water levels behind the dam and the pumping of water shows that the earthquake swarm was a one-off event related to the pumping operation.

Keywords Triggered seismicity · Earthquake swarm · Water transfer · Beni Haroun dam · Mila, northeastern Algeria · Pore fluid pressures

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1 Introduction

It is well known that earthquakes can be triggered not only by tectonic forces, volcanoes, and tidal forces, but also by human activities such as dam construction (Carder 1945, 1970), high-rise building (Lin 2005), and the production of oil and gas (McGarr 1991). Indeed, hydraulic fracturing is used to fracture tight geological formations and release natural gas trapped within them by injecting water deep into the ground at high pressures, and that process can cause small earthquakes. In various case studies, some authors have proposed that the injected fluids trigger earthquakes after a delay and that the delay in aftershocks can be explained by fluid flow (Nur and Booker 1972). In addition, the effect of fluids on fault strength is evident in the earthquake swarms that follow fluid injections into wells, particularly when the crust is in a critical state of stress. We can cite the examples of the Matsushiro experiment in Japan in 1970 (Ohtake 1974), the KTB project in Germany (Zoback and Harjes 1997), and the Nojima fault zone experiment in Japan in 1997 (Tadokoro et al. 2000).

Furthermore, the induced or triggered seismicity that follows a change in pore fluid pressure can be caused by seasonal groundwater recharge (Saar and Manga 2003) or by heavy rainfall (Jiménez and Garcia–Fernandez 2000; Husen et al. 2007). For example, Hainzl et al. (2006) demonstrated a statistically causal relationship between rainfall and earthquake activity in the Mt Hochstaufen region, and Roth et al. (1992) and Kraft et al. (2006), among others, have found evidence for earthquake activity induced or triggered by pore fluid pressure changes due to intense precipitation. The evidence is clear, therefore, that pore pressure changes, induced by intense precipitation, are able to trigger earthquake activity, even at some kilometers depth, through the mechanism of fluid diffusion.

In northern Algeria, in December 2007, the population of the two small villages of Ain Tinn and Sidi Khelifa (Fig. 1), located about 30 km west of Constantine (the main city in northeastern Algeria), experienced an unusual seismic crisis over a period of about 1 month, a few months after the activation of the nearby Beni Haroun hydraulic complex, which consists of a main dam and reservoir, a secondary reservoir, and a system for transferring water between the two reservoirs. The seismic activity was marked by dozens of felt earthquakes ($Md \leq 3.9$), and although it did not damage any building or infrastructure, it did frighten the population unaccustomed to such phenomena. In addition to the shaking, loud explosion-sounding noises preceded the earthquakes.

It should be noted that the seismic crisis occurred near the secondary Oued Athmania reservoir, located about 15 km to the south of the main dam. The two reservoirs are connected by a system of pipelines in order to provide water to the nearby districts (Fig. 1). The pipe system crosses a rocky mountain 1256 m high (Jebel Akhal) through a tunnel about 6 km long. A pumping station ensures the transfer of water (up to 600,000 m^3 per day) between the two reservoirs. The pumping commenced on November 21 and was stopped on December 11 during the earthquake swarm that started on November 30. The peak of recorded seismicity was reached on December 13. A long-term comparison of the seismic activity in the region versus water levels in the dam and pumping rates showed that the increase in seismic activity was a one-off event related to the pumping operation. Field investigations carried out with the engineers of the Hydraulic Service of the ANBT (National Agency of Dams and Transfers) led us to conclude that the seismic activity in this area was triggered by an unintentional process of water injection. The following two findings support our conclusion: (1) during the pumping operation, it was observed that only 45 % of the pumped water was recovered at the Oued Athmania reservoir; (2) complementary investigations made by the company charged with the construction of the

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tunnel through Jebel Akhal have found defective joints between the prefabricated segments of the lined tunnel. A large amount of leaked water penetrated the soil deeply with the assistance of a large number of fractures, faults, and even karsts.

Because anthropogenic activities can have an impact on tectonic stresses and pore pressures, any seismicity that occurs in spatial or temporal proximity to such activities is usually thought to be triggered or induced. For a long time, the two adjectives "triggered or induced" were used interchangeably, and McGarr and Simpson (1997), and more recently Gupta (2011), have discussed this question well. In our case, we will use the adjective "triggered," because the earthquake swarm occurred only a few days after the start of the causative activity, which suggests that the existing faults were critically stressed.

In this paper, we present the results of our study of this earthquake swarm that occurred in northeast Algeria during the transfer of water between the main Beni Haroun dam/reservoir and the secondary reservoir of Oued Athmania. A long-term comparison of seismic activity versus water levels in the dam shows that the peak of seismicity occurred during this period of water transfer. Almost all the located earthquakes are clustered 3–4 km SE of Jebel Akhal, a rocky mountain between the two reservoirs (Fig. 1), and which is a well-known karst region near the Mila basin. A triggering of seismicity is suspected to have taken place in this region following the unintentional injection of water from the tunnel into the ground. The main objective of this contribution is to determine the causes of the earthquake swarm by investigating the spatial and temporal distribution of earthquakes and their fault plane solutions. To this end, we processed the records of more than one hundred selected low-magnitude earthquakes. The records were acquired during the period November 30 2007 to January 19, 2008.

This study is interesting for several reasons: on the one hand, this is the first time that the region of Mila has experienced an earthquake swarm of this scale, and on the other hand, the seismicity is located in the vicinity of the largest dam in Algeria.

Among the main findings of this study is that the crust in some parts of the Mila region (Tellian Chain) could be so close to failure that even pressure variations as small as that produced by an unintentional injection of slightly pressured water (leak in the tunnel) can trigger earthquakes at a few kilometers depth. Hainzl et al. (2006), assuming homogeneous conditions in their analysis, found that a stress change in the order of a millibar is able to trigger the earthquakes.

In addition to the natural moderate level of seismicity in the region, triggered seismicity might increase the risk of the landslides that characterize the region of Mila.

2 The geological context of the Beni Haroun hydraulic complex

The Beni Haroun hydraulic complex was built on the Rhummel River that flows mainly from Constantine to the Mediterranean Sea. The complex consists of a main dam and reservoir, a secondary reservoir, and a water transfer system (Fig. 1).

The dam is 120 m in height, and can contain about 1 billion m^3 of water. It was built in order to provide water to several cities in the eastern region of Algeria. A transfer system made of pipelines was added to the dam to connect it with the Oued Athmania reservoir (35 million m^3 capacity) about 15 km to the south. A tunnel was dug in the Jebel Akhal, a rocky hill, in order to reach the Oued Athmania reservoir. The length of the tunnel is 5.9 km. The diameter of the pipe varies from 1.4 m at the pumping station to 3.6 m in Jebel Akhal. Near Jebel Kernenou (east of the city of Mila), the pipe diameter is 2.4 m. The water pumping station uses 180 megawatts of power and has a capacity of 23 m^3/s. The

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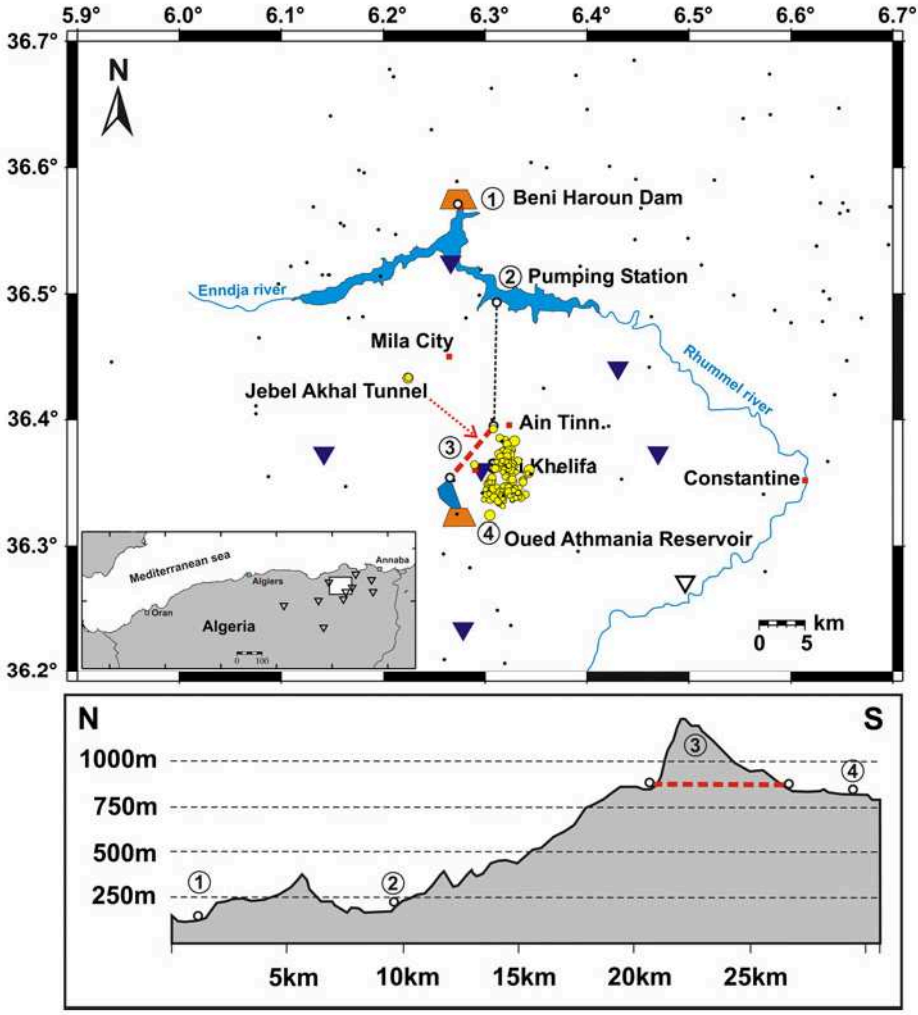

Fig. 1 Introductory map of the study area. *Top* Water conveyance system (*dashed lines*) from the pumping station at Beni Haroun dam to the Oued Athmania storage dam. The *red dashed line* shows the tunnel passing through Jebel Akhal. *Black dots* represent 2 years of seismicity (2006–2007). The reverse triangles show the temporary network. The *yellow full circles* show the earthquake swarm discussed in this study. Also shown in the inset are outlines of the dam and reservoir, the main rivers supplying the dam, and the permanent seismic recording network (*open triangles*). Bottom: topographic profile (N–S) of the water conveyance system

amount of water transferred daily can be as much as 600,000 m^3. The transfer system crosses several geological units (Figs. 1, 2).

The dam and its transfer system were built on the Mio–Pliocene basin of Constantine. This basin is characterized by post-nappe deposits, and it has been slightly affected by recent tectonic deformation. During the Quaternary, several series of lacustrine continental deposits (Villafranchian) were laid down in the basin.

To reach the Oued Athmania reservoir, the pipes of the transfer system cross several geological domains from north to south. Starting near the Beni Haroun dam, they first cross

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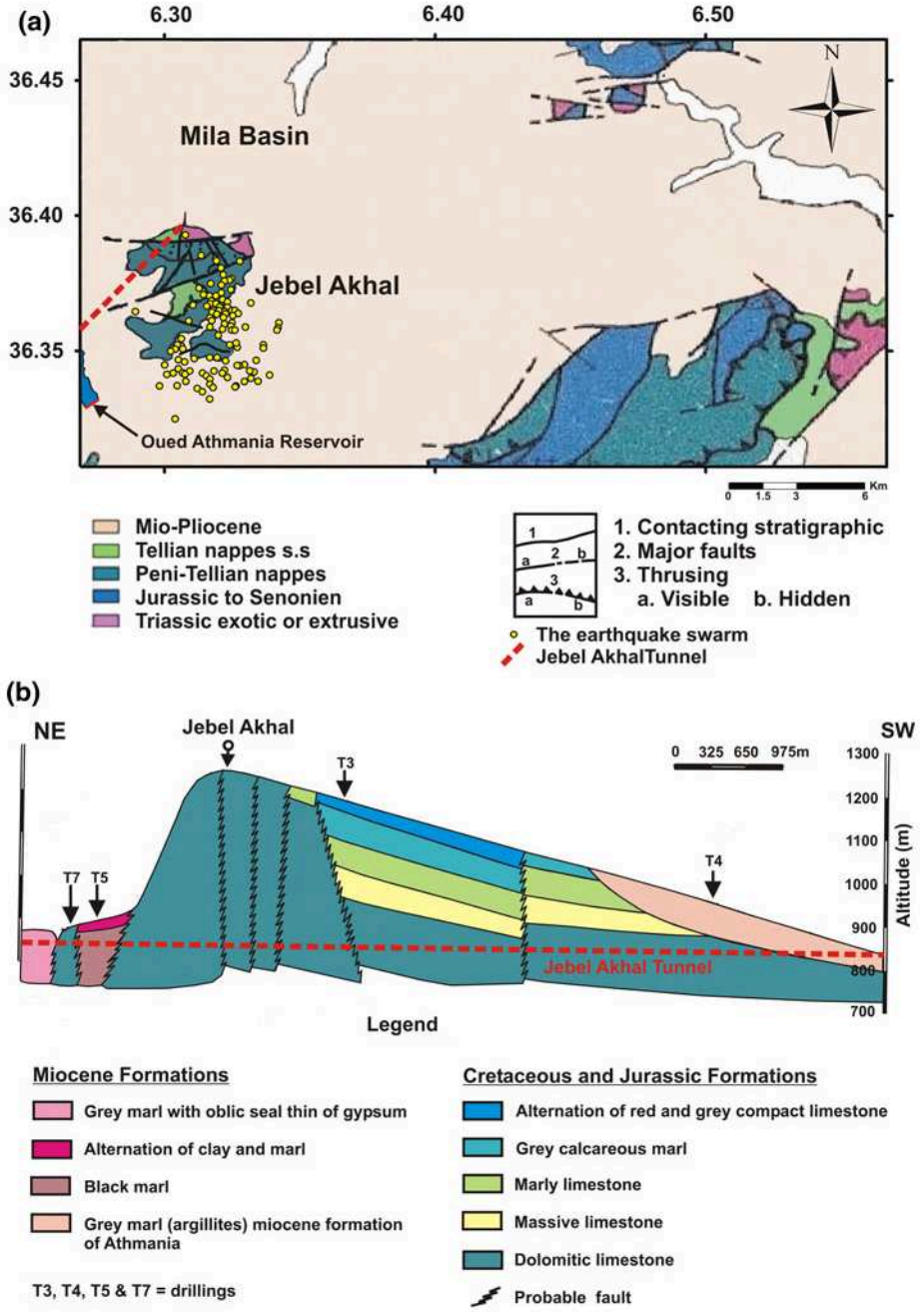

Fig. 2 *Top* Geology of the region of interest (modified after Vila 1977). Also shown are the preliminary located epicenters (*yellow dots*), the tunnel as a *red dashed line*, and part of the Oued Athmania reservoir. *Bottom* Schematic tectonic profile of Jebel Akhal (taken from the ANBT report, the National Agency of Dams and Transfers)

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the geological units of the Mila basin, and then the hills of Jebel Kernenou, which are characterized by an E–W structure made of gray marls. From there, the system crosses the Ain Tinn basin, a depression filled by Miocene gray marls, and then the Jebel Akhal, a 1,256-m hill that is aligned E–W, and which is made up of thrust layers of Jurassic and Cretaceous sandstones, dolomites, and marls (Fig. 2). This massif, considered to be para-autochthonous, belongs to the sandstone units of the Tellian nappes (Vila 1980). It is bound to the north and south by several subvertical faults oriented mainly E–W, and it is crosscut by several fractures associated with strike-slip movements (Fig. 2). From the hydrogeological point of view, this massif collects and stores rainwater, as attested by the many springs around its periphery. The pipe system terminates in the Oued Athmania reservoir, which is located on the sandstones and calcareous marls of the Mila basin.

From the seismotectonic point of view, the Mila basin is one of the Neogene basins of the Tellian Atlas. It is marked by moderate seismic activity, mainly concentrated along active structures many kilometres long that trend NE–SW, and which are like the main faults in the system of the Tellian region (Yelles et al. 2006). In addition, since the installation of the Beni Haroun hydraulic complex, there has been no particular increase in seismic activity in the region, except for the earthquake swarm of December 2007.

3 Seismic networks and data

To analyze this earthquake swarm, eight temporary three-component digital stations were deployed in the region. This temporary seismic network was designed in the first place with an interstation spacing of between 15 and 25 km (full inverse triangles in Fig. 1). A few days later, the geometry of the network was slightly modified for the purposes of reducing the station spacing and enhancing the quality of the records by moving away from sites with a high noise level.

The Sidi Khelifa station (SP04) was added to improve the precision of locating the seismic events in the epicentral area, and it was done after we had identified the position of that area from the first recorded events.

Whenever possible, we also used phase readings from the broadband and one-component short-period waveforms recorded by the 10 nearest stations of the ADSN (Algerian Digital Seismic Network), marked by the open triangles in Fig. 1.

The eighteen stations used in this study employ two types of equipment: the first type consists of a Geodevice EDAS-24IP digitizer and two different sensors, a 1 Hz Geodevice FSS-3 and a BBVS-60 s (Chinese equipment); the second type consists of a Q330 digitizer with a 1 Hz Kinemetrics SS-1 sensor.

The temporary stations, equipped with solar panels, recorded at a sampling rate of 100 Hz. Time was controlled by GPS to ensure a 2 ms accuracy.

The recording periods at each site are summarized in Fig. 3. This figure also shows that the maximum number of pickings is obtained at the nearest station (SP04).

During this earthquake swarm, more than 7200 microearthquakes were recorded. Most of the events were too small to be recorded by more than two or three stations of the network, and could not be precisely located. Furthermore, we decided that events recorded by less than five stations should be rejected, because of the difficulty in determining a sufficiently precise location of these epicenters. For these reasons, we took into account only 231 of the seismic events after data analysis.

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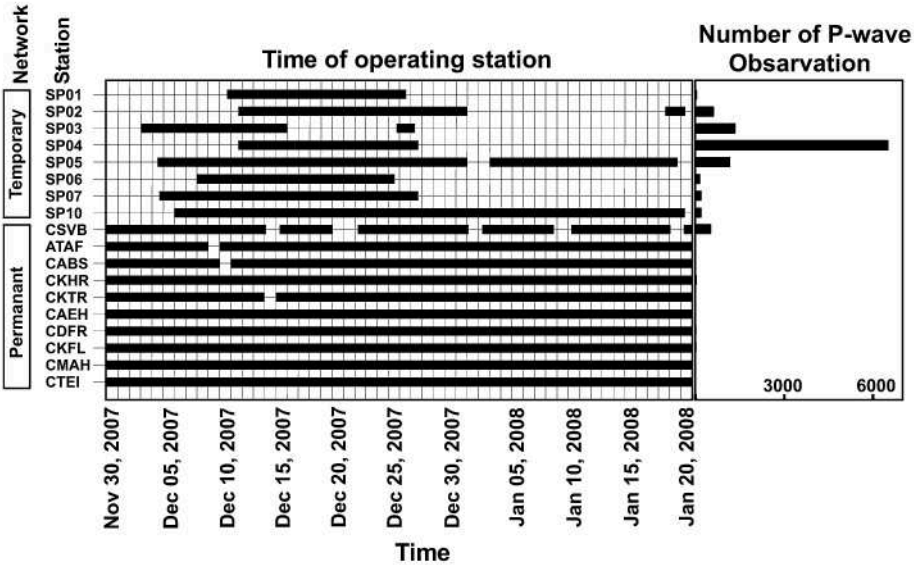

Fig. 3 Station operation time schedule. The number of *P* phase pickings for each station is shown at top right

4 Methods and results

4.1 Velocity model and earthquake distribution

In order to build a reliable minimum 1-D P-wave velocity model, we used the number of phases to select the most reliable events. The set of 231 events, preselected as described in Sect. 3, was used as input to the earthquake location program HYPO2000 (Klein 2002). We have tried two regional and one local velocity models. One of the regional velocity models was suggested by Doser et al. (1997), and the other was obtained from deep seismic sounding data in Tunisia (Research Group for Lithospheric Structure in Tunisia 1992), since the results of the SPIRAL project carried out in northern Algeria are not yet available (Yelles et al. 2010). The local velocity model was obtained by using a trial and error approach with the first week of data recorded in Mila (Abacha 2009). This model was later chosen as a preliminary velocity model as it gives better results. We then formed a subset by selecting 115 events with an azimuthal gap of less than 180°, a horizontal error (ERH) less than 1 km, and a vertical error (ERZ) less than 2 km. A 1-D P-wave velocity model was estimated by applying the iterative 1-D inversion algorithm VELEST (Kissling et al. 1994). The code was run for simultaneous inversions of both velocity and hypocenter locations using the outputs of the HYPO2000 algorithm as a starting model, and by splitting each layer of the starting model into several layers. Fifteen inversion steps, each with nine iterations, were applied to the initial velocity model. The inversions were applied until the velocity model and the error location stabilized, which happened after the thirteenth iteration.

The earthquake relocation process was repeated using the HYPO2000 algorithm and the preferred 1-D P-wave velocity model. This increased our data set from 115 to 129 events, with the same location errors as those mentioned above. Moreover, this data set allowed us to calculate the *Vp*/*Vs* ratio using the Wadati diagram.

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Event magnitudes were estimated from signal duration (T) and the epicentral distance (D) using the formulae $Md = -1.9 + 2.8 \log(T)$ for $D < 40$ km, and $Md = -1.2 + 2.2 \log(T) + 0.0033\, D$, for 40 km $\leq D \leq$ 200 km (CRAAG, unpublished work 1981).

Furthermore, to get a sharper and more detailed image of the clusters, we relocated the 129 events in the data set by applying the Hypo-DD double-difference algorithm (Waldhauser and Ellsworth 2000). We performed a number of iterations to optimize the damping and control parameters of the program, and to obtain good conditioning and convergence of the solution. About 95 % of the events were relocated (122 events).

Table 1 shows the final and preferred velocity model that gives the lowest RMS residual of 0.2 s. P-wave velocities increase from 3.8 km/s in the upper layer to 6.5 km/s at 30 km depth. We point out that velocity values below 12 km depth are poorly constrained, since only a few events occurred at this depth range. In general, the depth to the Moho in the North Africa region is 25–30 km, where the P-wave velocity approaches 8 km/s (e.g., Research Group for Lithospheric Structure in Tunisia 1992).

As mentioned in Sect. 3, a station (SP04) was installed on December 11, after the first events had been located (1 week of data), and as expected, this station provided better constraints on the depths of the seismic events in the epicentral area. Indeed, the uncertainty in measuring depth has been substantially reduced (the average vertical error is 0.7 km). Figure 4 shows the relocated epicenters. Events recorded before the installation of the SP04 station indicated depths of around 6–12 km. After installation of the SP04 station, almost all the recorded seismicity is put at shallower depths of 0–3 km (Fig. 4).

For further work, we should therefore consider only those events recorded after installation of the SP04 station, for which the determined depths are more reliable. They represent 72 % of the subset selected above.

Magnitudes of the seismic events range from 0.8 to 3.9, and the seismicity is concentrated between the two small villages of Sidi Khelifa and Ain Tinn, about 3–4 km southeast of Jebel Akhal (Figs. 1, 4). No clear geometrical pattern can be observed on the seismicity map. However, a closer look at the distribution of epicenters allows us to identify a preferred alignment of epicenters in the NNW–SSE direction, and with a length of about 5 km and a width of 1.5 km.

4.2 Focal solutions

Focal mechanisms were calculated for the best located events, recorded by a maximum number of stations, by using P-wave first motion polarities detected on the vertical component at each station (Rivera 1991; focal mechanism determination SPHERA code, Research Report EOST, Strasbourg, France). The code SPHERA provides two parameters,

Table 1 Velocity model in the vicinity of the Beni Haroun dam (northeast Algeria)

P-wave velocity (km/s)	Depth (km)
3.85	0
4.96	2
5.60	12
6.54	30

The *Vp*/*Vs* ratios were estimated using the Wadati diagram and found to be equal to 1.75, which is in agreement with *Vp*/*Vs* values estimated in adjacent regions (Beldjoudi et al. 2009)

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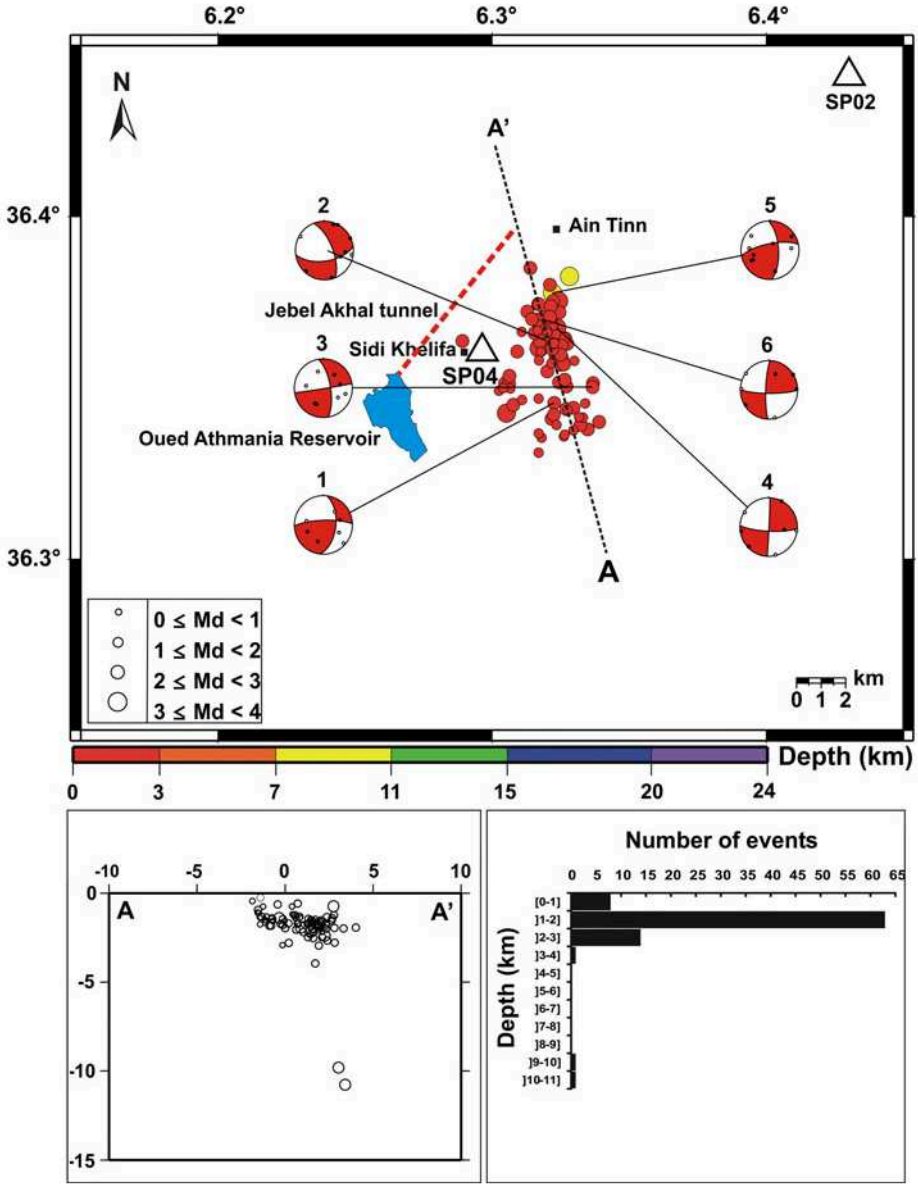

Fig. 4 *Horizontal* and *vertical* distributions of epicenters. The histogram shows the distribution of precisely located events with depth. Triangles are the nearest temporary stations to the epicenters. The *red dashed line* is the tunnel passing through the Jebel Akhal. A–A′ is a cross-section showing earthquake hypocenters. The figure also shows the 6 focal solutions calculated in this study. The reservoir of Oued Athmania is shown in *blue*

score and quality, which measure the misfit (the differences between polarities observed and predicted by the model).

In Fig. 4, we selected focal solutions with no discrepancies, and with a minimum number of P-wave polarities equal to seven. The six focal solutions show more or less the

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same shape. They show that strike-slip faulting was dominant along near vertical faults that trend close to the maximum horizontal stress direction, which is NW–SE (e.g., Beldjoudi et al. 2009, 2011). The solutions obtained are shown in Table 2, and all the largest events shown in this table were recorded at the same stations. The vertical errors for all these events are less than 0.7 km. It is probable that event number 5, which was at the greater depth of about 10 km, has nothing to do with the triggered seismicity at shallower depths. One of the fault plane solutions has a nearly vertical dip and strikes N–S (with a slight variation in the strike angle between N170° and N180°), and this is consistent with the overall alignment of seismic events (NNW–SSE), which suggests that the deformation is accommodated on a single planar structure.

4.3 Stress axis calculations

The calculated earthquake focal mechanisms were applied to the GRIDSIM_STRESS algorithm (Delouis et al. 2002) to determine the orientations of the main axes of the stress tensor ellipsoid. The critical thing when selecting the best stress tensors is to minimize the angular deviation between the observed and computed slip vectors on the nodal planes of the focal mechanisms. We defined a score, which is the average of the scalar products of the observed and computed slip vectors (a scalar product equal to one means a perfect matching, or in other words, no angular deviation). Minimizing the angular deviations is equivalent to maximizing the score. The stress tensor shape factor R is defined as in Rivera and Cisternas (1991), where the domains ($R < 0$), ($0 < R < 1$), and ($R > 1$) correspond, respectively, to compressional, strike-slip, and extensional stress regimes.

The main axes of the stress tensor ellipsoid are plotted in Fig. 5. The results were obtained using the 6 focal solutions calculated in the previous section. The best score value is 0.99, which indicates the results are well constrained. The best solution (black circle) indicates that the compression axis σ_1 is close to horizontal, and oriented NW–SE (N314° ± 25°), as found by Beldjoudi et al. (2009) for the 2006 Laalam earthquake (Mw = 5.2) about 80 km west of the dam region. The extension axis σ_3 is also close to horizontal, and oriented NE–SW (N212° ± 25°). The value of the shape factor R is 0.5, which corresponds to a strike-slip stress regime.

5 Discussion and conclusions

According to the Algerian earthquake catalogue, the area around the hydraulic complex is not considered to be very active, and no significant activity around this area had been observed before the 2007 December earthquake swarm. Because the locations where the first events were felt are in the area of the hydraulic construction, the probability that this seismicity could have been triggered is not negligible. In many cases, small and moderate seismic events have been associated with the poroelastic response of the crust to water level changes behind dams and in reservoirs (Gupta 1983, 1985; Talwani 1997, and references therein; Torcal et al. 2005; Valoroso et al. 2009), particularly when the observed seismic activity is located within a 10–25 km radius of the dam (Simpson 1976; Rastogi et al. 1997; Zhong et al. 1997; Torcal et al. 2005). Therefore, it was obvious that we should consider, in our preliminary analysis of the earthquake swarm, that the seismic activity recorded in the Mila region could have been caused by human activity.

The analysis of the temporal water level variation behind the dam versus the seismic activity shows three periods when there is an association with changes in the sign of the

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Table 2 Focal mechanisms parameter calculated for the largest events located between the Beni Haroun dam and the Oued Athmania reservoir

No	Date	Origin time	Md	Location		Depth (km)	Nodal plane A			Nodal plane B		
				Lon (°) E	Lat (°) N		Strike (°)	Dip (°)	Rake (°)	Strike (°)	Dip (°)	Rake (°)
1	07/12/12	18:51:31	2.7	6.32276	36.34559	0.568	6.3	54.4	15.7	267.9	77.2	143.4
2	07/12/13	07:0.6:33	3.3	6.31956	36.36381	1.902	82.1	60.3	−153.2	338.1	66.9	−32.6
3	07/12/13	13:31:13	2.9	6.33669	36.35022	1.731	80.9	88.7	−159.6	350.4	69.6	−1.4
4	07/12/17	22:23:50	3.0	6.32365	36.36763	1.725	2.7	89.0	−11.6	92.9	78.4	−179.0
5	07/12/18	11:49:41	3.9	6.32214	36.37768	9.812	258.6	72.1	154.5	356.6	65.8	19.7
6	07/12/18	20:31:56	3.3	6.31918	36.36984	2.386	90.4	80.3	170.4	182.0	80.6	9.9

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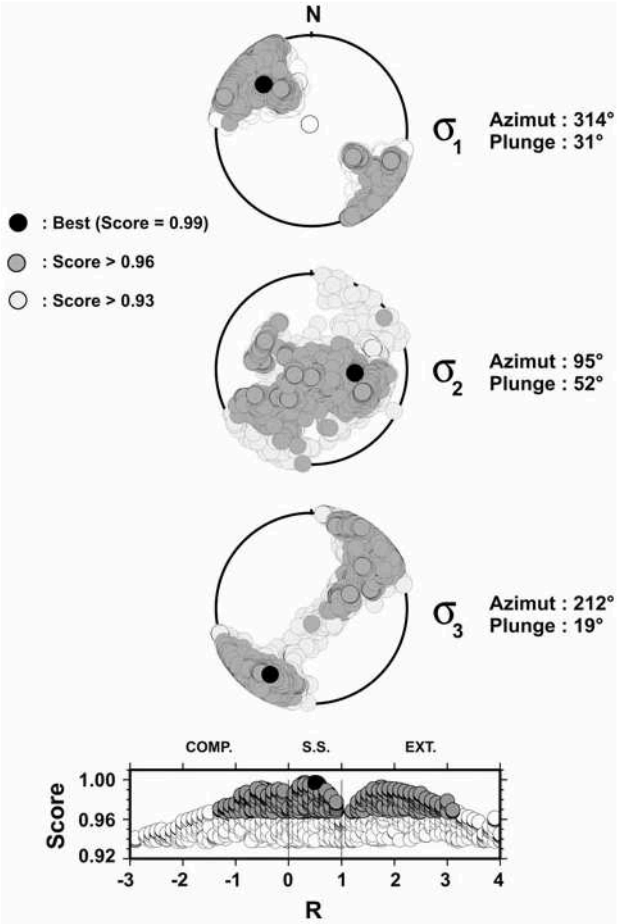

Fig. 5 Stress tensor inversion. Solutions for σ_1, σ_2, and σ_3 are plotted in three lower hemisphere equal-area projections. The azimuths and plunges of the principal stress directions are shown. The best solution (score) is represented by a *black dot*. The score here is the average of the scalar products of the observed and computed slip vectors. Solutions with a score superior to 0.96 and 0.93 are represented by *dark gray* and *light gray*, respectively. At the bottom, the same solutions are plotted as a function of their R values. The shape factor $R = (\sigma_z - \sigma_x)/(\sigma_y - \sigma_x)$, where σ_x, σ_y, and σ_z are the principal stresses, with $\sigma_y > \sigma_x$, and σ_z close to vertical, in order to interpret the shape factor R in a tectonic sense. The relationship of the non-ordered Eigen values σ_x, σ_y, and σ_z with respect to the ordered Eigen values $\sigma_1 > \sigma_2 > \sigma_3$ depends on R, and is shown in Legrand et al. (2005). For the strike-slip regime $(0 < R < 1)$ $\sigma_y > \sigma_z > \sigma_x$, then $\sigma_y = \sigma_1$, $\sigma_z = \sigma_2$, and $\sigma_x = \sigma_3$. The domains corresponding to the compressional $(R < 0)$, strike-slip $(0 < R < 1)$, and extensional $(R > 1)$ regimes are labeled, respectively, COMP., S.S., and EXT. Here, the R factor is equal to 0.5, indicating a strike-slip regime

slope of the water level curve (Fig. 6). From the end of October 2007 till November 21, 2007, the date at which the pumping operation started (for the water conveyance operation), we can observe an elevation in the water level behind the dam (phase 1). During this phase, only the permanent seismic network was operational, and at the nearest seismic station, about 20 km away, no microearthquakes were observed. During the following days, from November 21 to December 11 (phase 2), the date at which the pumping operation was stopped, the water level declined behind the Beni Haroun dam. The first

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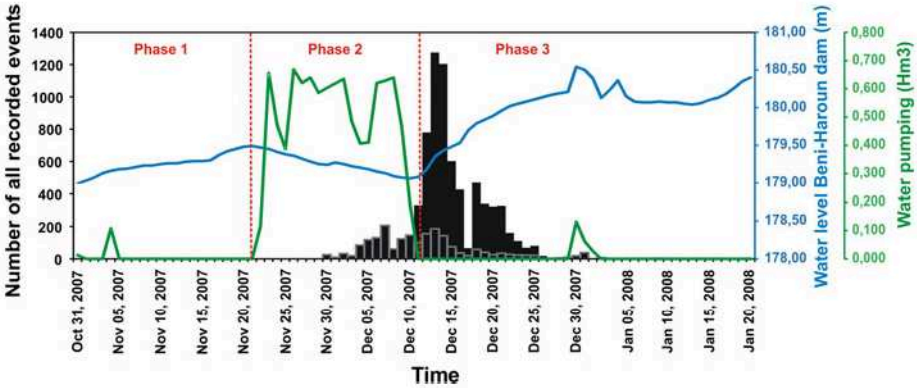

Fig. 6 Daily number of recorded events. The *black histogram* shows events recorded by the entire network (permanent and temporary), and the gray histogram shows events recorded by the entire network minus SP04 (epicentral added station). The water level behind the Beni Haroun dam is shown as a *blue line*, and the pumped water volume as a *green line*. This figure shows the pattern of activity and the timing of peak activity without any artifact due to the installation of the epicentral station (SP04)

triggered events were recorded on November 30, 9 days after the beginning of the pumping operation. After that, the seismic activity continued to increase until it reached its maximum on December 13, after which it declined rapidly and almost vanished completely on December 31 (late phase 3). The correlation of seismic activity with the variations in water level shows that seismic activity did not start during the impoundment (phase 1), but began during the pumping operation (phase 2), a few days after it was launched. The seismic activity continued to increase until reaching a maximum on December 13, even though the pumping operation had been stopped 2 days before that. The fact that seismicity is absent or significantly reduced during the two phases of impoundment (phase 1 and late phase 3) suggests that the seismic activity is not related to the changes in water level behind the dam, but rather to the water pumping operation.

For the purposes of showing the pattern of activity, including the timing of peak activity, and removing any artifact that could have been caused by the installation of the epicentral station (SP04), the gray histogram of Fig. 6 shows the daily number of events recorded by the entire network, minus the events recorded by the epicentral station SP04. The daily number of events recorded by the entire network (including station SP04) is shown by the black histogram. If one compares the black histogram with the pumping curve, one could say that activity peaked significantly when pumping stopped (Fig. 6). However, this is actually an artifact of the SP04 station installation, and the gray histogram shows the activity without that artifact.

From January 2006 to December 2011 (Fig. 7), a long-term comparison of the seismic activity in the region versus the pumping rate and the water levels behind the dam showed clearly that the sudden increase in seismic activity was a one-off event related to the pumping operation. The seismicity shown in Fig. 7 was recorded on the permanent network, and this figure shows a smaller peak of seismicity in August 2007. This date corresponds to a pumping test in early August, before the official start of the pumping operation on November 21, 2007. Seismic activity returned to normal a few months after these pumping operations, and after the problem of the leak in the tunnel had been resolved.

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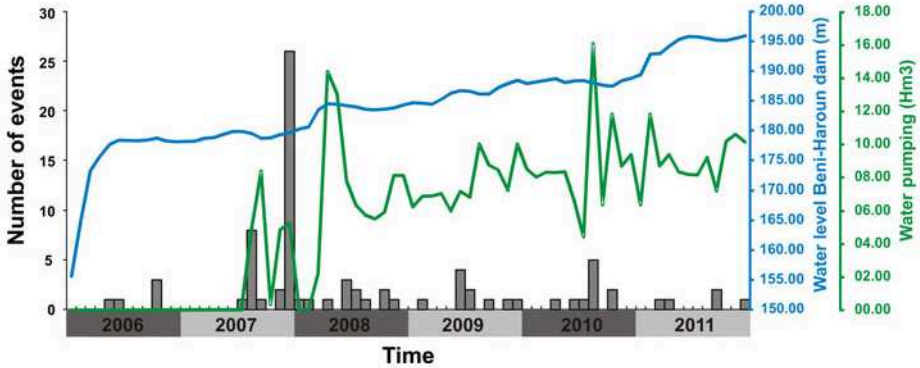

Fig. 7 Long-term context. The histogram shows the seismic activity in the region recorded by the permanent network ($M > 2$) from January 2006 to December 2011 versus the level of water behind the dam (*blue curve*) and the pumped volume of water (*green curve*)

This investigation, carried out in conjunction with the engineers of the Hydraulic Service of the ANBT (National Agency of Dams and Transfers), leads us to conclude that the seismic activity of the studied area was triggered by an unintentional water injection process. This deduction is validated by several findings: (1) during the pumping operation, it was observed that only 45 % of the pumped water was recovered at the Oued Athmania reservoir; (2) a complementary investigation made by the company in charge of the construction of the tunnel through the Jebel Akhal found defective joints between the prefabricated segments of the lined tunnel inside the Jebel Akhal. We concluded, therefore, that the defective joints allowed a large part of the pressurized water to penetrate the Jebel Akhal and its surrounding faults and fractures, triggering the seismicity observed. Another important point is that eyewitness reports noted hot water coming out of the hill through preexisting fissures during the pumping operation phase. The preexisting fractures in Jebel Akhal may have provided effective and favorable conditions for the penetration of pumped water into the ground, with the resultant increase in pore pressure reducing the rock strength and triggering the release of elastic strain of tectonic origin, as seen elsewhere around the world (Bell and Nur 1978; Shapiro et al. 2006). In our case, the triggered seismicity was observed 3–4 km SE of the Jebel where the faults may possibly have been closer to failure. With regards to the heating of the water, we have had to leave this problem for future research, due to a lack of literature and other data related to what might have caused the heating.

Triggered seismicity following the injection of water, mainly from boreholes, has been observed in several places elsewhere. In 1970, in central Japan (Matsuhiro), an experimental test for examining the relationship between pore pressure and earthquakes was conducted in a deep well. Several days after the beginning of fluid injection, a series of microearthquakes were initiated around the well at a distance of 2–4 km (Ohtake 1974). Other examples of water injection induced earthquakes include the Nojima fault zone experiment in Japan, and the KTB project in Germany, both of which have some similarities with our case, such as the distance of the epicenters from the injection point (about 3–4 km in all cases). There are also some differences, as for example in the time elapsed after the beginning of water injection. In the Matsushiro area, seismicity was activated between 4.8 and 9.3 days after the beginning of injections (there were three injections), whereas in the case of KTB, it started only 2 h after the beginning of injection. In our case

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in Algeria, seismicity started 9 days after the beginning of the unintentional water injection.

The problem of the water leaking inside the tunnel was resolved a few months after these seismic events. Today, the reservoir behind the dam is full to maximum capacity, with 1 billion m^3 of water stored since the fall of 2011, and the transfer of water between the two reservoirs is now regularly done without triggering any more seismic events.

In conclusion, we note that this is the first time that Algeria has experienced seismic activity associated with human activity. The low-magnitude shallow earthquakes, located mainly with the aid of a temporary digital seismic network, showed that the earthquake swarm was concentrated in an area immediately adjacent to the Beni Haroun hydraulic complex. The vertical and horizontal distributions of hypocenters show one main cluster that is aligned in a NNW–SSE direction, consistent with one of the fault plane focal solutions, and which may be related to a strike-slip vertical fault that trends close to the maximum horizontal stress direction. No correlation was found between the changes in water level behind the Beni Haroun dam and the earthquake swarm, but a clear correlation exists between the water pumping operation and the seismic activity. A leakage of pressurized water through defective joints in the tunnel through the Jebel Akhal would have increased the interstitial pore water pressure in the rock matrix and the faults in the vicinity of the Jebel, thus reducing the strength of the rocks and resulting in the releases of tectonic elastic strain. The triggered seismicity was located 3–4 km southeast of the Jebel Akhal, even though faults reportedly exist within the Jebel Akhal itself. This means that the faults within the immediate vicinity of the Jebel are either not active or not close to failure. Today, seismicity levels are back to normal as a result of resolving the leakage problem in the tunnel through the Jebel.

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