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Variations latérales de la déformation crustale en Himalaya

Théo Berthet

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UNIVERSITE DE MONTPELLIER 2 - SCIENCES ET TECHNIQUES DU LANGUEDOC

ÉCOLE DOCTORALE SIBAGHE

THÈSE

pour obtenir le titre de

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Spécialité : Sciences de la Terre et de l'Univers

VARIATIONS LATÉRALES DE LA DÉFORMATION CRUSTALE EN HIMALAYA

Présentée et soutenue publiquement par :

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soutenue le 27 Novembre 2013, devant un jury composé de :

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Lateral variations of crustal deformations in the Himalayas

Abstract

Several major earthquakes have struck the 2500 km long Himalayan range during the past century. However, both the maximum size of such earthquakes and the probability of occurrence of a magnitude 9 megaquake are still a matter of debate. In this framework it is of key importance to investigate the lateral segmentation of the Himalayan arc in order to deepen our understanding of the mechanisms that control such events as well as their extents. Besides the remarkably uniform tectonic set-up and the fact that most studies on the Himalayas have focused on Central Nepal, several indications of along strike changes can be gathered. This thesis brings new constraints on lateral variations of crustal structure and of deformation between western Nepal ($\sim 80^\circ\text{E}$) and eastern Bhutan ($\sim 92^\circ\text{E}$).

The first part of this thesis focuses on the lateral variations of crustal structure. In the past three years we have carried out four land gravity campaigns. We have established 366 new gravity measurement points to fill the data gaps in central and western Nepal as well as in Bhutan. This new dataset, together with available seismological data, is used to constrain thermo-mechanical models of the bending of the India plate underneath the Himalayas. While the inferred crustal geometry does not exhibit major along-strike variations over central and eastern Nepal, the shorter wavelength flexure of the lithosphere in Bhutan is associated with a weaker upper mantle rheology.

The second part focuses on the lateral variations of recent deformation in the Himalayas. Fifteen years of paleoseismological investigations have extended the catalogue of major Himalayan events over the last millennium. Combining these information with secular loading, we assess the spatial and temporal stress changes on the Main Himalayan Thrust along the orogen over the last nine centuries. Our calculations indicate that inter-seismic loading has now nearly overcome the Coulomb stress decrease caused by the great ~ 1100 medieval earthquake. Our results also point out that the 1897 Shillong plateau earthquake does not have a major influence on the stress accumulated on the Main Himalayan Thrust since then. In order to better characterize active tectonics in the Bhutan Himalayas where no studies were done so far, we carried out a morphotectonic analysis in the south-central part of the kingdom. We show that the same amount of Holocene deformation is accommodated on the frontal thrust in Bhutan as on the neighbouring portions of the Himalayan arc. We also find evidences for two major ($M > 8$) earthquakes on this thrust in the Bhutan Himalayas during the last millennium. Our results therefore show that Bhutan cannot be considered as a seismic gap. They also challenge the interpretation of one single ~ 1100 medieval mega-event that would have ruptured the Main Frontal Thrust from central Nepal to eastern Bhutan.

Variations latérales de la déformation crustale en Himalaya

Résumé

Au cours du dernier siècle, plusieurs séismes majeurs ont affecté l'Himalaya. Cependant, la taille maximale de ces événements et la probabilité d'occurrence de méga-séismes avec des magnitudes proches de 9 sont toujours matière à débat. L'étude de la segmentation de l'arc Himalayen est donc primordiale afin de comprendre les mécanismes qui contrôlent ces séismes ainsi que leur extension spatiale. La compréhension du cycle sismique en Himalaya est aujourd'hui essentiellement basée sur des études menées au Népal central, ce qui limite notre connaissance de son fonctionnement tri-dimensionnel. Ce travail de thèse permet d'apporter de nouvelles contraintes sur les variations latérales de la déformation crustale dans la zone comprise entre l'ouest Népal ($\sim 80^\circ\text{E}$) et l'est Bhoutan ($\sim 92^\circ\text{E}$).

La première partie de cette thèse est consacrée à l'étude des variations latérales de la structure lithosphérique. Quatre campagnes gravimétriques ont été réalisées entre 2010 et 2012. Elles ont permis d'étendre le réseau gravimétrique Népalais jusqu'à l'ouest Népal et d'établir le premier réseau gravimétrique Bhoutanais. Ce nouveau jeu de données, combiné avec les données sismologiques disponibles, permet de contraindre des modèles thermo-mécaniques de la flexure de la plaque Indienne sous l'Himalaya. Les résultats suggèrent qu'il n'existe pas de variations latérales majeures du comportement mécanique de la lithosphère Indienne entre le centre et l'est Népal contrairement au Bhoutan où une rigidité flexurale plus faible est nécessaire pour expliquer les données.

La deuxième partie de cette thèse est consacrée à l'étude des variations latérales des déformations récentes dans le prisme Himalayen. Les études paléosismologiques menées depuis 15 ans en Himalaya ont permis d'étendre le catalogue des événements majeurs sur le dernier millénaire. En intégrant à la fois les contraintes disponibles sur ces séismes majeurs et le chargement séculaire, nous étudions les variations spatiales et temporelles de l'état de contrainte du chevauchement Himalayen principal le long de l'arc. Nos calculs montrent que même dans le cas où un méga-séisme se serait produit en ~ 1100 en Himalaya, le chargement séculaire a quasiment compensé la chute de contrainte associée. Les modèles montrent aussi que le séisme du Shillong ne joue pas un rôle majeur sur les contraintes accumulées sur le chevauchement Himalayen depuis 1897 au niveau du Bhoutan. Enfin, nous présentons les premières contraintes sur la tectonique active au Bhoutan. L'étude morphotectonique réalisée au centre du Sud-Bhoutan a permis de montrer que la majeure partie de la déformation Holocène est accommodée au niveau du front Himalayen, comme au Népal. Nous montrons aussi des évidences de ruptures de surface avec des décalages verticaux de plusieurs mètres associés à deux séismes majeurs sur le dernier millénaire. Ces évidences remettent en cause l'interprétation d'un méga-séisme en ~ 1100 .

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CHAPITRE 1

INTRODUCTION

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1.1 Contexte et problématique générale de la thèse

Les séismes majeurs ont lieu principalement dans les zones de convergence lithosphérique et notamment au niveau des zones de subduction qui concentrent 80 % des séismes de magnitude supérieure à 8. Ces zones de subduction sont, à ce jour, les seuls endroits sur Terre ayant généré des méga-séismes instrumentaux dont la magnitude est supérieure à 9 (Tohoku 2011 - M9.0; Sumatra 2004 - M9.2; Alaska 1964 - M9.2; Chili 1960 - M9.5). Ce type de séisme est d'autant plus dangereux qu'il peut générer des tsunamis parfois plus meurtriers que le séisme en lui-même comme ce fut le cas en 2004 suite au tremblement de terre de Sumatra (M9.2).

Les zones de subduction sont donc des zones d'études privilégiées depuis plusieurs dizaines d'années pour essayer de comprendre les mécanismes contrôlant ces méga-ruptures. De nombreuses données géophysiques ont permis d'étudier les mécanismes d'accommodation de la déformation des ces zones de subduction ainsi que d'en imager la structure (eg *Klingelhoefer et al. 2012*). Les mesures de déformations cosismiques réalisées grâce aux stations GPS à terre pendant le séisme de Sumatra (*Chlieh et al. 2007*) ou encore celui de Maule en 2010 (*Vigny et al. 2011*) ont permis une meilleure compréhension des mécanismes associés aux ruptures. Si les récentes observations de déformations cosmiques importantes près de la fosse lors du séisme de Tohoku (*Kodaira et al. 2012*) ont beaucoup interrogé la communauté scientifique et continue à le faire, elles ont aussi mis en évidence une des limitations majeures des études en zones de subduction. En effet, leur caractère immergé limite la distribution spatiale des mesures de déformation de surface et également la localisation précise des séismes et ce malgré le développement de l'instrumentation fond de mer.

Au contraire des zones de subduction, l'arc Himalayen permet le déploiement de réseau instrumentaux avec une meilleure distribution spatiale pour mesurer les déformations intersismiques (eg *Ader et al. 2012*), localiser la sismicité (eg *Pandey and Agrawal 1999*) ou encore imager les structures crustales (eg *Nabelek et al. 2009*). De plus, l'accès au terrain proche de la zone frontale (l'équivalent de la fosse en zone de subduction) permet d'étudier les déformations récentes (*Lavé and Avouac 2000*) ou encore l'occurrence de paléo-séismes (*Kumar et al. 2010*). Au cours du dernier siècle, plusieurs séismes majeurs ont affecté l'Himalaya (Cachemire 2005 - M7.6; Assam 1950 - M8.5; Bihar-Népal 1934 - M8.1; Kangra 1905 - M7.8). Cependant, la taille maximale des séismes et la probabilité

d'occurrence de méga-séismes avec des magnitudes proches de 9 est toujours matière à débat.

La magnitude d'un séisme dépend principalement de la quantité de glissement et de l'extension de la rupture associée. C'est pourquoi l'étude de la segmentation des zones de convergence est primordiale afin d'estimer la longueur des tronçons pouvant être rompus. Dans les zones de subduction, la segmentation est classiquement liée au comportement frictionnel sur l'interplaque. La structure de la plaque subduite joue aussi un rôle essentiel : les reliefs océaniques (failles, édifices volcaniques) peuvent à la fois servir d'aspérités pouvant déclencher la rupture ou de barrière à sa propagation (eg *Dominguez et al. 2000*). À partir de données InSAR, *Bejar-Pizarro et al. (2013)* ont récemment proposé que les structures de la plaque supérieure puissent aussi influencer les propriétés frictionnelles sur l'interplaque et donc sa segmentation. En Himalaya, la compréhension du cycle sismique est essentiellement basée sur des études menées au Népal Central (eg *Avouac 2003*) ce qui limite notre connaissance de son fonctionnement tri-dimensionnel. Plusieurs questions fondamentales restent donc en suspens :

- Est-il possible de générer des méga-séismes dans le contexte Himalayen ?
- Quelles sont les différences fondamentales entre l'Himalaya et les zones de subduction ?
- Qu'est ce qui contrôle la segmentation des séismes majeurs le long de l'Himalaya ?
- Existe-t-il des variations latérales de couplage intersismique à l'origine de la segmentation des séismes comme c'est le cas dans les zones de subduction.
- Quelle sont les interactions entre le chargement séculaire et les séismes majeurs en Himalaya ?
- La notion de séismes caractéristiques souvent employée au niveau des grands décrochements intracontinentaux (eg *Rizza et al. 2011*) est-elle applicable en Himalaya ?
- Existe-t-il des variations structurales majeures dans l'arc Himalayen ? Si oui, sont-elles à l'origine de la segmentation des séismes majeurs en Himalaya ?
- Quel est l'état de contrainte des failles majeures en Himalaya et présente-t-il des variations latérales ?

Répondre à ces questions nécessite une approche qui se doit de combiner diverses disciplines comme la gravimétrie, la géodésie, la sismologie, la morphotectonique, la paléosismologie ou encore la modélisation numérique. Ce travail de thèse permet d'apporter des éléments de réponse à certaines de ces questions dans une zone couvrant plus de 1200 km

de l'arc Himalayen comprise entre l'ouest Népal ($\sim 80^\circ\text{E}$) et l'est Bhoutan ($\sim 92^\circ\text{E}$).

Dans la suite de ce chapitre, après une rapide présentation de l'accommodation de la convergence Inde-Eurasie, je montrerai que malgré l'apparente cylindricité topographique et géologique de la chaîne Himalayenne, il existe des évidences de variations latérales de premier ordre. Ensuite, je présenterai les différentes études menées au cours de cette thèse.

1.2 Accommodation de la convergence Inde-Eurasie

L'Himalaya et le plateau Tibétain résultent de la collision entre L'Inde et l'Eurasie (eg *Powell and Conaghan 1973*). Suite à la séparation du Gondwana, il y a 120-130 Ma (*Metcalfe 1996*), le déplacement de la plaque Indienne vers le nord à une vitesse de $\sim 15 \text{ cm.an}^{-1}$ a entraîné la subduction progressive de la portion de l'océan Téthysien comprise entre l'Inde et l'Eurasie (*Dercourt et al. 1993*). La reconstruction du mouvement de l'Inde par rapport à l'Eurasie montre une diminution brutale du taux de convergence vers 50 Ma (Fig. 1.1). Ce processus est attribué à l'arrivée de la marge continentale passive Indienne au contact de la marge active Eurasiatique et a pour conséquence de diminuer le taux de convergence de $\sim 10 \text{ cm.an}^{-1}$ pour se stabiliser à $\sim 5 \text{ cm.an}^{-1}$ (eg *Guillot et al. 2003*). La collision a d'abord commencé à l'Ouest Himalayen et a migré vers l'Est (eg *Rowley 1996*; *De Sigoyer et al. 2000*; *Ding et al. 2005*).

Le principe de sous-plaquage du continent Indien sous l'Eurasie a initialement été évoqué par *Argand (1924)* (Fig. 1.2A). Il aura fallu attendre près d'un siècle pour imager de manière claire ce phénomène et montrer la présence du sous-continent Indien sous le Tibet sur plus de 450 km au Nord du front Himalayen (Fig. 1.2B). Depuis le début de la collision continentale, l'accommodation de la déformation se fait par une combinaison d'épaississement crustal et d'extrusion latérale (*Molnar and Tapponier 1975*). Le raccourcissement total estimé par la reconstruction du mouvement relatif des plaques Indienne et Eurasiatique est de $2600 \pm 900 \text{ km}$ dont 1700 ± 600 sont liés à la déformation du bloc Eurasiatique (*Patriat and Achache 1984*; *Achache et al. 1984*; *Besse and Courtillot 1988*) et le reste par le raccourcissement Himalayen (*Le Pichon et al. 1992*).

Actuellement, les taux de convergence déterminés par GPS au travers de la chaîne Himalayenne sont de l'ordre de $15\text{-}20 \text{ mm.an}^{-1}$ (*Bilham et al. 1997*; *Larson et al. 1999*; *Jouanne et al. 2004*; *Bettinelli et al. 2006*; *Jade et al. 2007*; *Mukul et al. 2010*; *Ader et al.*

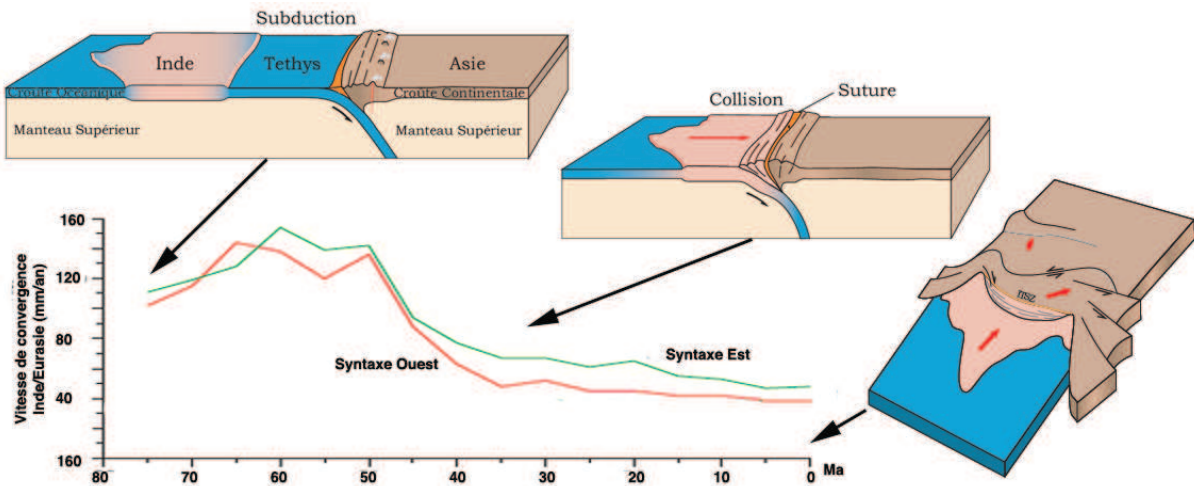


FIGURE 1.1 Vitesse de l'Inde par rapport à l'Eurasie au cours du temps. La ligne rouge correspond à l'extrémité Ouest de la suture Himalayenne et la verte à l'extrémité Est. Les schémas géodynamiques illustrent la subduction océanique de la Tethys sous le continent asiatique (~ 80 Ma), l'entrée en subduction de la marge continentale Indienne et les effets actuels de la collision Indo-Eurasienne. Modifié d'après *Malavieille et al. (2002)*.

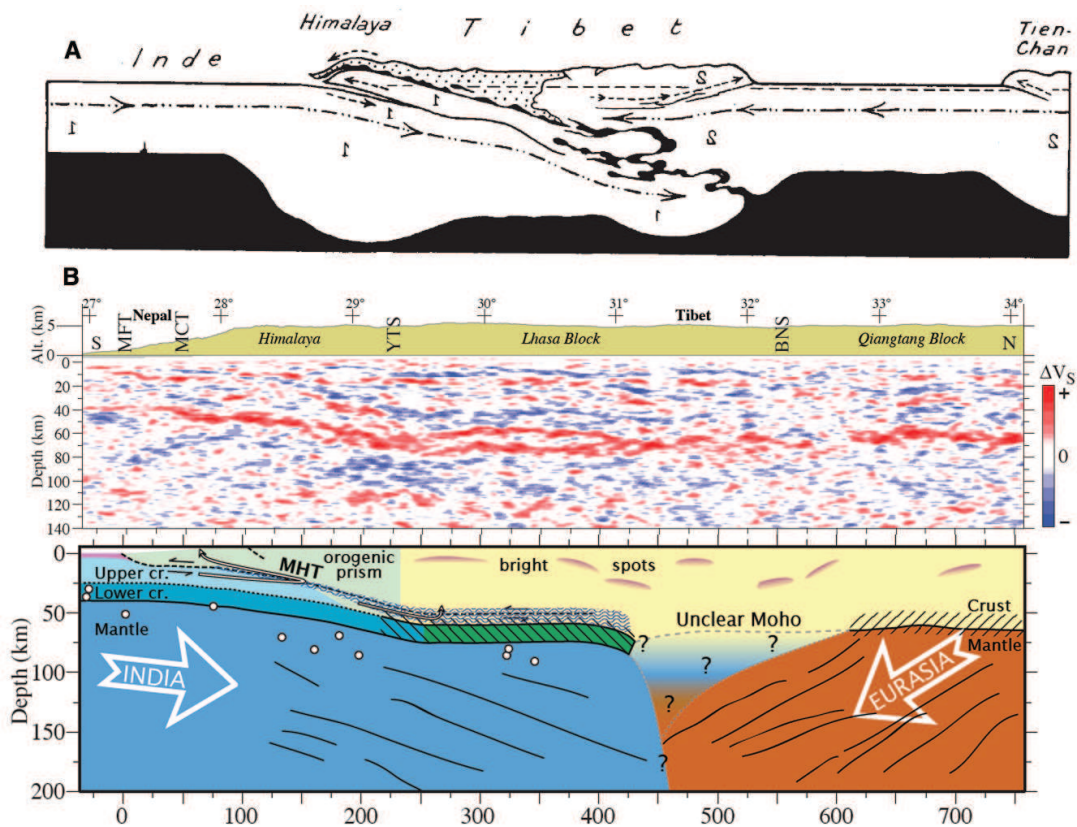


FIGURE 1.2 (A) Coupe schématique interprétative du système Himalaya-Tibet par *Argand (1924)*. (B) Profil de fonctions récepteurs obtenu par l'expérience Hi-CLIMB et schéma interprétatif (*Nabelek et al. 2009*).

2012). La plupart de cette déformation est accommodée au niveau du front Himalayen comme l'attestent les taux de convergence Holocène déterminés à partir de la datation des terrasses alluviales (*Lavé and Avouac 2000; Burgess et al. 2012*). Cependant, l'existence d'une activité hors-séquence du MCT est proposée par certains auteurs (eg *Wobus et al. 2005*). La majeure partie du chargement séculaire est relâchée lors de séismes majeurs comme celui de Bihar-Népal en 1934 (M 8.1). Depuis 15 ans, les études paléosismologiques en Himalaya ont permis de mettre en évidence de nombreux séismes majeurs en Himalaya depuis le dernier millénaire. Cependant l'existence de méga-événements avec des magnitudes approchant 9 est toujours débattue (*Lavé et al. 2005; Kumar et al. 2010; Mugnier et al. 2013*). L'étude de la segmentation, ou plus précisément des variations latérales le long de l'arc Himalayen est donc cruciale pour aborder l'aléa sismique de cette région.

1.3 Cylindricité de la chaîne Himalayenne ?

Afin d'étudier ces variations latérales, je présente ici les différentes évidences géomorphologiques, géologiques et géophysiques qui remettent en cause la vision classique d'un arc Himalayen à géométrie cylindrique.

1.3.1 Aspects géomorphologiques

Au premier ordre, la chaîne Himalayenne, longue de 2500 km, présente une topographie relativement cylindrique séparée en 4 unités principales (Fig. 1.3A). Cette subdivision est essentiellement basée sur la situation au Népal, du Sud au Nord on trouve :

- Les plaines du Gange et du Brahmapoutre qui présentent un relief faible et une altitude de l'ordre de 100 à 200 m.
- Le piémont et le moyen-pays, larges d'une centaine de km, et dont l'altitude augmente régulièrement avec une moyenne de l'ordre de 2000 m.
- La haute chaîne, zone de très forts reliefs, qui est constituée par les plus hauts sommets Himalayen. Elle s'étend sur une zone d'environ 50 km où l'altitude moyenne est de 5000 m et où les pentes sont très fortes.
- Le plateau Tibétain dont l'altitude moyenne est aussi de l'ordre de 5000 m mais qui présente des reliefs moins marqués que dans la haute chaîne.

Néanmoins, une analyse plus détaillée permet de remettre en cause cette cylindricité de premier ordre. *Duncan et al.* (2003) montrent à partir de profils N/S d'altitudes et de pentes moyennées sur une largeur de 50 km des différences importantes entre le Népal central et le Bhoutan. La principale transition topographique au Népal central est située au sud de la haute chaîne. Elle permet de passer d'une altitude moyenne de 5000 m au Nord à 2000 m sur une largeur de quelques dizaines de km (Fig. 1.3C). Cette transition majeure n'est pas visible sur le profil Bhoutanais (Fig. 1.3D). À cette longitude, les pentes sont beaucoup plus régulières et la seule transition topographique marquée est située au front de la chaîne.

La comparaison de la topographie et des précipitations moyennes montre une corrélation très nette (Fig. 1.3A-B). Au Népal central, les précipitations sont réparties en deux bandes parallèles qui correspondent aux deux transitions topographiques situées au niveau du front du moyen-pays et de la haute chaîne. Au niveau du Bhoutan, les précipitations sont concentrées au front de la chaîne. Sur la moitié est du Bhoutan, la présence du plateau du Shillong 150 km au sud du front de la chaîne, constitue une barrière orographique qui concentre une partie des précipitations. Ceci induit une variation latérale de la quantité de précipitation au Bhoutan, la partie est étant plus 'sèche' que la partie ouest.

Si une corrélation a été proposée entre cette variation latérale de précipitation au Bhoutan et les taux d'érosion calculés à partir des données thermochronologiques de traces de fissions sur apatites (*Grujic et al.* 2006), l'étude récente d'*Adlakha et al.* (2013) à partir d'un jeu de données plus complet et incluant les données de *Grujic et al.* (2006) montre qu'il n'existe pas de lien direct entre les précipitations et les taux d'érosion et de déformation long-terme. Au Népal, *Godard et al.* (2013) propose plutôt un contrôle tectonique des taux d'érosion déduits de données de cosmonucléides ^{10}Be .

1.3.2 Aspects géologiques

D'un point de vue structural, l'Himalaya est limité au sud par le Main Frontal Thrust (MFT) et au nord par la suture de l'Indus-Tsangpo. En plus des bassins flexuraux d'avant-pays, le prisme orogénique est composé d'une série de failles d'échelle crustale séparant 4 domaines géologiques continus incluant du sud au nord (Fig. 1.4) :

- Les Siwaliks ou Sub-Himalaya, constitués des dépôts Miocène à Quaternaire des

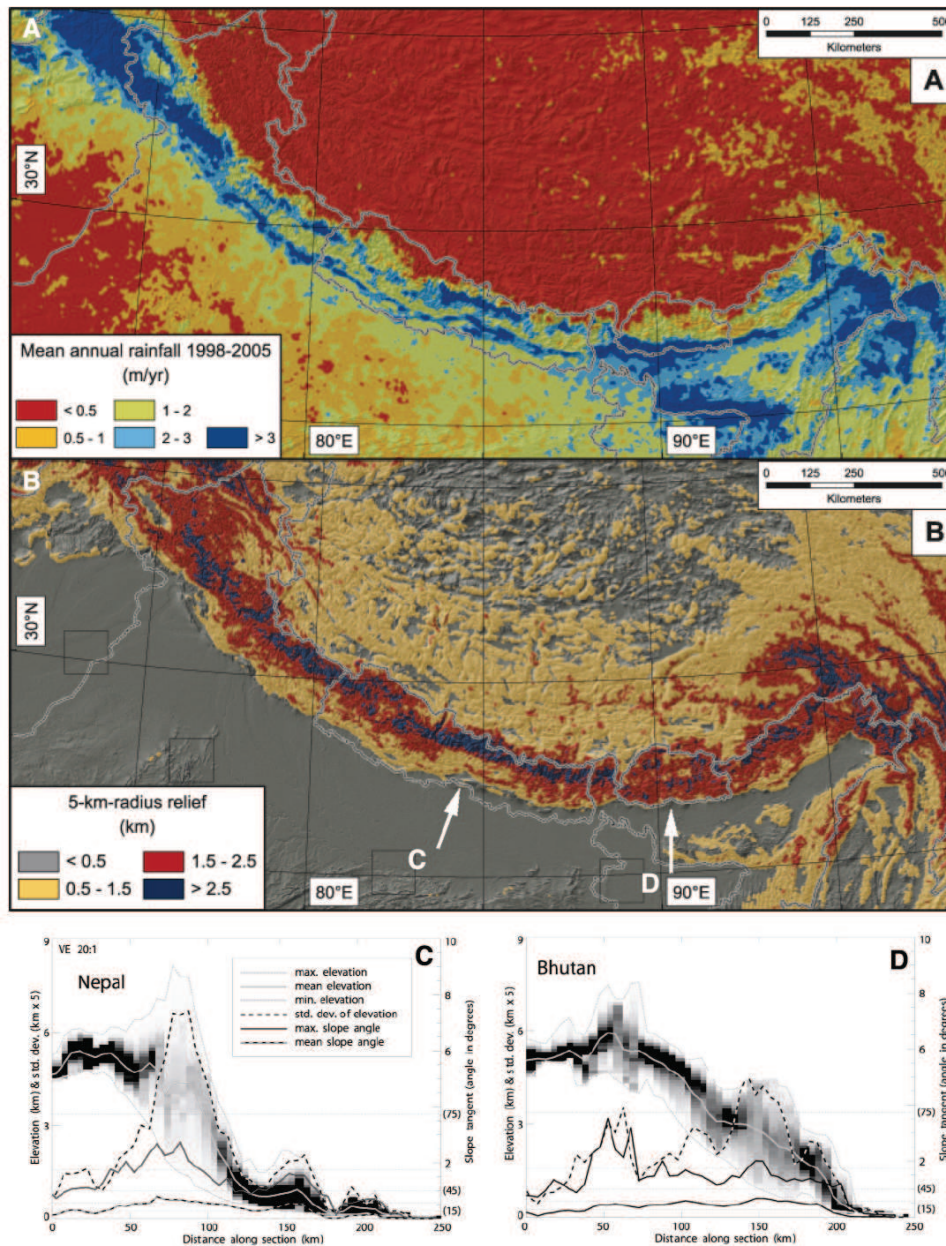


FIGURE 1.3 (A) Moyenne des précipitations lors de la mousson de 1998 à 2005. (B) Relief calculé sur un rayon de 5 km à partir des modèles numériques de terrain SRTM (résolution : 90 m). Une bande à fort relief est marquée au niveau de l'Himalaya central. Elle est absente au Bhoutan. D'après *Bookhagen and Burbank (2006)*. Profils N/S de l'altitude et des pentes (moyennés sur une largeur de 50 km) au Népal (C) et au Bhoutan (D). D'après *Duncan et al. (2003)*.

bassins d'avant-pays et plissés lors de la propagation frontale de la déformation. Ces séries chevauchent les dépôts du bassin d'avant pays actuel au niveau du MFT qui délimite donc leur extension au Sud. Elles sont principalement composées d'une alternance de grès, d'argiles et de conglomérats.

- Le moyen-pays ou Lesser Himalaya (LH), composé principalement de roches sédimentaires précambriennes déposées sur la marge Indienne et présentant un faible degré métamorphique (faciès schiste vert) (*Stöcklin 1980*). Au nord, ce domaine inclut également des métapelites (*Upreti 1999*). Le LH chevauche les Siwaliks au niveau du Main Boundary Thrust (MBT).
- Le Haut Himalaya ou Higher Himalaya (HH), constitué de métasédiments du Protérozoïque supérieur au Cambrien intrudés par des complexes cristallins (*Gansser 1983*). Il présente un degré de métamorphisme moyen à fort (faciès amphibolite). Ce domaine qui constitue une partie des plus hauts sommets Himalayen est délimité au sud par le Main Central Thrust (MCT).
- Les séries Téthysiennes ou Tethyan Himalaya (TH), composées de roches sédimentaires silicoclastiques et carbonatées déposées sur la marge Indienne entre le Cambrien et l'Éocène. On les retrouve entre la suture de l'Indus Tsangpo et les hauts sommets Himalayens où elles sont délimitées par la faille normale du Sud Tibet (STD) ainsi que sous forme de klippe plus au sud.

Au Népal, le Sub-Himalaya affleure sur plusieurs dizaines de km (~ 40 km à l'ouest Népal) et présentent des chevauchements internes comme le Main Dun Thrust, alors qu'il est beaucoup moins large au Bhoutan (< 10 km). À l'ouest de Katmandou ($\sim 82^\circ\text{E}$), les séries du LH sont beaucoup plus exposées. On observe aussi deux déflexions importantes du MCT vers le Nord au niveau du Sikkim et de l'Est Bhoutan impliquant des variations latérales de largeur du LH à l'affleurement. La présence du chevauchement du Kakhtang au Nord du MCT, qui permet de doubler l'épaisseur du HH au Bhoutan, est aussi une différence majeure par rapport au reste de l'arc.

D'un point de vue cinématique, *Najman et al. (1997)* montre que la chaîne ne se développe vraiment que depuis 28 Ma. L'utilisation des données radiochronologiques permet de mettre en évidence la séquence d'activité des différents chevauchements à l'origine de la formation du prisme Himalayen.

Dans l'ensemble de l'Himalaya, le fonctionnement du MCT débute après 23 Ma (*Hubbard 1989; Hodges et al. 1996; Daniel et al. 2003*). Au Bhoutan, les mouvements sur le

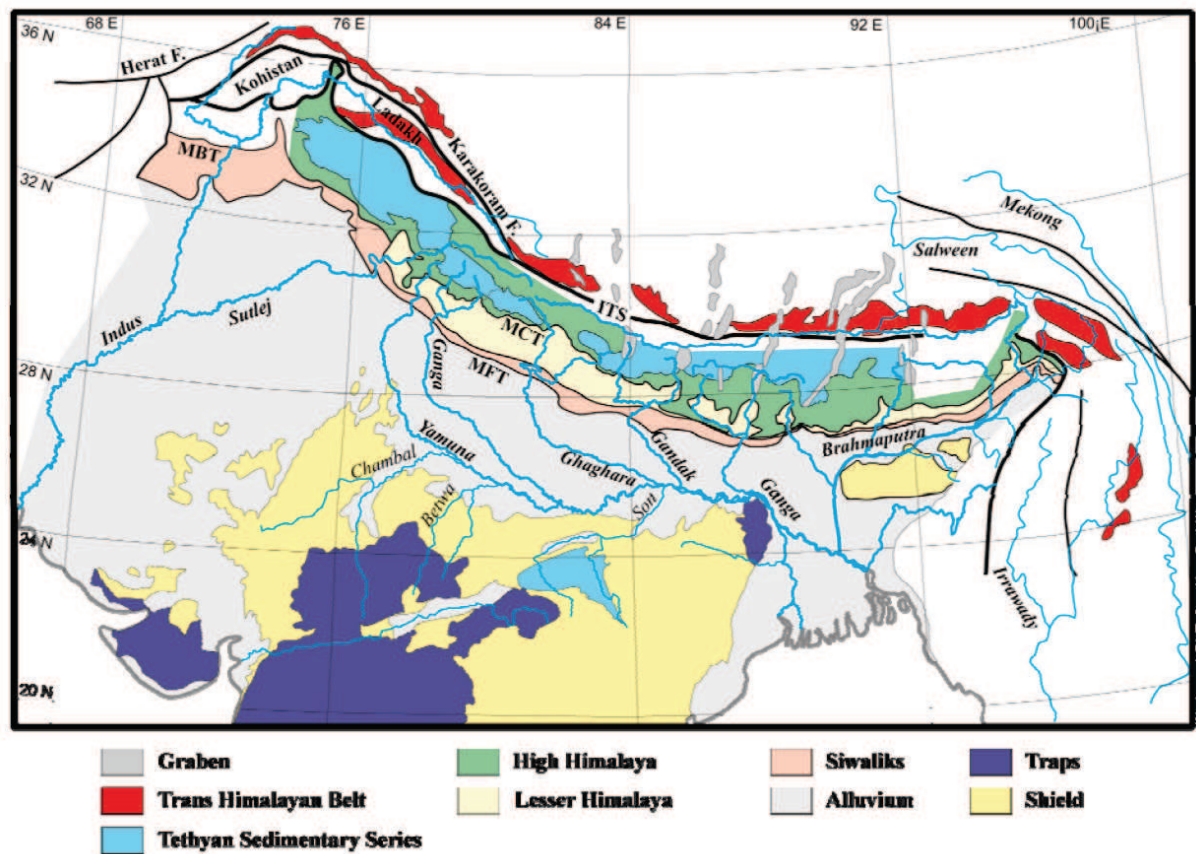


FIGURE 1.4 Carte géologique simplifiée de l'Himalaya et de l'Inde montrant les principales unités tectoniques, d'après *Galy* (1999).

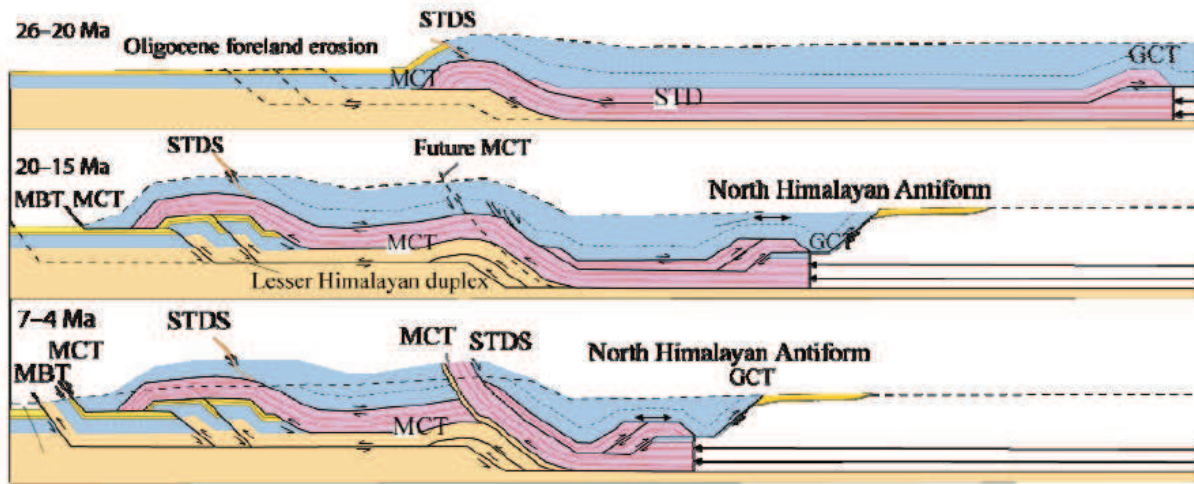


FIGURE 1.5 Modèle d'évolution cinématique du prisme Himalayen depuis 26 Ma illustrant sa formation incrémentielle, modifié d'après *Yin* (2006).

MCT sont datés entre 20 et 15 Ma (*Tobgay et al. 2012*) et son initiation semble être légèrement plus tardive à l'Est Bhoutan de ~ 3 Ma. Il faut noter l'activité hors séquence du chevauchement de Kakhtang à 15 Ma au Bhoutan (*Grujic et al. 2002; Daniel et al. 2003*) contemporain à la mise en place et à la formation du duplex inférieur dans le LH et au plissement des séries du HH. À partir de ~ 10 Ma, le duplex supérieur se met en place (*McQuarrie et al. 2008*). L'exhumation du LH dans l'Himalaya occidental et central est aussi datée à environ 10 Ma à partir d'analyses détritiques de sédiments du bassin d'avant pays (*Huyghe et al. 2001; Robinson et al. 2006; Najman et al. 2009*). Au Népal, le MFT aurait été initié aux environs de 2 Ma (*Mugnier et al. 2004*) et est actuellement actif comme le prouve la surrection des terrasses alluviales au front de la chaîne (*Lavé and Avouac 2000; Mugnier et al. 2004*) et les ruptures de surfaces observées (*Kumar et al. 2010; Sapkota et al. 2013*). Au Bhoutan, les séries Siwaliks sont beaucoup moins développées, suggérant une activation plus tardive du MFT.

Une synthèse cinématique est proposée sur la figure 1.5 (*Yin 2006*). La compilation des estimations de raccourcissement long terme à travers toute la longueur de la chaîne Himalayenne faite par *Long et al. (2011)* n'indique pas d'augmentation systématique vers l'Est comme pourrait le suggérer l'obliquité de la convergence entre l'Inde et l'Eurasie.

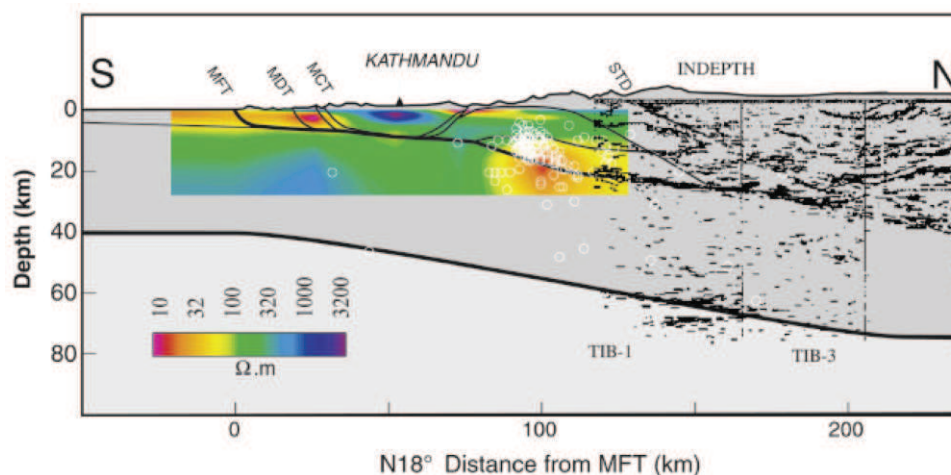


FIGURE 1.6 Coupe géologique construite à partir des contraintes géophysiques au Népal Central, d'après *Avouac (2003)*. Le profil de résistivité est obtenu à partir d'expériences magnétotelluriques (*Lemonnier et al. 1999*), les contraintes de sismiques par les expériences INDEPTH (*Zhao and Nelson 1993*). Les ronds blancs correspondent à la micro-sismicité enregistrée par le réseau sismologique Népalais, localisée au niveau de la zone de transition fragile-ductile sur le MHT.

1.3.3 Aspects géophysiques

Au cours des 30 dernières années, de nombreuses campagnes géophysiques (eg *Zhao and Nelson 1993*; *Lemonnier et al. 1999*; *Cattin et al. 2001*; *Nabelek et al. 2009*) ont permis d'imager les structures lithosphériques. Comme suggéré par les études géologiques (*Pêcher 1978*; *Brunel 1986*), les profils sismiques et l'analyse des fonctions récepteurs ont montré que les chevauchements majeurs du prisme Himalayen (MFT, MBT et MCT) se connectent en profondeur au Main Himalayan Thrust (MHT). Au Népal central, les données géodésiques suggèrent que ce chevauchement accommode l'essentiel de la déformation élastique intersismique dans sa partie bloquée (~100 km au nord du MFT) avant de s'enraciner au nord dans une zone de cisaillement ductile sub-horizontale. Cette transition de comportement mécanique est marquée par la localisation d'un essaim de micro-sismicité (*Pandey and Agrawal 1999*). À partir des profils sismiques profonds INDEPTH (*Zhao and Nelson 1993*; *Brown et al. 1996*; *Nelson et al. 1996*) et du profil de résistivité obtenu par *Lemonnier et al. (1999)*, *Avouac (2003)* propose une géométrie du MHT en plat-rampe-plat avec une rampe située sous le front du HH dont le pendage vers le nord est d'environ 15°(Fig. 1.6).

Les variations latérales de micro-sismicité au Népal semble être corrélées avec des variations latérales des structures géologiques avec des limites proches de ~82°E, ~85°E et

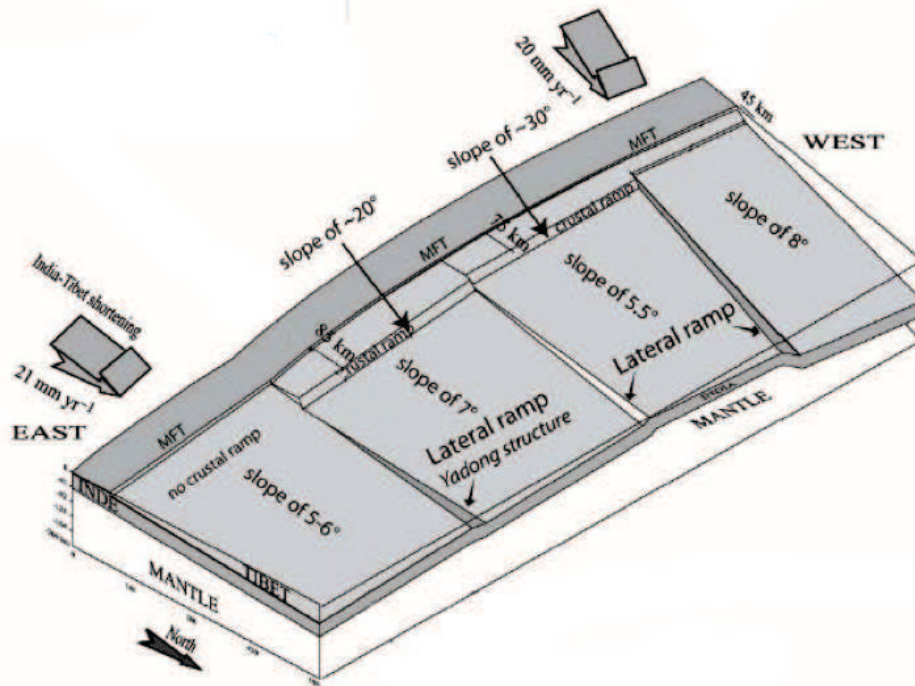


FIGURE 1.7 Modèle schématique de la segmentation du MHT entre l'ouest Népal et le Bhoutan contraint par la modélisation de la déformation intersismique (*Berger et al. 2004*) et la modélisation des données thermochronologiques (*Robert et al. 2011*), d'après *Robert et al. (2011)*.

~87°E. À partir de ces observations, *Berger et al. (2004)* propose une segmentation du MHT au niveau du Népal pour modéliser la déformation intersismique. Cette segmentation présente trois portions distinctes avec la présence d'une rampe mi-crustale prononcée dans les sections est et centre Népal et beaucoup moins à l'ouest Népal (Fig. 1.7). Le pendage moyen du MHT est aussi variable. Ce modèle est cohérent avec les modélisations thermiques basées sur des données thermochronologiques effectuées par *Robert et al. (2011)* qui complète ce modèle au niveau du Bhoutan en proposant un MHT plat sans rampe bien distincte. Cependant, les récentes modélisations au Bhoutan basées sur un jeu de données thermochronologique plus complet, nécessitent la présence d'une rampe crustale et même l'activité d'un chevauchement hors séquence (*Adlakha et al. 2013*).

La segmentation de la géométrie du MHT nécessite l'existence de rampes latérales. À partir des différentes observations et modélisations évoquées dans le paragraphe précédent, *Mugnier et al. (2013)* propose des limites à ~82°E (Jumla-Nepalgunj transfert zone), à ~85°E (Trisouli transfert zone) et ~89°E (Yadong transfert zone). Cependant si les observations topographiques et géologiques montrent des évidences nettes de segmentation

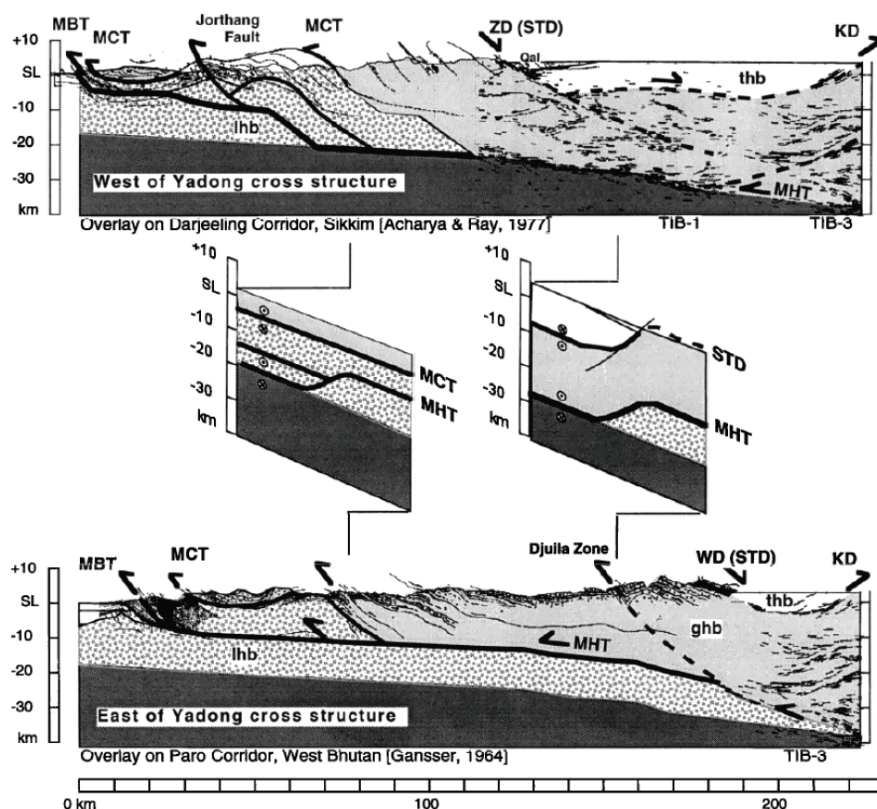


FIGURE 1.8 Coupes géologiques (à partir de *Acharaya and Ray* (1977) et *Gansser* (1964)) et contraintes sismiques de part et d'autre de la Yadong cross structure ($\sim 89^\circ\text{E}$) d'après *Hauck et al.* (1998).

la long de l'arc, ni les données intersismiques ni les données thermochronologique ne permettent de contraindre de manière précise la géométrie du MHT. En utilisant les profils sismiques INDEPTH et les coupes géologiques de *Acharaya and Ray* (1977) et *Gansser* (1964), *Hauck et al.* (1998) proposent une interprétation de la variation latérale à $\sim 89^\circ\text{E}$. La géométrie plane du MHT sous l'Himalaya du Bhoutan n'est cependant pas contrainte par les données sismiques (Fig. 1.8).

La relation entre la géométrie du MHT et celle de la plaque Indienne sous l'Himalaya n'est pas directe. En effet la géométrie du MHT dépend non seulement du comportement flexural de la plaque Indienne mais aussi des structures superficielles de la plaque Indienne accrétées et sous plaquées au niveau du prisme Himalayen. Malgré les nombreuses expériences géophysiques en Himalaya permettant d'imager les structures profondes et notamment la géométrie du Moho (INDEPTH, *Nelson et al.* (1996); HIMNT, *Schulte-Pelkum et al.* (2005); Hi-CLIMB; *Nabelek et al.* (2009)), il n'existe pas d'évidence de segmentation de la géométrie de la plaque Indienne le long de l'arc Himalayen entre le

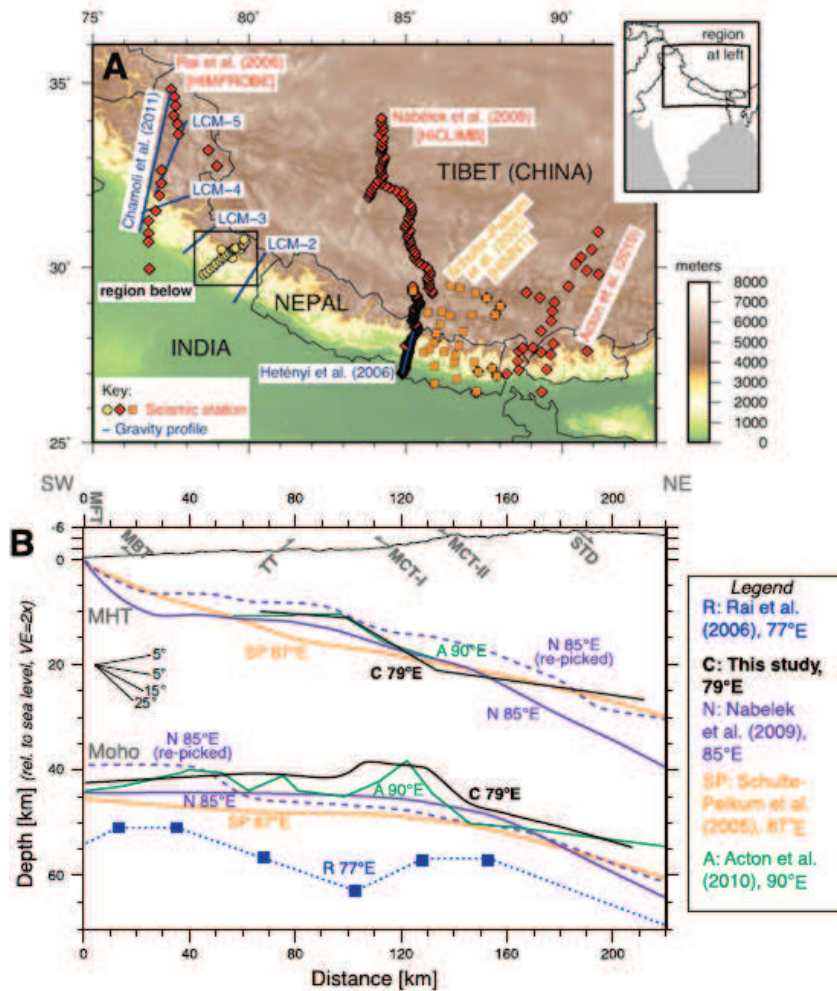


FIGURE 1.9 (A) Carte présentant la localisation des expériences sismologiques et des profils gravimétriques déjà étudiés en Himalaya. (B) Comparaison des contraintes obtenues sur le MHT et le Moho par la méthode des fonctions récepteurs, d'après *Caldwell et al. (2013)*

Népal et le Bhoutan (Fig. 1.9). Ceci est en partie dû au fait que certaines d'entre elles proviennent de résultats moyennés latéralement et que ces études n'utilisent pas le même modèle de vitesse pour obtenir la profondeur des interfaces.

1.3.4 Bilan et démarche suivie au cours de la thèse

L'ensemble des observations précédentes, synthétisées dans le tableau 1.1, montre clairement que l'hypothèse de cylindricité de l'arc Himalayen est trop simpliste. Ces observations sont cependant relativement parcellaires et une analyse systématique des variations latérales est indispensable. Pour ce faire, cette thèse s'articulera autour de deux parties :

- La première partie est consacrée aux données gravimétriques acquises au cours de cette thèse entre l'ouest Népal et le Bhoutan et à la mise en évidence de variations latérales structurales majeures dans cette zone. Dans le chapitre 2, je présenterai l'acquisition et le traitement des données gravimétriques. Ces données gravimétriques sont ensuite utilisées pour contraindre les modèles thermomécaniques de flexure de la plaque indienne sous le Népal (chapitre 3) et sous le Bhoutan (chapitre 4). Ces deux derniers chapitres sont présentés sous la forme d'articles publiés.
- La deuxième partie de la thèse sera consacrée à l'étude des déformations récentes du prisme Himalayen. Après avoir présenté au chapitre 5 l'état actuel des connaissances dans ce domaine, je montrerai l'évolution de l'état de contrainte du MHT entre le Népal et la Bhoutan sur le dernier millénaire (chapitre 6). Dans le chapitre 7, je présenterai les premières contraintes sur la vitesse de déformation Holocène ainsi que les premières évidences d'occurrence de séismes majeurs au Bhoutan. Ces deux derniers chapitres sont présentés sous la forme d'articles soumis.
- Enfin, après une conclusion reprenant l'essentiel des résultats obtenus, je présenterai plusieurs perspectives dans le cadre de notre problématique générale et les travaux en cours sur l'utilisation des données gravimétriques satellitaires GOCE.

TABLE 1.1 Synthèse des observations de variations latérales en Himalaya.

Observations	Variations latérales			Références
	Non	Oui	?	
Topographie	<input type="checkbox"/>	<input checked="" type="checkbox"/>	<input type="checkbox"/>	(eg <i>Duncan et al. 2003</i>)
Précipitations	<input type="checkbox"/>	<input checked="" type="checkbox"/>	<input type="checkbox"/>	(eg <i>Bookhagen and Burbank 2006</i>)
Érosion	<input type="checkbox"/>	<input checked="" type="checkbox"/>	<input checked="" type="checkbox"/>	(eg <i>Godard et al. 2013</i>)
Largeur unité structurale	<input type="checkbox"/>	<input checked="" type="checkbox"/>	<input type="checkbox"/>	(eg <i>Long et al. 2011</i>)
Cinématique des failles	<input type="checkbox"/>	<input checked="" type="checkbox"/>	<input checked="" type="checkbox"/>	(eg <i>McQuarrie et al. 2008</i>)
Géométrie du MHT	<input type="checkbox"/>	<input checked="" type="checkbox"/>	<input checked="" type="checkbox"/>	(eg <i>Robert et al. 2011</i>)
Géométrie du Moho	<input type="checkbox"/>	<input checked="" type="checkbox"/>	<input checked="" type="checkbox"/>	Chapitres 3 et 4
Déformations récentes	<input type="checkbox"/>	<input type="checkbox"/>	<input checked="" type="checkbox"/>	Chapitres 5, 6 et 7

Première partie

Apports de la gravimétrie à l'étude des
variations latérales de flexure de la
plaque indienne entre le Népal et le
Bhoutan

CHAPITRE 2

CAMPAGNES D'ACQUISITION GRAVIMÉTRIQUE EN HIMALAYA

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2.1 Introduction

La gravimétrie consiste en la mesure et l'analyse des variations dans l'espace et dans le temps du champ de pesanteur terrestre. Ses applications sont nombreuses :

- L'étude du champ de pesanteur permet d'étudier la forme de la Terre et est donc étroitement liée à la géodésie (e.g. *Pavlis et al. 2012*).
- L'analyse des variations temporelles de la pesanteur permet d'étudier les changements de la répartition des masses dans la Terre et trouve des applications dans des domaines divers tels que l'hydrogéologie (e.g. *Deville et al. 2012*), la volcanologie (e.g. *Jousset et al. 2000*), la géodynamique (e.g. *Masson et al. 2012*) et même la sismologie (e.g. *Panet et al. 2007*).
- Les variations spatiales de la pesanteur permettant d'appréhender la structure interne de la Terre de l'échelle de la subsurface (e.g. *Fournier et al. 2004*) jusqu'à celle du noyau terrestre (e.g. *Mandea et al. 2012*). Elle permet aussi de caractériser le comportement mécanique de la lithosphère (*Karner and Watts 1983*).

La pesanteur varie à la surface de la Terre du fait de sa forme ellipsoïdale, de l'altitude de la mesure et des hétérogénéités de masse dans le sous-sol. Notre objectif étant d'étudier la structure et le comportement mécanique de la lithosphère en Himalaya, seules les variations de masse sous la surface terrestre nous intéressent. Des corrections doivent donc être apportées aux mesures de pesanteur pour obtenir des anomalies gravimétriques que l'on pourra interpréter en termes de structures lithosphériques. Ces corrections sont présentées dans la section 2.2.

La compilation des anomalies de Bouguer publiées ou publiques dans la région de l'Himalaya (*Hetényi 2007*) montre une variation d'environ 500 mGal de l'Inde au Sud Tibet (Fig. 2.1). Cette variation reflète principalement l'épaississement crustal sous l'Himalaya et le Tibet et permet d'étudier le comportement flexural de la plaque indienne sous le système Himalaya-Tibet (*Cattin et al. 2001*; *Hetényi et al. 2007*). L'un des objectifs de ce travail de thèse est d'acquérir de nouvelles données gravimétriques avec une précision de l'ordre du mGal afin de couvrir les lacunes actuelles au centre/ouest Népal et au Bhoutan. Ces nouvelles données permettront d'étudier la variabilité latérale des structures lithosphériques et le comportement mécanique de la lithosphère grâce à un jeu de données géophysiques cohérent le long de l'arc Himalayen. La méthodologie d'acquisition ainsi que les différentes campagnes gravimétriques effectuées entre 2010 et 2012 au Népal

et au Bhoutan seront présentées dans la section 2.3. Enfin le traitement des données sera exposé dans la section 2.4.

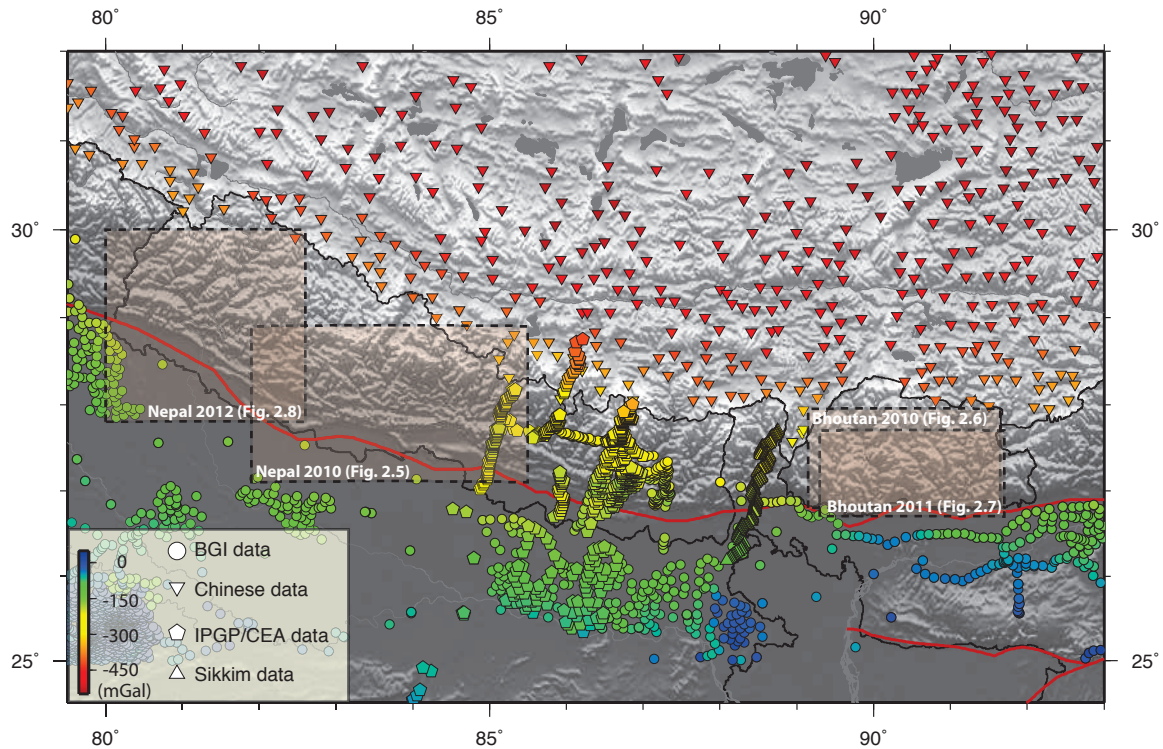


FIGURE 2.1 Carte des anomalies de Bouguer disponibles avant les campagnes gravimétriques effectuées au cours de cette thèse. Chaque figuré correspond à une source de données différentes (voir légende). Les encarts rouges correspondent aux figures 2.5, 2.6, 2.7 et 2.8, chacune associée à une campagne gravimétrique effectuée durant cette thèse.

2.2 De la mesure de la pesanteur à l'anomalie de Bouguer

2.2.1 Mesure de la pesanteur

Les mesures de l'accélération de la pesanteur consistent en la mesure simultanée d'une distance et d'un temps. Deux types de mesures sont possibles : les mesures absolues et les mesures relatives.

Les mesures absolues de la pesanteur permettent d'accéder directement à la valeur du champ de pesanteur terrestre. Si les premières mesures réalisées l'ont été à l'aide de

pendules (eg *Bessel 1830*), la méthode couramment utilisée aujourd'hui est celle de la chute libre d'un corps. Les gravimètres absolus portables tels que le *FG5* et le *A-10* de *Micro-g Solutions* permettent d'obtenir une précision de l'ordre du μGal ($1 Gal = 10^{-2} m.s^{-2}$).

Les gravimètres relatifs permettent de mesurer uniquement une variation de la valeur de l'accélération de la pesanteur entre deux points. La connaissance des valeurs de la pesanteur à partir de mesures relatives est donc dépendante de la connaissance de la valeur absolue en un point (base gravimétrique). Il existe plusieurs ordres de bases gravimétriques : les bases de premier ordre où des mesures absolues ont été réalisées, les bases de second ordre qui ont été rattachées aux premières grâce à des mesures relatives et ainsi de suite.

Les gravimètres relatifs *CG5* de *Scintrex* permettent une précision de mesure inférieure à 10^{-2} mGal. Ils utilisent un ressort vertical en quartz au bout duquel est fixé une masse soumise aux variations de pesanteur et placée dans un système capacitif. La tension électrique nécessaire à la compensation des variations de position de la masse est convertie en variations de pesanteur. Cette conversion nécessite au préalable l'étalonnage de ce signal électrique par rapport à des variations de pesanteur connues. Au cours de la mesure, deux corrections internes au gravimètre sont effectuées, l'une liée à la température et l'autre à l'horizontalité de l'instrument. De par les caractéristiques du dispositif, la mesure est soumise à une dérive instrumentale. La valeur de la dérive est variable selon les instruments mais aussi selon leurs conditions d'utilisation. Elle peut généralement être exprimée sous la forme d'une fonction affine sur un jour et par un polynôme de degré 2 à plus long terme. Il apparaît donc nécessaire de répéter les mesures en un (ou plusieurs) même points lors d'une série de mesures afin de corriger cette dérive.

Les valeurs mesurées doivent ensuite être corrigées des variations temporelles du champ de pesanteur qui sont liées à quatre raisons principales : les marées terrestres, les marées océaniques, les variations de pression atmosphérique et les mouvements du pôle de rotation terrestre. La précision visée pour nos campagnes de mesure étant de l'ordre du mGal, nous nous affranchissons des variations liées à la pression atmosphérique ($0,3 \mu Gal/hPa$) et au mouvement du pôle de rotation terrestre ($\sim \mu Gal/semaine$) à cause de leurs faibles amplitudes. Les marées terrestres et océaniques sont dues aux attractions directes de la Lune et du Soleil et aux déformations associées. L'amplitude maximale de la marée terrestre est de l'ordre de $0,3 mGal$ alors que celle des marées océaniques est plus faible, d'autant plus si l'on se trouve à distance des masses océaniques comme pour notre cas

d'étude ($\sim 10^{-2} mGal$).

2.2.2 Valeur théorique de la pesanteur

La forme de la Terre est définie par une surface équipotentielle du champ de pesanteur qui se confond avec la surface moyenne des océans au repos : le géoïde. La forme du géoïde dépend de la répartition des masses à l'intérieur de la Terre et l'ellipsoïde de référence correspond à la position d'une surface équipotentielle du champ de pesanteur théorique qui s'en rapproche au mieux. La valeur théorique de la pesanteur calculée sur cet ellipsoïde ne dépend que de la latitude :

$$g_{theorique} = g_0(1 + k_1 \sin^2 \lambda - k_2 \sin^2 2\lambda) \quad (2.1)$$

avec g_0 la pesanteur à l'équateur et k_1 et k_2 des constantes dépendantes de la forme et de la vitesse de rotation de la Terre et λ la latitude. Du fait de l'aplatissement de la Terre aux pôles et de sa rotation sur elle même, la pesanteur théorique calculée sur l'ellipsoïde de référence varie de près de 5000 mGal entre l'équateur et le pôle. C'est pourquoi il est nécessaire de prendre en compte ces variations latitudinales sur les analyses des mesures gravimétriques effectuées.

2.2.3 Corrections et anomalies gravimétriques

La comparaison des valeurs mesurées (g_{mesure}) et théoriques ($g_{theorique}$) du champ de pesanteur n'est pas immédiate et nécessite des corrections. Il est en effet indispensable de tenir compte de la distance entre les surfaces où l'on calcule la valeur théorique (surface de référence) et celle où l'on mesure, ainsi que de corriger de la présence de corps massiques entre ces surfaces (Fig. 2.2).

La correction à l'air libre permet de ramener la surface de référence à l'altitude du point mesuré. Pour ce calcul, on néglige l'effet des masses situées entre le point de mesure et le niveau de référence. Le gradient de la pesanteur est de $0,3086 mGal.m^{-1}$. L'anomalie à l'air libre est donc obtenue avec l'équation suivante :

$$\Delta g_{AL} = g_{mesure} - (g_{theorique} - 0,3086z) \quad (2.2)$$

avec Δg_{AL} en $mGal$ et z l'altitude en m, comptée positive si le point de mesure se situe au dessus du niveau de référence (ce qui est toujours le cas dans notre situation).

Il faut maintenant tenir compte de l'effet gravitationnel des masses excédentaires situées entre le niveau de référence et le point de mesure. Cela se fait en deux étapes permettant d'obtenir l'anomalie de Bouguer simple et l'anomalie de Bouguer complète. Pour la première on considère un relief tabulaire d'épaisseur z (en m) et de densité ρ dont la perturbation en $mGal$ équivaut à $0,0419\rho z$. La seconde permet de prendre en compte les variations de pesanteur induites par la présence de reliefs à proximité du point de mesure et correspond à la correction de terrain (δ_{topo}). L'anomalie de Bouguer complète s'obtient donc avec l'équation suivante :

$$\Delta g_{BC} = g_{mesure} - (g_{theorique} - 0,3086z + 0,0419\rho z - \delta_{topo}) \quad (2.3)$$

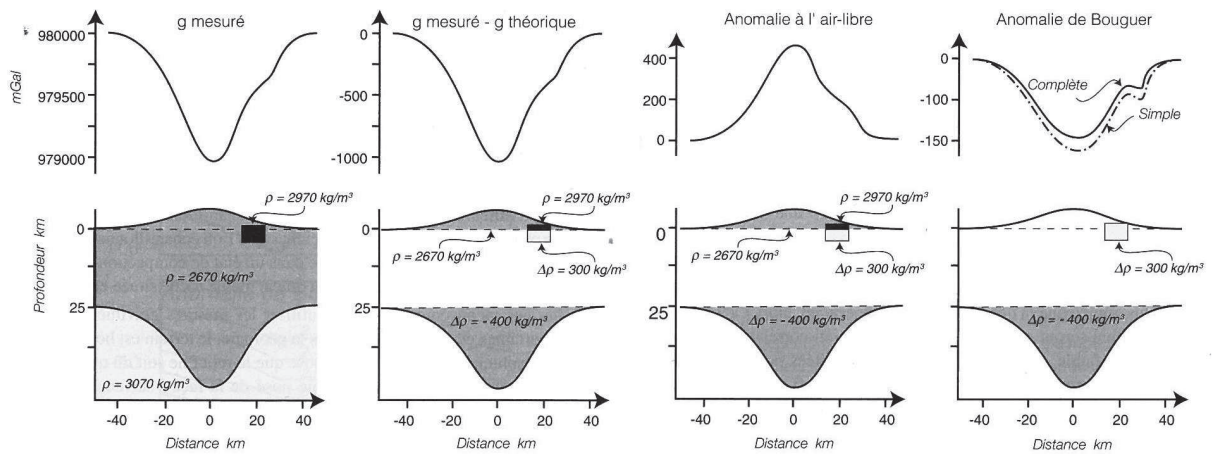


FIGURE 2.2 Schémas et signaux associés aux différentes étapes de correction à apporter à la mesure de pesanteur pour obtenir l'anomalie de Bouguer, d'après *Dubois et al. (2011)*

Le calcul de l'anomalie de Bouguer complète est donc dépendante de l'altitude du point de mesure et de la densité utilisée. Avec la densité traditionnellement utilisée de 2,67 (*Hinze 2003*), les corrections à l'air libre et de plateau correspondent à un gradient de $0,1967 mGal.m^{-1}$. Une incertitude sur la position verticale des points de mesure de $\pm 5 m$ correspond donc à une incertitude de l'anomalie de Bouguer d'environ $\pm 1 mGal$. Comme la précision recherchée dans le cadre des campagnes effectuées est de l'ordre du $mGal$, cela nécessite une précision métrique sur le positionnement vertical des points de mesures. Pour ce faire nous utilisons des systèmes de positionnement par satellites.

2.2.4 Système de positionnement par satellite

Le *Global Positioning System* (GPS) permet de déterminer la position précise d'un récepteur GPS situé à la surface du globe grâce à la mesure des distances entre ce dernier et les satellites GPS. Trois satellites sont nécessaires à l'obtention d'un point d'intersection des trois sphères (dont le rayon correspond à la distance récepteur-satellite) ainsi qu'un quatrième pour la synchronisation des horloges. La constellation des 24 satellites est telle que l'on peut observer en tout points du globe, et à tout instant, 4 satellites au minimum. Chaque satellite émet deux ondes porteuses L1 et L2, modulées par des codes pseudo aléatoires *Coarse Acquisition* sur L1 et *Precision* sur L1 et L2. Le temps mis par ces signaux pour atteindre le récepteur permet de calculer la distance qui sépare les satellites et le récepteur. Elle peut être obtenue soit par la corrélation entre le signal reçu et le signal propre du récepteur sur les codes (pseudo-distance), soit par la mesure du déphasage des porteuses L1 et/ou L2. Il existe divers modes de positionnement qui diffèrent par la précision obtenue et qui nécessitent différents dispositifs d'acquisition. Nous présentons ci-après les trois modes utilisés pour le traitement des données.

1. **Le positionnement absolu :** Il est possible d'obtenir une position absolue en temps réel avec des récepteurs de faible coût en mesurant (au minimum) quatre pseudo-distances. Cependant ce type de positionnement est peu précis (plurimétrique).
2. **Le positionnement relatif par mesure de phases :** Il permet d'obtenir une précision millimétrique dans des conditions optimales (petites lignes de base et enregistrements continus sur 24h). Le positionnement relatif consiste en la détermination d'un vecteur (ligne de base) reliant deux récepteurs dont un est de position connue (*Herring et al. 2006*). L'utilisation simultanée des données provenant de plusieurs satellites à plusieurs stations permet, par double différence, de diminuer fortement l'influence des erreurs d'orbites, des corrections de propagation et des décalages des horloges satellites et récepteurs. Cette méthode de positionnement nécessite l'enregistrement simultané de plusieurs récepteurs GPS.
3. **Le positionnement pontuel précis (PPP) :** Cette méthode permet un positionnement absolu avec une précision approchant celle du positionnement relatif par double différence (*Kouba and Heroux 2001*). Pour atteindre ces précisions, cette technique utilise des modèles des différentes sources de perturbation de la mesure

GPS (orbites, horloge des satellites, paramètres de rotation de la Terre, surcharge océaniques, variations du centre de phase des antennes, *etc*). Cette méthode de positionnement nécessite un seul récepteur GPS contrairement au positionnement relatif.



FIGURE 2.3 Instruments utilisés et méthodologie d'acquisition lors des campagnes gravimétriques. (A) Photographie de l'ensemble du matériel utilisé lors des missions de terrain. (B) Estimation de la hauteur de l'antenne GPS par rapport au gravimètre. (C) Point de mesure gravimétrique au Bhoutan. Les parapluies permettent de protéger les gravimètres d'une exposition directe aux rayons solaires.

2.3 Campagnes d'acquisition gravimétrique

2.3.1 Instruments utilisés et méthodologie d'acquisition

Au cours de ma thèse, j'ai participé à l'ensemble des campagnes d'acquisition gravimétrique au centre/ouest Népal et au Bhoutan. Les mesures gravimétriques ont été réalisées le long de routes carrossables en densifiant les mesures le long des profils perpendiculaires à la chaîne. L'utilisation de gravimètres relatifs *CG5* a nécessité la connexion de nos mesures avec des bases absolues. La dérive instrumentale des *CG5* (cf. section 2.2) nous a contraint à boucler les mesures soit en remesurant le premier point de mesure en fin de journée (Fig. 2.4A), soit en effectuant les mesures entre deux points de valeurs connues (Fig. 2.4B). Etant données les grandes distances à couvrir et la précision recherchée, nous n'avons pas réoccupé de points intermédiaires à l'intérieur d'une même boucle. Nous avons effectué nos boucles sur une seule journée de terrain afin de pouvoir considérer la dérive comme une fonction affine. Avant toute mesure sur le terrain, les *CG5*

sont mis en chauffe au minimum 48h afin de stabiliser la température interne du gravimètre. Les boucles journalières sont parcourues en voiture. Au minimum 5 mesures de 90 s sont réalisées à chaque site. Ces mesures sont considérées stables si la variation entre les différentes valeurs mesurées n'excède pas $\pm 10\mu Gal$. Les mesures sont faites à l'ombre ou avec un parasol afin d'éviter une chauffe du gravimètre ou du trépied pouvant nuire à la stabilité de la mesure (Fig. 2.3C). Un positionnement absolu en temps réel du point de mesure est fait grâce à un récepteur GPS *GARMIN* à main (positionnement GPS de type 1). Cette localisation nous permet de répartir nos mesures régulièrement le long de nos profils. Nous utilisons en parallèle un système d'acquisition composé d'un récepteur *Ashtec Z-12 Sensor* et d'une antenne GPS *Trimble Zephyr* afin d'obtenir une précision métrique. Avant chaque mesure, nous vérifions que le nombre de satellites visibles par l'antenne GPS soit suffisant. Un minimum de 15 minutes d'enregistrement pour chaque point est effectué, avec une réception d'au moins 5 signaux issus de satellites différents. La différence de hauteur entre le centre de phase de l'antenne GPS et le point de mesure est mesurée ($\pm 5cm$) et utilisée pour corriger la position verticale finale du gravimètre (Fig. 2.3B).

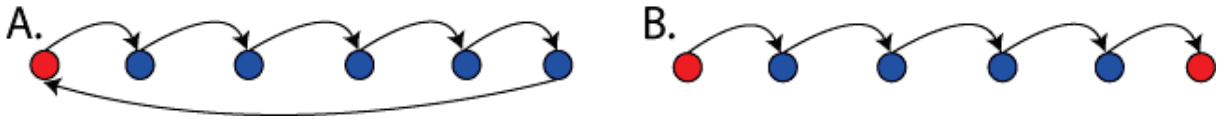


FIGURE 2.4 Schémas illustrant les deux principes d'acquisition permettant de s'affranchir de la dérive des gravimètres relatifs. Les points rouges représentent des bases gravimétriques où la valeur de pesanteur est connue et les points bleus, des points inconnues. (A) Boucle sur un seul point mesuré deux fois, en début et en fin de journée. Ce type de stratégie de mesure implique de faire l'aller-retour dans la journée. (B) Le premier point mesuré et le dernier ne sont pas identiques cependant leur valeur de pesanteur est connue. Cette stratégie permet d'effectuer un aller simple entre les deux points rouges.

2.3.2 Népal 2010

La première des 4 campagnes gravimétriques du projet Bhutanepal s'est déroulée du 23 Octobre au 4 Novembre 2010 au Népal. Une première équipe a effectué des mesures gravimétriques le long de la east-west motorway au sud du Népal entre Katmandou et Tulsipur. La seconde équipe a effectué des mesures le long d'un profil N/S au centre Népal le long de la route entre Siddhartanagar et Jomosom. Au total 18 boucles ont été réalisées

pour un total de 120 points mesurés. Afin de caler ce réseau en absolu, nous avons relié nos mesures à 4 bases absolues (Kathmandu, Simra, Siddharthanagar et Tansen) mesurées par le NOAA en 1996 (Fig. 2.5).

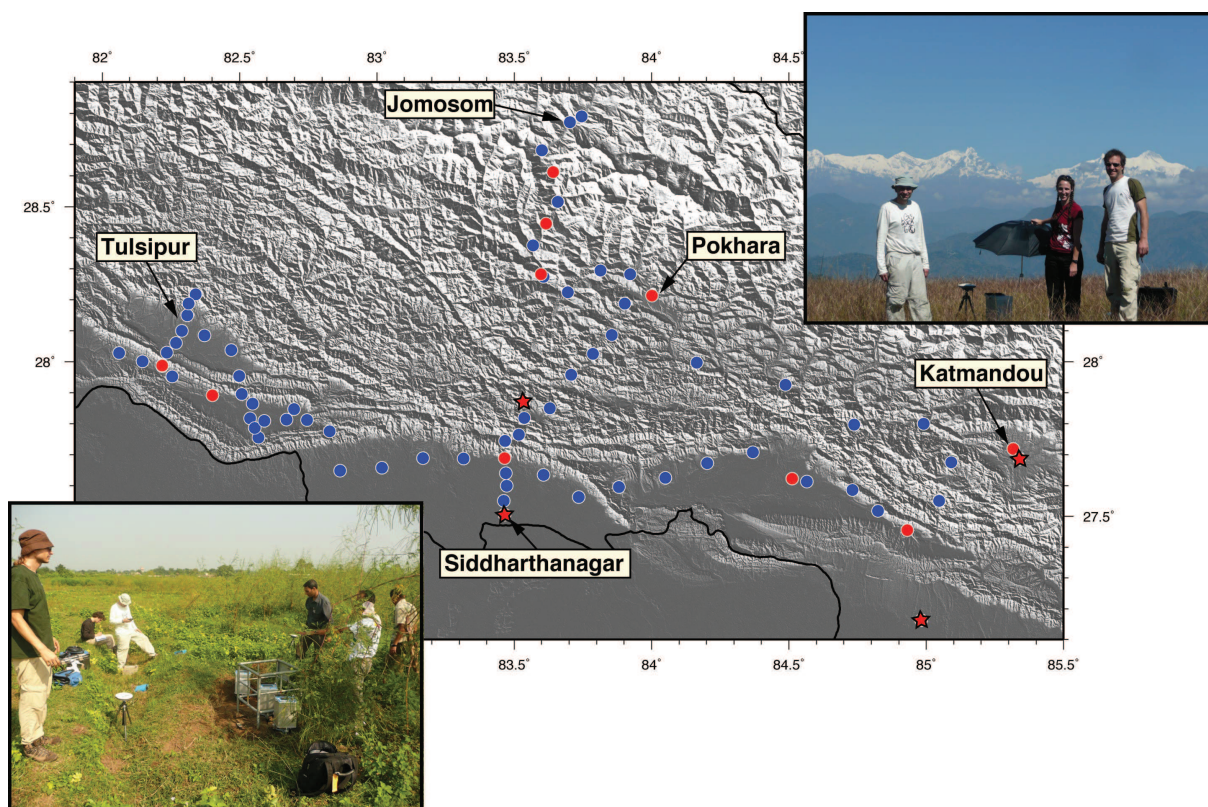


FIGURE 2.5 Carte topographique du centre Népal montrant les points mesurés lors de la campagne Népal 2010. Les points rouges correspondent aux bases gravimétriques occupées en début et en fin de boucle ; les étoiles rouges aux bases absolues utilisées pour caler notre réseau. En haut à droite, Rodolphe Cattin, Sarah Lechmann et Erik Doerflinger mesurant à la base gravimétrique de Tansen (base absolue mesurée par le NOAA), en arrière plan les Annapurnas. En bas à droite, Cédric Champollion, Rodolphe Cattin, Prithvi Shrestha et Théo Berthet effectuant une mesure gravimétrique au point absolu de Simra.

2.3.3 Bhoutan 2010

La campagne de 2010 au Bhoutan a permis d'effectuer les premières mesures gravimétriques dans ce pays. Elle s'est déroulée du 5 au 16 Novembre 2010. La première équipe a couvert la seule route E/O du pays située à $\sim 27,5^\circ\text{N}$, qui relie la capitale, Thimphou, à Trashigang. La seconde équipe a mesuré un profil N/S dans l'ouest du pays le long de la route entre Phuentsholing et Gasa. Les régions de Ha et de Paro ont aussi été couvertes pour un total de 179 mesures et 20 boucles. De par l'absence de mesures gravimétriques

absolues au Bhoutan, le nouveau réseau gravimétrique bhoutanais n'est actuellement calé en absolu qu'avec une précision de $\pm 10 \text{ mGal}$. L'établissement de plusieurs bases relatives dans le pays facilitera la connexion de notre réseau aux mesures absolues prévues en 2014.

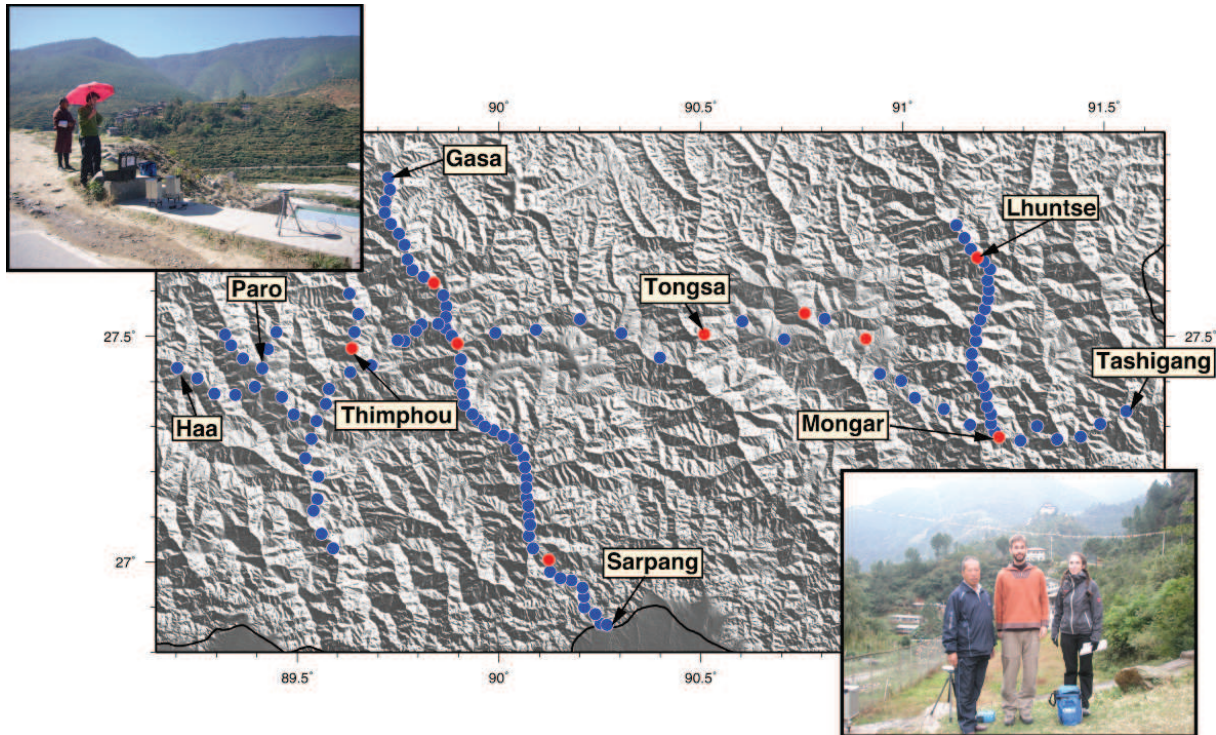


FIGURE 2.6 Carte topographique du Bhoutan montrant les points mesurés lors de la campagne Bhoutan 2010. Les points rouges correspondent aux bases gravimétriques occupées en début et en fin de boucle. En haut à gauche, mesures gravimétriques à Wangdu Phodrang. En bas à droite, Mr Wangchuk (notre chauffeur), György Hetényi et Sarah Lechmann à Lhuentse lors de la dernière mesure effectuée au cours de cette mission.

2.3.4 Bhoutan 2011

Cette mission, qui s'est déroulée entre le 12 Novembre et le 3 Décembre 2011, a permis de compléter le réseau de mesures gravimétriques bhoutanais débuté en 2010. Au cours de cette campagne, que j'ai co-organisé, 3 nouveaux profils N/S à l'est, au centre et à l'ouest Bhoutan ont été couverts. Au total, 102 mesures réparties dans 12 boucles ont été effectuées. Afin de connecter ces nouvelles mesures au réseau existant, nous avons réoccupé les bases gravimétriques établies à Trashigang, Tongsa et Thimphou lors de la précédente mission. Les mesures gravimétriques acquises au cours de ces deux missions ont permis de couvrir tout le réseau routier bhoutanais actuel.

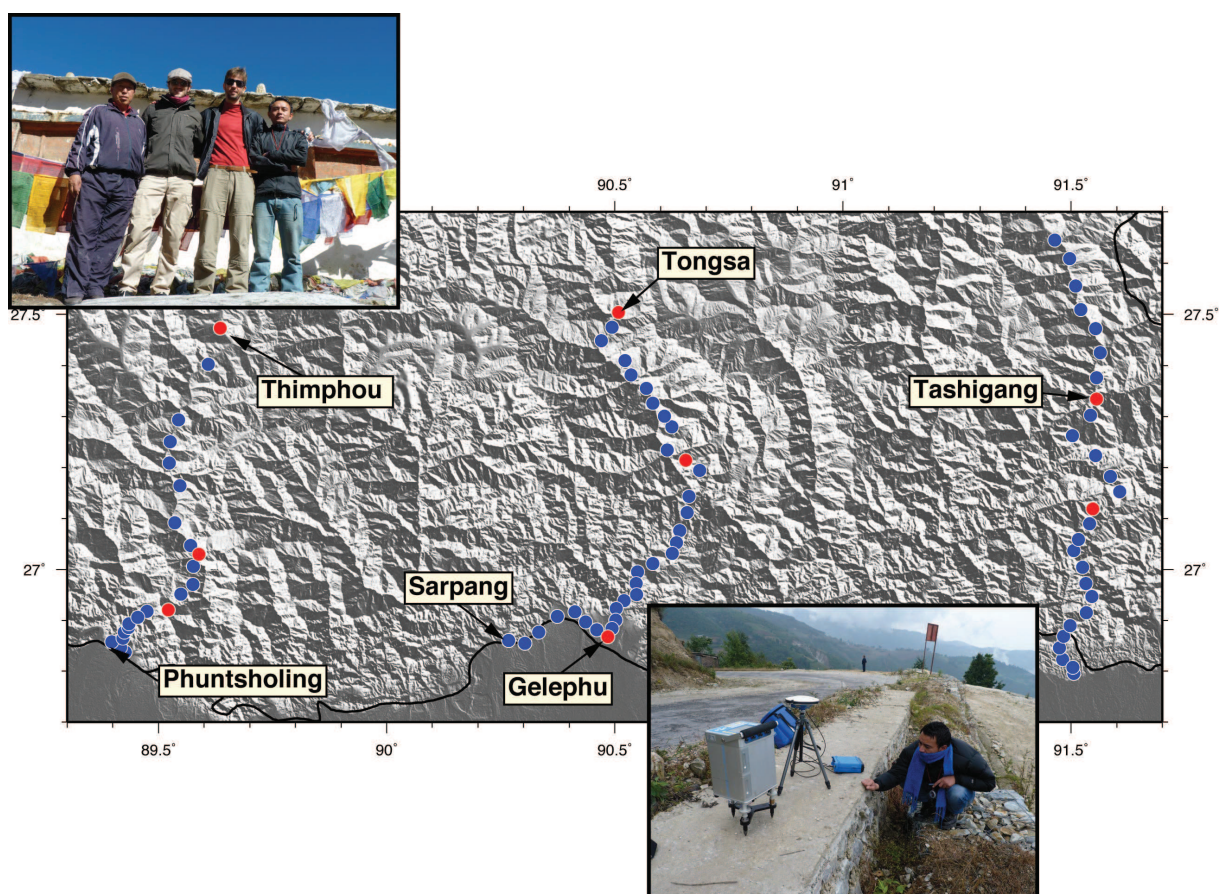


FIGURE 2.7 Carte topographique du Bhoutan montrant les points mesurés lors de la campagne Bhoutan 2011. Les points rouges correspondent aux bases gravimétriques occupées en début et en fin de boucle. En haut à gauche, composition de l'équipe ayant effectuée les mesures. De gauche à droite : Mr Wangchuk, Théo Berthet, György Hetényi et Jamyang Chophel. En bas à droite, Jamyang Chophel en train de vérifier le nombre de signaux satellites enregistrés par le récepteur GPS.

2.3.5 Népal 2012

Fort de l'expérience acquise au cours des précédentes missions, j'ai organisé et dirigé la mission gravimétrique de 2012 dans l'ouest Népal. Elle s'est déroulée du 14 Mars au 6 Avril 2012 afin d'étendre le réseau gravimétrique népalais à l'ouest du pays. Cette campagne a permis de terminer de couvrir la east-west motorway jusqu'à la frontière Indienne à l'ouest du pays en reliant ces nouvelles mesures aux bases relatives établies en 2010. Un nouveau profil N/S à l'ouest du pays a été mesuré le long de la route entre Danghadi et Darchula. Durant cette campagne, 104 mesures incluses dans 13 boucles ont été effectuées.

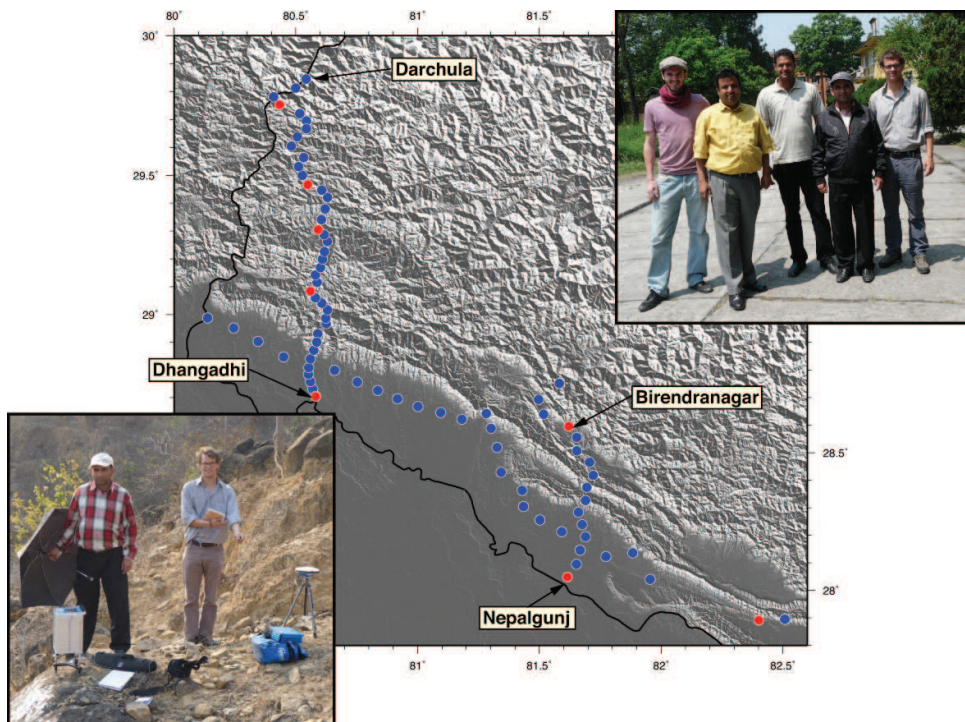


FIGURE 2.8 Carte topographique de l'Ouest Népal montrant les points mesurés lors de la campagne Népal 2012. Les points rouges correspondent aux bases gravimétriques occupées en début et en fin de boucle. En bas à gauche, dispositif standard d'un point de mesure gravimétrique avec le gravimètre (à gauche) et le récepteur GPS (à droite) à proximité. En haut à gauche, composition de l'équipe ayant effectué le mission Népal 2012. De gauche à droite : Théo Berthet, Dr. Som Nath Sapkota, Thakur Kandel, Lok Adikari et Paul Hammer.

2.4 Traitement des données et calcul des anomalies gravimétriques

2.4.1 Traitement GPS

Dans un premier temps, les données GPS enregistrées par le récepteur sont converties en fichier *RINEX* à l'aide du logiciel *teqc* développé par l'UNAVCO (University Navstar Consortium). Ces fichiers *RINEX* sont ensuite traités en utilisant soit la méthode de positionnement relatif par mesure de phases soit le positionnement ponctuel précis. Afin de comparer ces méthodes de positionnement, nous avons effectué ces deux types de traitement sur les données enregistrées lors de campagnes Népal 2010 et Bhoutan 2010. En ce qui concerne le positionnement relatif, 6 stations GPS fixes du réseau Népalais - géré par le DGM (*Department of Geology and Mines de Katmandou*) en collaboration avec le DASE/CEA (*Département Analyse, Surveillance, Environnement du Commissariat à l'Energie Atomique*) et le *California Institute of Technology* - ont été utilisées pour le traitement des points de la campagne Népal 2010 et 4 (dont la station hors réseau Népalais située à Lhasa) pour les points de la campagne Bhoutan 2010 (Fig 2.9). Ces données, associées aux orbites précises des satellites fournies par l'IGS (International Global Navigation Satellite System Service), ont été traitées avec le logiciel *Trimble Geomatic Office* (TGO). Pour ce qui est du traitement par positionnement ponctuel précis, nous avons utilisé le logiciel *NRCAN's PPP* développé par le *Canadian Department of Natural Resources* (*Kouba and Heroux 2001*).

La comparaison des hauteurs ellipsoïdales obtenues avec les logiciels *TGO* et *NRCAN's PPP* ne montre pas de biais systématique (Fig. 2.10). Si l'on met de côté les valeurs aberrantes montrant des différences supérieures à 2,5 m, l'histogramme des différences de hauteurs obtenues par ces deux types de traitement suit une distribution gaussienne centrée sur -0,03 m avec un écart type de 0,70 m. Les valeurs aberrantes peuvent être liées au traitement PPP et à l'enregistrement parfois discontinu du signal provenant des satellites.

Les erreurs RMS formelles fournies par le logiciel *NRCAN's PPP* et *TGO* sur le positionnement vertical sont représentées sur la figure 2.11. La moyenne des erreurs RMS calculée par *TGO* est de 0,18 m avec un écart type de 0,16 m alors qu'elle est de 0,62 m

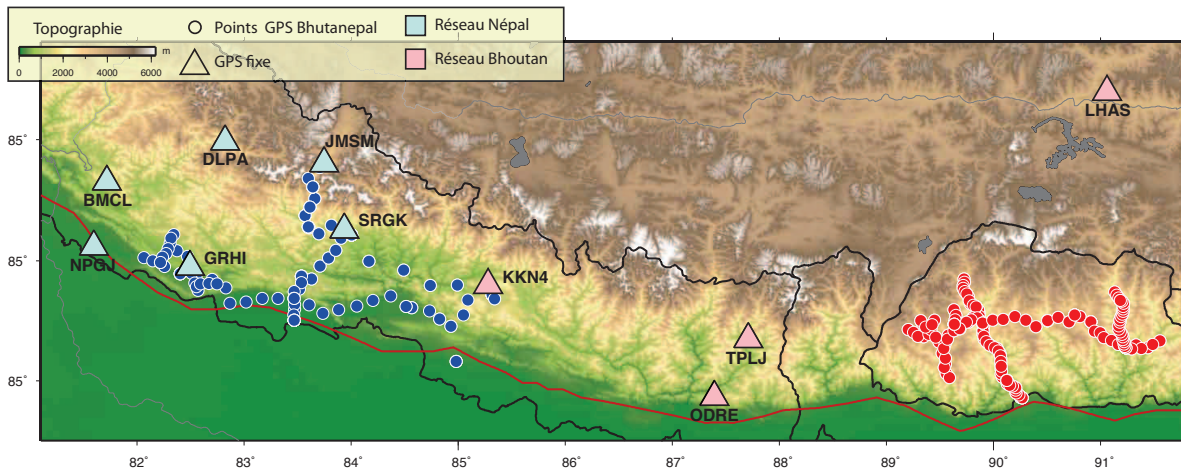


FIGURE 2.9 Carte de localisation des stations GPS fixes utilisées pour le positionnement relatif par mesure de phase. Les triangles bleus correspondent aux stations fixes utilisées pour la traitement des points de la campagne Népal 2010 (ronds bleus) et les rouges pour la campagne Bhoutan 2010.

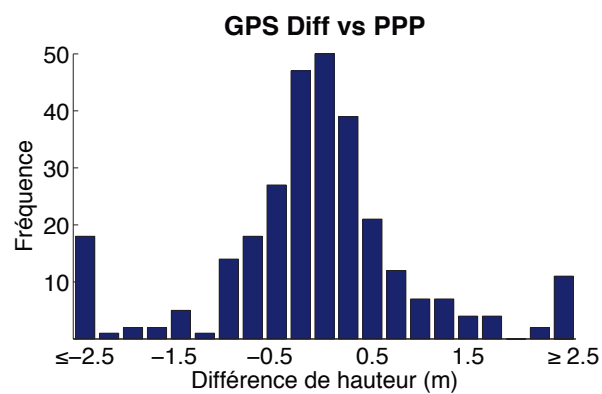


FIGURE 2.10 Histogramme de fréquence des différences de hauteur ellipsoïdales obtenues avec la méthode de positionnement ponctuel précis et le positionnement relatif.

avec un écart type de 0,81 m pour PPP. Cependant, le traitement des erreurs formelles étant propre à chaque logiciel, elles ne peuvent pas être comparées.

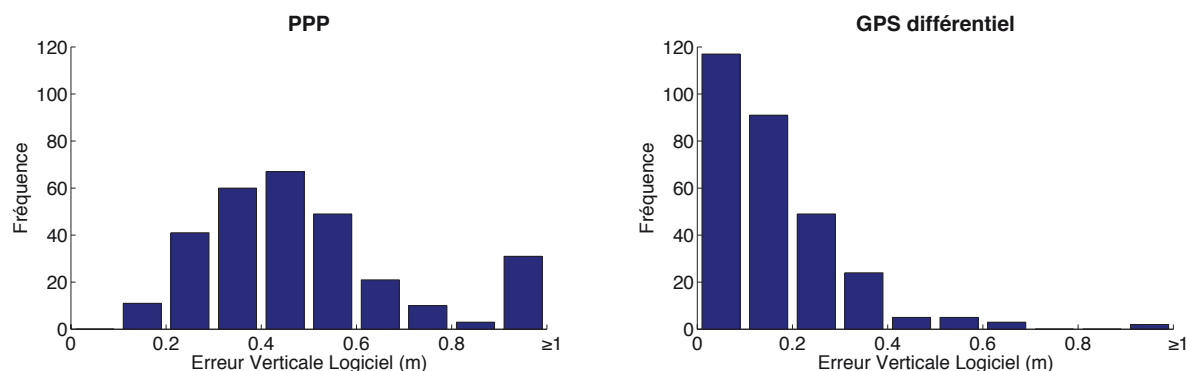


FIGURE 2.11 Histogrammes de fréquence des erreurs verticales fournies par les logiciels de post-traitement GPS. À gauche, les erreurs associées au positionnement ponctuel précis. À droite, les erreurs fournies par *Trimble Geomatic Office* et le positionnement relatif par mesure de phases.

Afin d'estimer une incertitude sur le positionnement vertical de nos points plus cohérente que les erreurs formelles fournies par les logiciels utilisés, nous tirons parti des mesures GPS répétées à certaines bases gravimétriques pour étudier la répétabilité propre à chaque méthode de positionnement. Nous avons comparé les hauteurs obtenues par les 3 méthodes de positionnement décrites dans le paragraphe 2.2.4 pour 7 bases gravimétriques, chacune mesurée au moins 4 fois. Ces différences de hauteur sont montrées dans la figure 2.12 et suggèrent une répétabilité d'environ $\pm 0,3$ m pour le positionnement relatif, ± 2 m pour le positionnement ponctuel précis et ± 10 m pour le positionnement absolu. Ces valeurs seront utilisées comme incertitudes sur le positionnement vertical lors de l'estimation des erreurs associées au calcul des anomalies gravimétriques.

De par sa meilleure répétabilité verticale, nous utilisons les données de positionnement obtenues par traitement relatif en double différence pour les campagnes Népal 2010 et Bhoutan 2010. Les données des stations GPS fixes du réseau Népalais étant indisponibles au moment du traitement des données, les positions des points mesurés lors des campagnes Bhoutan 2011 et Népal 2012 sont obtenues par le positionnement ponctuel précis sauf pour certains points où les données GPS enregistrées ne permettent pas ce type de post traitement. Dans ce cas nous utilisons les positions fournies par la méthode absolue.

Les traitements GPS effectués fournissent des hauteurs ellipsoïdales. La conversion de ces hauteurs ellipsoïdales en altitude permet de s'affranchir de l'anomalie gravimétrique

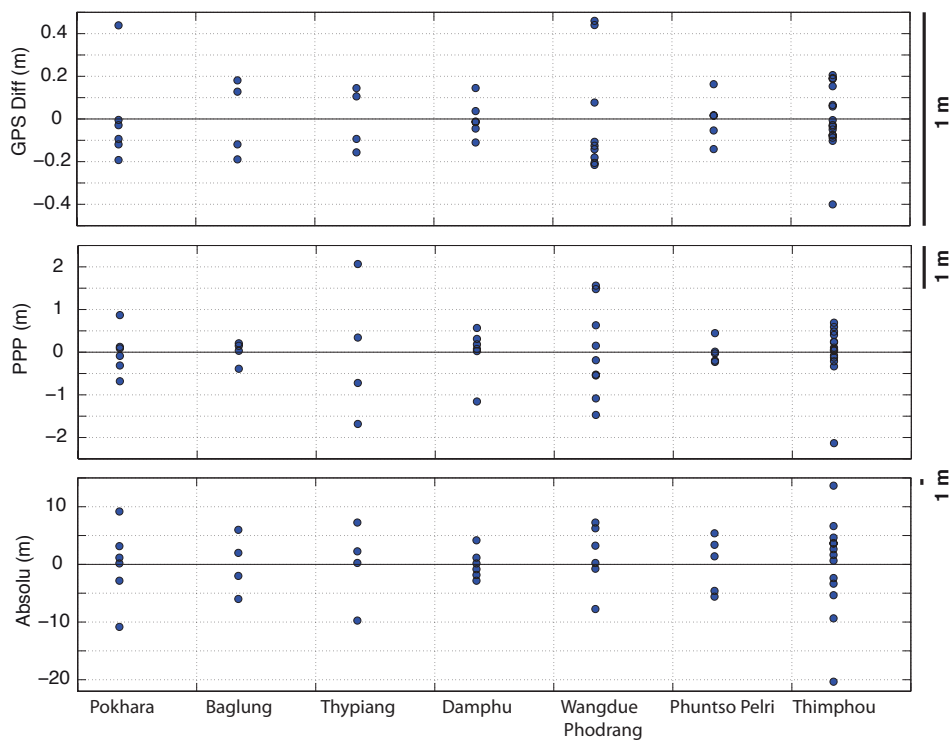


FIGURE 2.12 Répétabilité des mesures de positionnement vertical à 7 bases gravimétriques pour les trois types de positionnement GPS. En haut, répétabilité associée au positionnement relatif ($\pm 0,30 m$). Au milieu, répétabilité associée au positionnement ponctuel précis ($\pm 2 m$). En bas, répétabilité associée au positionnement absolu ($\pm 10 m$).

régionale à grande longueur d'onde responsable des ondulations du géoïde principalement causées par les distributions de masses en profondeur dans le manteau (*Ricard et al. 1988*). En Himalaya, le géoïde se situe au dessus de l'ellipsoïde (30-40 m) de référence WGS84, ce qui implique de soustraire la hauteur géoïdale de la hauteur ellipsoïdale pour obtenir l'altitude (Fig. 2.13). Le logiciel *hsynthWGS84* fournie par la NGA (National Geospatial-Intelligence Agency) a été utilisé afin de convertir les hauteurs ellipsoïdales en altitudes par rapport au modèle de géoïde EGM08 dont l'incertitude est de $\pm 0,45m$ (*Pavlis et al. 2012*).

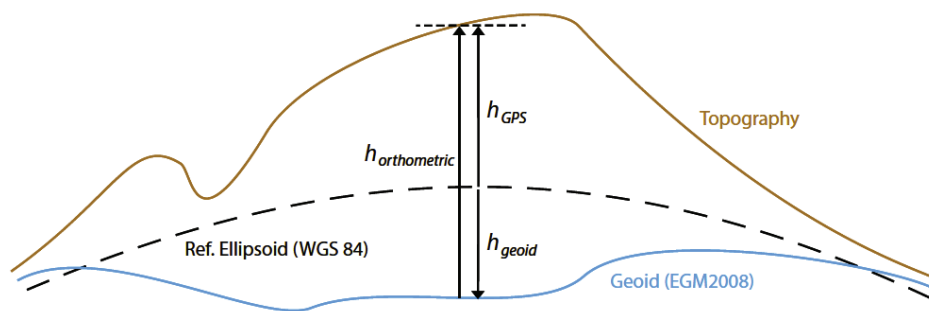


FIGURE 2.13 Schéma illustrant la différence entre la hauteur ellipsoïdale (h_{GPS}) provenant du positionnement GPS et l'altitude ($h_{orthometric}$) qui correspond approximativement au niveau de la mer. La différence entre ces deux distances correspond à la hauteur géoïdale (h_{geoid}). Le modèle EGM08 est utilisé dans notre étude.

2.4.2 Corrections de la pesanteur mesurée

Les mesures de pesanteur effectuées durant les campagnes gravimétriques sont tout d'abord corrigées des effets liés aux marées terrestres et océaniques. De par la dispersion spatiale et temporelle des points de mesures, nous avons mis en place une routine fortran permettant de calculer en tout points de l'espace et du temps l'effet des marées terrestres et océaniques. Pour ce faire, nous utilisons le package ETERNA (*Wenzel 1996*) afin de calculer l'effet des marées terrestres. La surcharge océanique, bien que très faible dans notre cas (quelques μGal), est calculée en utilisant le modèle de marée FES2004 dont les coefficients sont disponibles en ligne¹. La comparaison des corrections de marées obtenues par notre méthode à celles calculées avec le logiciel *TSoft* (*Van Camp and Vauterin 2005*)

1. <http://holt.oso.chalmers.se/loading/>

montre des différences inférieures au μGal (Fig. 2.14).

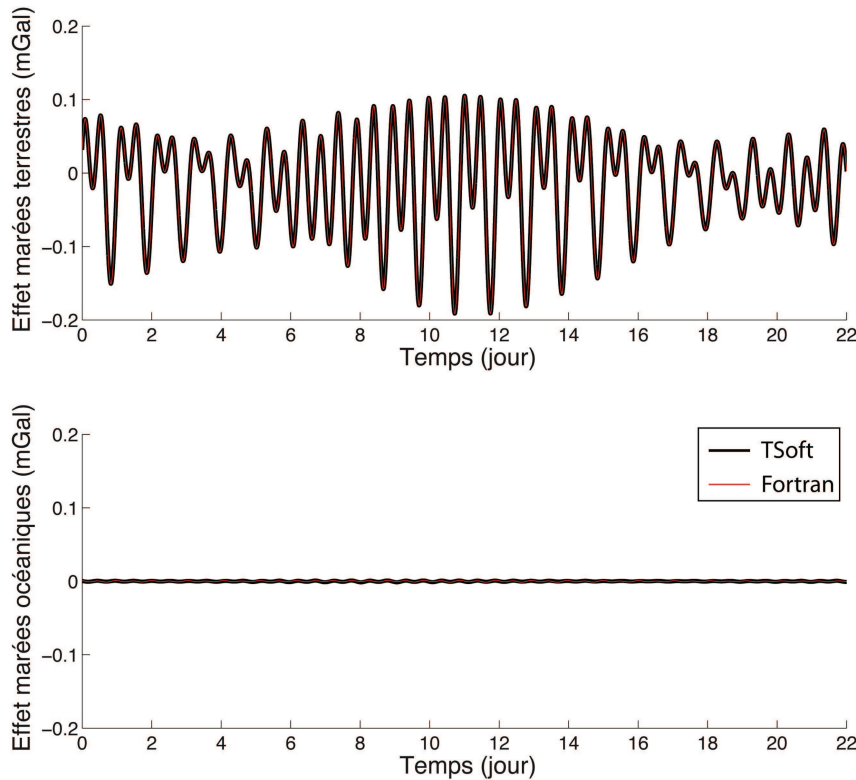


FIGURE 2.14 Effet des marées terrestres (en haut) et des marées océaniques (en bas) calculé à partir de notre routine Fortran (en rouge) et du logiciel *TSoft* (en noir) à Sarpang (point mesuré le plus proche des masses océaniques) durant toute la durée des missions Népal et Bhoutan 2010.

Les valeurs de pesanteur obtenues sont ensuite corrigées de la dérive instrumentale des *CG5* grâce aux boucles journalières effectuées. En considérant la dérive linéaire sur une journée, nous calculons sa valeur et retranchons aux différents points mesurés à l'intérieur de la boucle la valeur de la dérive associée à l'instant de la mesure. Les valeurs de dérive calculées pour les gravimètres utilisés lors des 4 campagnes sur chaque boucle de mesure sont représentées sur la figure 2.15. De manière générale, les dérives sont inférieures au $mGal/j$ et ont tendance à se stabiliser après ~ 3 jours de mesures. La valeur à laquelle se stabilise la dérive semble dépendre du gravimètre et donc de ses caractéristiques intrinsèques. Le *CG5#424* est celui qui présente la dérive la plus forte et la plus variable. Le *CG5#323* a été utilisé pour 3 missions et montre une dérive relativement constante d'environ $0,5 mGal/j$ consécutif aux trois premiers jours de mesure. La valeur de la dérive du *CG5#379*, utilisé sur deux missions, est légèrement inférieure à celle du *#323* : $\sim 0,4 mGal/j$. Le *CG5#167* est quant à lui un instrument qui dérive très peu étant donné que

sa dérive est $<0.1\text{mGal/j}$.

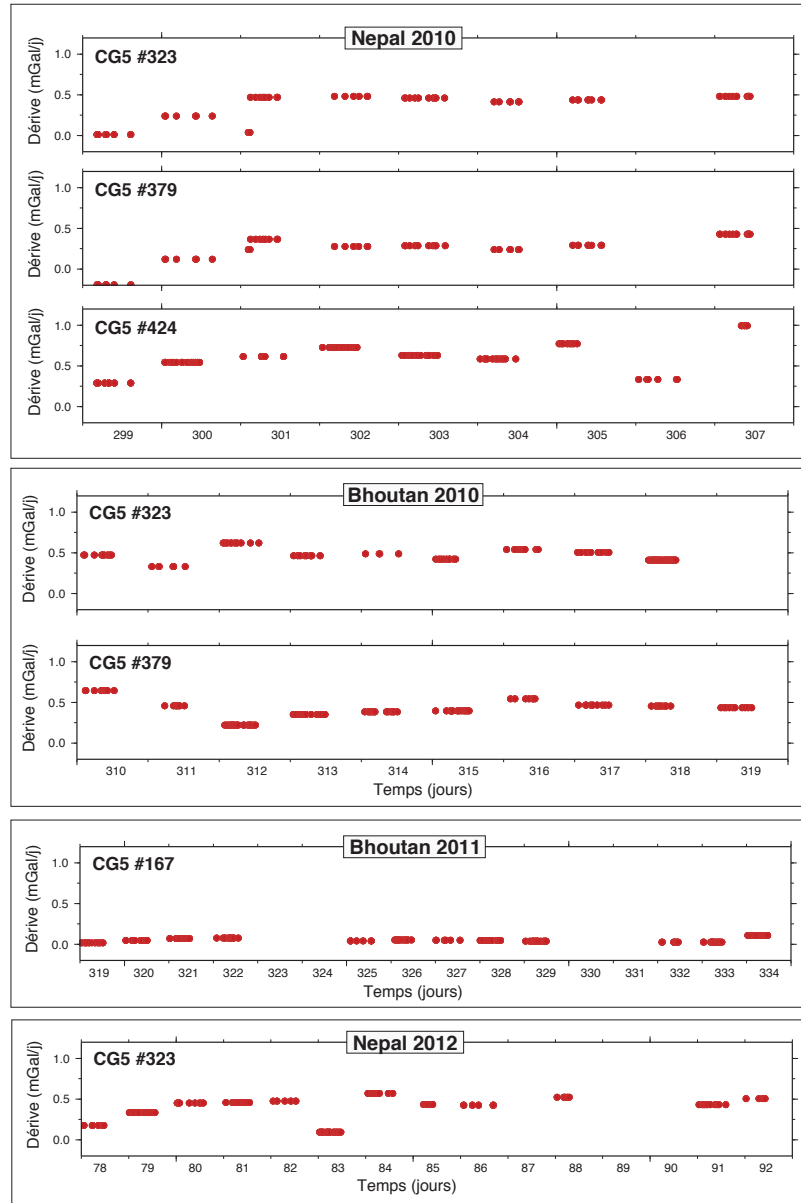


FIGURE 2.15 Dérives des gravimètres calculées pour chaque boucle de mesure effectuée. La dérive se stabilise de manière quasi systématique après ~ 3 jours de mesure. La valeur à laquelle se stabilise cette dérive est propre au gravimètre.

2.4.3 Calcul des anomalies de Bouguer complètes

Les mesures de pesanteur corrigées sont ensuite comparées aux valeurs théoriques sur l'ellipsoïde. Pour une meilleur cohérence avec les données gravimétriques pré-existantes,

la pesanteur théorique est calculée sur l'ellipsoïde GRS67 selon l'équation suivante :

$$g_{theorique} = 9,7803184558(1 + 0,005278895\sin^2\lambda + 0,000023462\sin^4\lambda) \quad (2.4)$$

Afin de ne pas introduire de biais avec les données pré-existantes, l'anomalie de Bouguer est calculée d'après l'équation 2.3 en utilisant une masse volumique ρ de 2670 kg.m^{-3} . Les incertitudes sur les positions verticales obtenues par positionnement relatif, positionnement ponctuel précis et positionnement absolu sont de $\pm 0,30 \text{ m}$, $\pm 2 \text{ m}$ et $\pm 10 \text{ m}$, respectivement. Ce qui correspond à des incertitudes de $\pm 0,06 \text{ mGal}$, $\pm 0,4 \text{ mGal}$ et $\pm 2 \text{ mGal}$, respectivement. La correction de terrain δ_{topo} est calculée avec *TCQ*, basé sur la méthode de quadrature de Gauss (*Hwang et al. 2003*). En champ proche du point de mesure, nous utilisons le modèle numérique de terrain (MNT) possédant la meilleure résolution sur la zone d'étude, *i.e.* le MNT SRTM3 à 3" d'arc. En champ lointain, ce MNT est dégradé à une résolution plus basse de 30" d'arc afin de limiter les temps de calcul sans pour autant détériorer la qualité des corrections calculées. Nous utilisons des rayons internes et externes de 20 et 200 km respectivement, permettant d'atteindre une précision moyenne inférieur au mGal (Fig. 2.16). Les corrections topographiques calculées pour nos points de mesures vont du mGal à $\sim 80 \text{ mGal}$ (Fig. 2.17)

2.4.4 Référencement du réseau gravimétrique bhoutanais

Le nouveau réseau de mesures gravimétriques au Népal est rattaché à des bases absolues de premier ordre. Au contraire, l'absence de bases gravimétriques absolue au Bhoutan rend le nouveau réseau gravimétrique Bhoutanais flottant. Une campagne de mesures absolues a été réalisée en Mars 2012, cependant des problèmes techniques sur le gravimètre absolu *A-10* n'ont pas permis de mesures absolues. L'établissement de bases gravimétriques de premier ordre permettant de connecter notre réseau relatif est prévu dans le cadre de l'ANR Bhutanépal. Afin de caler en absolu le nouveau réseau relatif Bhoutanais, nous l'avons fait coïncider avec les plus proches mesures gravimétriques disponibles en Inde et au Tibet le long des profils Ouest et Est Bhoutan (Fig. 2.18). Le décalage final du réseau bhoutanais se situe dans une fenêtre de $\sim 20 \text{ mGal}$, ce qui induit une erreur absolue de $\pm 10 \text{ mGal}$ sur le réseau Bhoutanais.

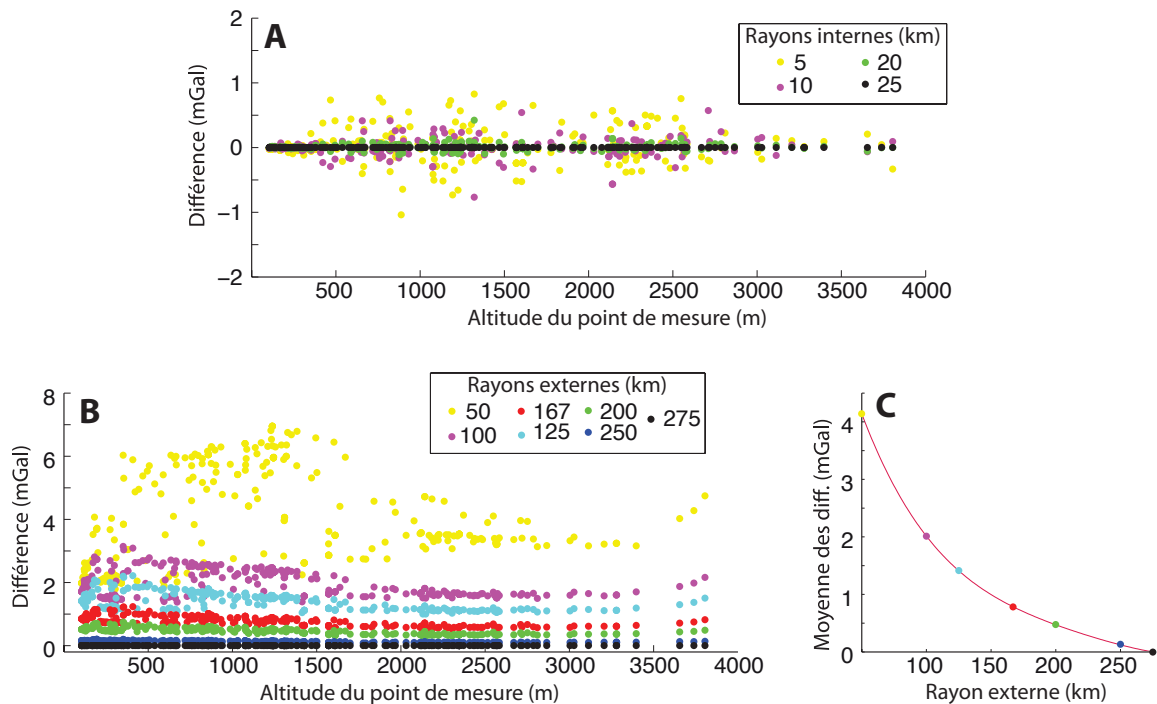


FIGURE 2.16 Influence de la taille des rayons utilisés dans les corrections de terrains. (A) Différences des corrections de terrains calculées avec différents rayons internes par rapport à un rayon de 25 km en fonction de l'altitude. Les différences étant inférieures au mGal à partir d'un rayon de 20 km, ce dernier est suffisant pour la précision recherchée de l'ordre du mGal. (B) Même chose que A avec les rayons externes par rapport à un rayon de 275 km. (C) Moyenne des différences des corrections de terrain en fonction du rayon externe. L'utilisation d'un rayon de 200 km permet une incertitude moyenne inférieure au mGal.

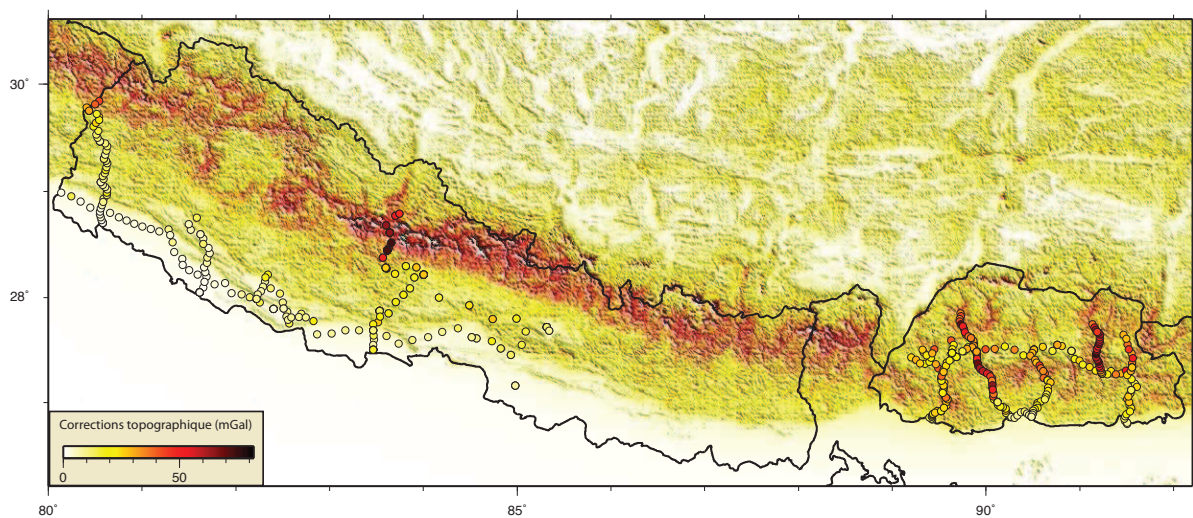


FIGURE 2.17 Carte des corrections de terrain calculées à partir du programme *TCQ* (Hwang *et al.* 2003) sur un MNT à 3" d'arc. Les cercles correspondent aux corrections de terrain calculées pour les points des campagnes gravimétriques Bhutanépal.

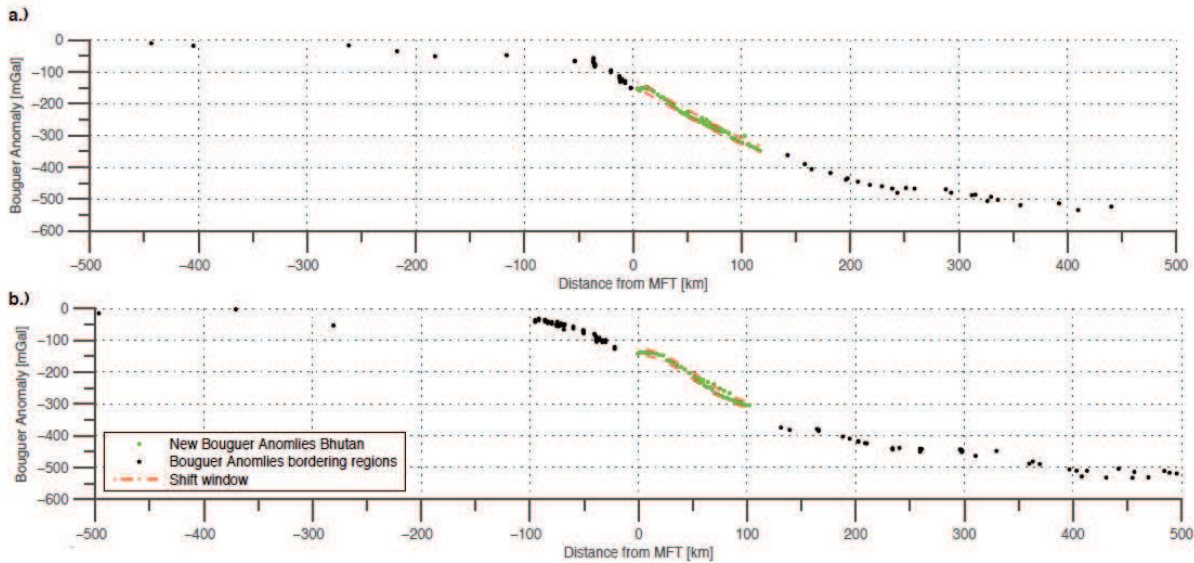


FIGURE 2.18 Anomalies de Bouguer projetées sur deux profils Nord/Sud à l'ouest du Bhoutan (en haut) et à l'est (en bas). Les points noirs correspondent aux données pré-existantes permettant le calage des nouvelles données en vert (les données sont projetées sur une largeur de $\pm 35 \text{ km}$). Les tirets rouges représentent la barre d'erreur de $\pm 10 \text{ mGal}$ associée au recalage des anomalies de Bouguer au Bhoutan.

2.5 Résultats et incertitudes

Les anomalies de Bouguer obtenues durant les missions au Népal et au Bhoutan sont sujettes à des erreurs qui proviennent : (1) des incertitudes relatives sur les valeurs mesurées (et corrigées) de la pesanteur, (2) du calage en absolu de ces mesures et (3) des erreurs associées au calcul des anomalies de Bouguer. Afin d'assigner une incertitude moyenne aux anomalies de Bouguer calculées, chaque source d'erreur doit être estimée et combinée.

(1) La répétabilité des mesures effectuées avec les gravimètres *CG5* est de $\pm 10 \mu\text{Gal}$. L'erreur associée aux corrections des variations de pesanteur dans le temps dues aux phénomènes de marées terrestres et océaniques sont de l'ordre de quelques μGal . Les dérives linéaires des gravimètres sur une journée de mesures excèdent rarement $0,5 \text{ mGal} \cdot \text{j}^{-1}$. Les variations potentielles de la dérive au cours de la journée pourraient donc introduire une erreur de quelques centièmes de mGal . Les incertitudes relatives des valeurs de pesanteur mesurées sont donc de l'ordre de $0,1 \text{ mGal}$.

(2) Le réseau relatif Népalais est relié à des bases absolues de premier ordre dont la précision de mesure est inférieure à $0,01 \text{ mGal}$. Les incertitudes relatives de notre réseau de mesures peuvent être propagées avec l'augmentation de l'ordre des bases gravimétriques. Cependant, les incertitudes absolues sur les mesures effectuées au Nord/Ouest Népal dans

la région de Darchula (Fig. 2.8) à partir de la base gravimétrique la plus éloignée d'une base absolue ne dépassent pas 1 *mGal*. Pour ce qui est du réseau Bhoutanais, en attendant les mesures absolues permettant de connecter notre réseau relatif, nous estimons une incertitude de $\pm 10mGal$ (voir section 2.4.4).

(3) Les incertitudes associées au calcul de l'anomalie de Bouguer proviennent des erreurs sur le calcul de la valeur théorique de la pesanteur et des incertitudes sur l'altitude du point de mesure et sur la correction topographique. L'erreur associée au calcul de la valeur théorique de la pesanteur sur l'ellipsoïde GRS67 est au maximum de 0,004 *mGal*². Les incertitudes sur l'altitude de mesure (erreurs combinées des hauteurs ellipsoïdales et des hauteurs géoïdales) impliquent des incertitudes sur les corrections à l'air libre et de Bouguer simple inférieures à 1 *mGal* sauf pour les quelques points où l'on est contraint d'utiliser le positionnement absolu ($\pm 2 mGal$). L'erreur moyenne sur les corrections topographiques est aussi inférieure au *mGal*, cependant les erreurs peuvent être plus importantes localement (de l'ordre de plusieurs mGal), notamment dans les zones à fort relief.

Les 4 campagnes réalisées dans le cadre de cette thèse entre 2010 et 2012 ont permis d'étendre le réseau gravimétrique Népalais jusqu'à l'Ouest Népal et d'établir le premier réseau gravimétrique au Bhoutan. Au total, ces campagnes ont permis l'acquisition de 366 nouveaux points gravimétriques avec une précision de l'ordre du mGal. Ces nouvelles données au Népal et au Bhoutan complètent les données d'anomalies de Bouguer préexistantes (Fig. 2.19). Ce nouveau jeu de données permet d'obtenir des profils transverses à la chaîne quasi-continue entre l'Inde et le Tibet de l'Ouest Népal jusqu'à l'Est du Bhoutan. Nous utilisons ces données dans les chapitres 3 et 4 afin de contraindre la flexure de la plaque indienne sous l'Himalaya en utilisant des modèles thermo-mécaniques.

2. <http://bgi.omp.obs-mip.fr/data-products/Documentation>

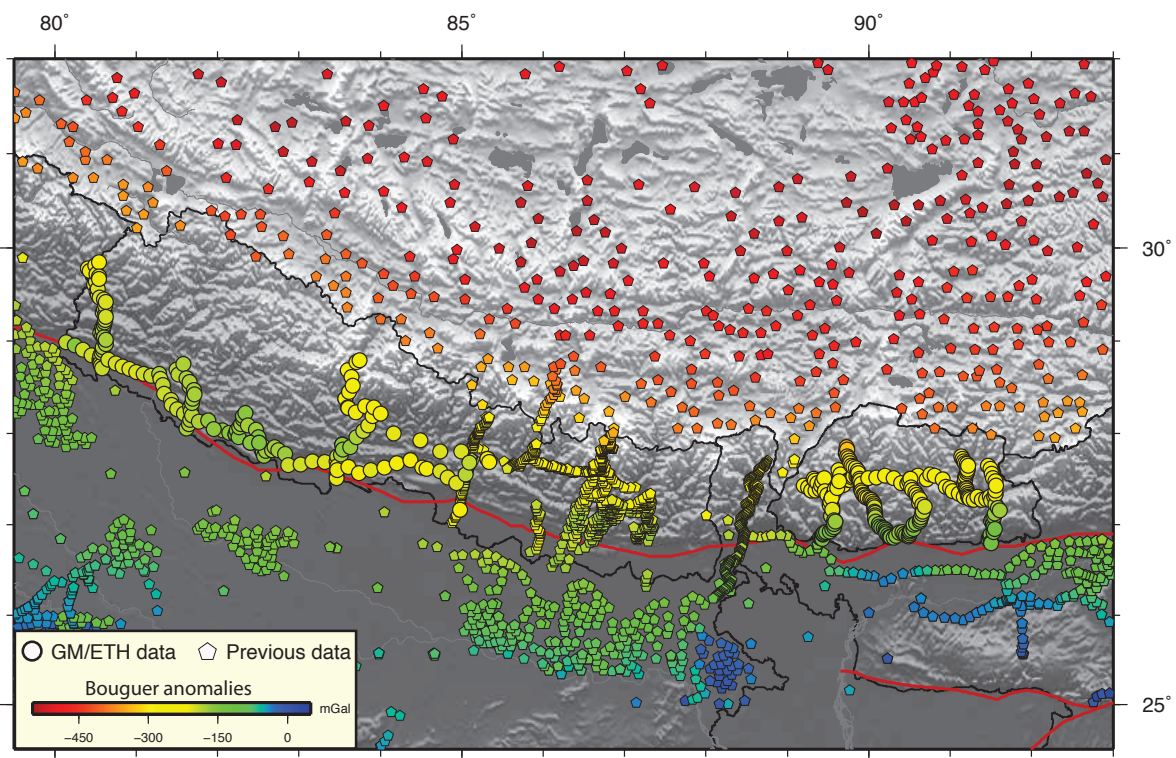


FIGURE 2.19 Carte des anomalies de Bouguer disponibles dans la région de l'Himalaya du Népal et du Bhoutan après les 4 campagnes gravimétriques effectuées au cours de cette thèse.

Retour d'expérience

Fort de l'expérience acquise autant sur l'acquisition que sur le traitement des données gravimétriques, je discute dans cette note de quelques points qui peuvent être appliqués lors de futures missions de ce type.

Traitement GPS : Le post-traitement PPP présente un avantage indéniable pour ce type de mission car il ne nécessite pas l'utilisation de données GPS fixes. Il est d'autant plus utile dans les régions où il n'existe pas de réseau GPS fixe ou encore dans les cas où l'accès aux données est limité par des stations non télé-métrées. Cependant nous avons vu que sa répétabilité verticale est moins bonne que le post-traitement par double différence sur des sessions courtes (quelques dizaines de minutes). L'intégration du réseau GLONASS et l'enregistrement des signaux satellites à une fréquence plus haute pourrait améliorer le résultat. Bien que la répétabilité des mesures soit fonction de paramètres externes comme la répartition des satellites, une étude systématique simple pourrait être menée afin d'estimer, en fonction des longueurs des sessions d'enregistrement, la répétabilité vers laquelle on peut tendre.

Dérive des gravimètres : La stabilité de la dérive des *CG5* est impressionnante et ce malgré la mauvaise qualité des routes parfois empreintées. Systématiquement, en plus des deux jours de mise en chauffe, la dérive se stabilise après environ ~ 3 jours de mesure à une valeur qui semble fonction du gravimètre utilisé. Il serait intéressant que les utilisateurs des gravimètres du parc INSU puissent avoir accès à ces données afin d'adapter leur stratégie d'acquisition.

Stratégie d'acquisition : Une stratégie alternative dans le cas de nos missions aurait été de contrôler la dérive en début de mission et de commencer les mesures une fois la dérive stabilisée. À partir de ce moment nous pourrions effectuer des boucles sur plusieurs jours tout en contrôlant la dérive quand le gravimètre est en station (par exemple la nuit). S'il s'avère, comme le suggère la figure 2.15 que certains gravimètres dérivent très peu, il est même possible de s'en affranchir en fonction de la précision recherchée tout en gardant à l'esprit la propagation des erreurs dans le temps.

Traitement des anomalies : Si la décomposition du calcul des anomalies en Bouguer simple et Bouguer complète présente un avantage pédagogique, l'abandon des abaques au profit de l'outil informatique rend aujourd'hui obsolète cette décomposition. D'autant plus que le calcul direct de l'anomalie de Bouguer complète sans passer par l'étape Bouguer simple est le même que pour le calcul de la correction topographique.

CHAPITRE 3

COMPORTEMENT FLEXURAL DE LA PLAQUE

INDIENNE SOUS LE NÉPAL

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3.1 Présentation

Dans ce chapitre, je tire partie des données gravimétriques nouvellement acquises afin d'étudier les variations latérales majeures de structures lithosphériques sous le centre et l'est Népal.

Pour minimiser la non-unicité des solutions du modèle direct, j'utilise aussi les contraintes obtenues par les expériences HIMNT (*Schulte-Pelkum et al. 2005*), Hi-CLIMB (*Nabelek et al. 2009*), et le traitement de 15 stations du réseau Hi-CLIMB (ligne NP) déployées le long du nouveau profil gravimétrique centre Népal. Ces données permettent de contraindre un modèle thermo-mécanique de la flexure de la plaque Indienne sous l'effet de la propagation du Tibet, de l'Himalaya et des sédiments accumulés dans le bassin flexural (*Cattin et al. 2001; Hetényi et al. 2006*). J'ai amélioré le modèle thermo-mécanique en appliquant un champ de pression vertical sur la plaque Indienne plus réaliste qui inclut des variations temporelles et spatiales et tenant compte à la fois des variations topographiques et des résultats de Hi-CLIMB sous le Tibet.

J'ai effectué une étude paramétrique permettant de mettre en évidence l'influence des propriétés élastiques, thermiques et rhéologiques de la plaque Indienne. Cependant, je montre qu'une rhéologie homogène de la plaque Indienne suffit à expliquer les données entre le centre et l'Est Népal avec un support flexural principalement contrôlé par le manteau lithosphérique Indien. Ceci suggère qu'il n'existe pas de variations latérales majeures du comportement mécanique de la lithosphère Indienne sur ~ 350 km le long de l'arc. Ce résultat suggère donc une certaine continuité géométrique sur cette zone. Cette étude est présentée sous la forme d'un article accepté à *Geophysical Journal International*.

3.2 Lateral uniformity of India plate strength over Central and Eastern Nepal

*Paper accepted in Geophysical Journal International*¹

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Abstract

The current understanding of the Himalayan lithosphere stems mostly from cross-sections through the range at the longitude of the Kathmandu basin. In this paper we laterally extend the analyses of structures and rheology along the Nepal Himalayas between the Pokhara valley and the Arun river. We take advantage of available informations and a new dataset including gravity measurements and a receiver function profile. It appears that the geometry of the Moho inferred from seismological profiles and long wavelength gravity anomalies does not exhibit major East-West variations within the 350 km wide study area. Using thermomechanical modelling, we show that the northward deepening of the Moho observed along profiles perpendicular to the main thrust faults can be interpreted simply as the bending of a strong India plate. This result suggests a gradual mechanical decoupling between the crust and the mantle, leading to a northward decrease of the effective elastic thickness of the Indian lithosphere from ~ 75 km to ~ 25 km beneath the Ganga Basin and the Tibetan Plateau, respectively. Our results also confirm (partially) eclogitized lower Indian crust beneath southern Tibet. At shorter wavelengths, the observed gravity profiles exhibit some small lateral variations that can be interpreted in terms of east-west variations of the thickness of sub-surface geological structures such as the Ganga Basin and the Tethyan Sedimentary Sequence.

Keywords : Gravity anomalies and Earth structure, Lithospheric flexure, Continental margins : convergent, Crustal structure, Dynamics : gravity and tectonics, Rheology : crust and lithosphere

3.2.1 Introduction

The Himalayan belt along the southern edge of the Tibetan Plateau is one of the most spectacular products of the ongoing collision between India and Eurasia. Over the last four decades many seismological and gravity experiments as well as GPS, structural geology, geomorphology and thermochronological campaigns were conducted in the Himalayas to image its deep structure and understand its dynamics (e.g. *Le Fort 1975; Hirn et al. 1984; Zhao and Nelson 1993; Brown et al. 1996; Alsdorf et al. 1998; Cattin et al. 2001; Jouanne et al. 2004; Schulte-Pelkum et al. 2005; Hetényi et al. 2006; Nabelek et al. 2009; Robert et al. 2011*). The main features of the collision zone are now relatively well-known : beneath the range, the India plate is bent and underthrusts the southern Tibetan plateau for about 450 km distance. The thickness of the crust increases by ~ 40 km from India to Southern Tibet. The shortening is accommodated along major frontal thrust faults connected to the Main Himalayan Thrust (MHT), which is a mid-crustal reflector imaged at a depth of ~ 25 to ~ 40 km. However, because most of these available data are located in central Nepal, the current understanding of the Himalayas is mostly limited to the cross section through the range at the longitude of the Kathmandu basin. The lateral variation of structures is nevertheless important. First, to unravel the mountain building processes. For instance, along strike variations can reveal rheological contrasts or inherited structures, which can play a major role in the orogenic evolution (*Gahalaut and Kundu 2012*). Second, to improve seismological hazard assessment due to self-similarity of earthquakes as the magnitude of future events can be estimated from fault lengths. Thus the size of lateral similarities of major structures in a seismically active area may control its seismic potential.

Several studies have recently demonstrated that along strike variations exist between the Nepal and Bhutan Himalayas in topography (*Duncan et al. 2003*) or geometry of the Main Himalayan Thrust (*Robert et al. 2011*). However, these studies do not exhibit significant lateral variations in the Nepal Himalayas. In Central and Eastern Nepal, geodetic data also show very little variations (*Grandin et al. 2012*). Due to lack and sparseness of geophysical data and the difficulty to assess the consistency of these different methods, the extent of lateral variations at depth is still unconstrained in Nepal.

Here, using a new dataset including gravity measurements and receiver functions profiles, we study the lateral variations in terms of behaviour of the India plate in flexure

along the Nepal Himalayas between the Pokhara valley and the Arun river (Fig. 1). After a short description of the study area, we present data acquisition and processing. We then combine them with available data in order to obtain four 1000 km long profiles within a 350 km wide area in Central and Eastern Nepal. We next describe the modelling approach and assumptions. Finally we discuss the different wavelengths of the gravity measurements and our findings in terms of flexure of the India plate and structural variation along the Nepal Himalayas.

3.2.2 Geodynamic setting of the Nepal Himalayas

As the entire Himalayan arc, the Nepal Himalayas can be divided into four distinct tectonic units (Fig. 3.1) (e.g. *Gansser 1964; Le Fort 1975; Yin 2006*). The northernmost zone, known as the Tethyan Sedimentary Series (TSS), consists of Cambrian to Eocene sedimentary and low grade metamorphic rocks. In Nepal, the TSS has a limited extent except in Central Nepal where the best sections can be found in the Annapurna-Dhaulagiri region. The TSS lies between the Indus-Yarlung Suture (IYS) which is the northern limit of the Himalayas, and the South Tibetan Detachment, a north dipping normal fault which marks the limit with the underlying Higher Himalaya (HH) (*Burchfield et al. 1992*). The HH consists of high-grade metamorphic rocks including gneisses, schists and migmatites. The width of this unit varies along Nepal, it is only a few kilometers where the unit is limited to its root zone like in Central Nepal whereas it can reach tens of kilometers where it has overthrust the Lesser Himalaya (LH) forming series of nappes and klippe in Western and Eastern Nepal (*Upreti 1999; DeCelles et al. 2001*). The LH underlies the HH along the Main Central Thrust (MCT) and forms a series of duplexes composed of low-grade metasedimentary rocks. In eastern Nepal, the LH is mainly exposed in tectonic windows that cut through the extensive crystalline thrust sheet. In Central Nepal, the width of LH reaches 100 km because of the absence of overthrust HH. Crystalline nappes and klippe reappear in western Nepal. The last unit, composed of Neogene to Quaternary syntectonic clastic sediments, is formed by the Siwalik thrust belt. It lies between the Major Boundary Thrust (MBT) to the North and the Major Frontal Thrust (MFT), which is the most recently activated one in the sequence (*Lavé and Avouac 2001*). South of the MFT is the undeformed foreland basin with relatively recent sedimentary infill up to 6 km in thickness (*Raiverman et al. 1983*).

The localization of the deformation is now well-constrained in Central and Eastern Nepal. The present day shortening rate of $17.8 \pm 0.5 \text{ mm.yr}^{-1}$ is related to interseismic coupling on the Main Himalayan Thrust fault (*Bilham et al. 1997; Larson et al. 1999; Jouanne et al. 2004; Ader et al. 2012*). At longer timescales, fluvial terrace analyses and tectonostratigraphic studies give a consistent rate (*Lavé and Avouac 2000; DeCelles et al. 2001; Mugnier and Huyghe 2006*), suggesting that the Himalayan arc absorbs about 20 km/Myr convergence by localized thrusting since mid-Miocene times.

Many structural geology field campaigns as well as several seismological and gravity experiments were performed to infer and image the deep structures of the Himalayas in Nepal, which includes the geometry of major faults and the depth of both the Moho and the foreland sedimentary basin. The three north dipping thrusts of the arc (MCT, MBT, MFT) all root on the MHT (*Zhao and Nelson 1993; Nelson et al. 1996; Lavé and Avouac 2000*). The MHT consists of a mid-crustal décollement where the India plate underthrusts the Himalayas and Tibet. Several studies on the shape of the MHT suggest that a ramp-flat-ramp-flat geometry prevails (e.g. *Cattin and Avouac 2000; Avouac et al. 2001; Nabelek et al. 2009*). The first ramp corresponds to the currently active MFT and the second one occurs at mid-crustal depth below the topographic front of the Higher Himalaya.

In the Nepal Himalayas, *Herman et al. (2010)* suggest a dip of 15° for a ramp that begins ~ 70 km North of the MFT. More recently, *Robert et al. (2011)* have proposed models in which the MHT exhibits lateral variations in terms of geometry along the Himalayan arc.

In eastern Nepal, the Himalayan-Nepal-Tibet seismic experiment (HIMNT) reveals the deepening of the Moho beneath the range from ~ 35 km under the MFT to ~ 75 km beneath the IYS (*Schulte-Pelkum et al. 2005*). West of this profile along $\sim 85^\circ$ E, the same trend is imaged by the Hi-CLIMB project (Himalayan-Tibetan Continental Lithosphere during Mountain Building) with receiver functions calculations. As this experiment runs farther north than HIMNT, it covers the underthrust and flattened Indian lower crust extending from the IYS until near the Banggong-Nujiang Suture (*Nabelek et al. 2009*). The top of the Indian lower crust (~ 15 km thick) is well imaged especially beneath southern Tibet where it is interpreted to be eclogitized (*Wittlinger et al. 2009*). Bouguer anomaly profiles across the range show a variation of about -500 mGal from India to Southern Tibet and confirm the northward thickening of the crust (*Karner and Watts 1983; Lyon-Caen and Molnar 1985; Jin et al. 1996*). Beneath the range, the gravity data are no longer consistent with

Airy isostasy suggesting that the weight of the Himalayas is supported by the strength of the underthrusting India plate (*Cattin et al. 2001*). By combining these data with thermo-mechanical modelling, *Hetényi et al. (2006)* have shown that the Effective Elastic Thickness (EET) decreases from 60-80 km in India to 20-30 km beneath Tibet where the Himalayas are mostly supported by the Indian lithospheric mantle. Based on forward gravity modelling and petrological models, *Hetényi et al. (2007)* suggest that the Indian lower crust has to be eclogitized in order to explain both seismological and gravity data.

The gravity data also exhibits a kink associated with the density contrast of the Ganga basin, with its depth mainly control by the flexural subsidence of the India plate (*Lyon-Caen and Molnar 1985*). South of Nepal, the foreland basin has a variable width with a mean extent of about 200 km and reaches a maximum depth of 6 km in the Gandak depression. The basement of the Ganga basin is affected by inherited structures showing a succession of spurs and depressions. Three depressions are separated by two highs in the basement : from west to east, the Sarda depression, the Faizabad ridge, the Gandak depression, the Munghyr Saharsa ridge and the Purnea basin. Each ridge is connected with an Archean massif which crops out south of the basin, respectively the Bundelkhand massif and the Satpura massif (*Raiverman et al. 1983; Gahalaut and Kundu 2012*).

3.2.3 Gravity and seismological data

3.2.3.1 Available gravity dataset

For this study, we merge three sources of existing gravity data (Fig. 3.1). In India, Bouguer anomalies are provided by the Bureau Gravimétrique International (BGI²). The data are tied to IGSN 71 reference with a reduction density of 2670 kg.m⁻³ but do not include terrain correction. North of Nepal, gravity anomalies are taken from the 5x5 arcmin gridded Bouguer anomaly map (*Sun 1989*) resampled at the measurement locations. Those Bouguer anomalies were also calculated using a reduction density of 2670 kg.m⁻³ and include terrain correction. In Nepal, gravity data are available from joint French-Nepali surveys along three profiles perpendicular to the Himalayan arc (*Cattin et al. 2001; Martelet et al. 2001*). The corresponding Bouguer anomalies computed with terrain correction and a reduction density of 2670 kg.m⁻³ have an accuracy ranging from

2. <http://bgi.omp.obs-mip.fr/>

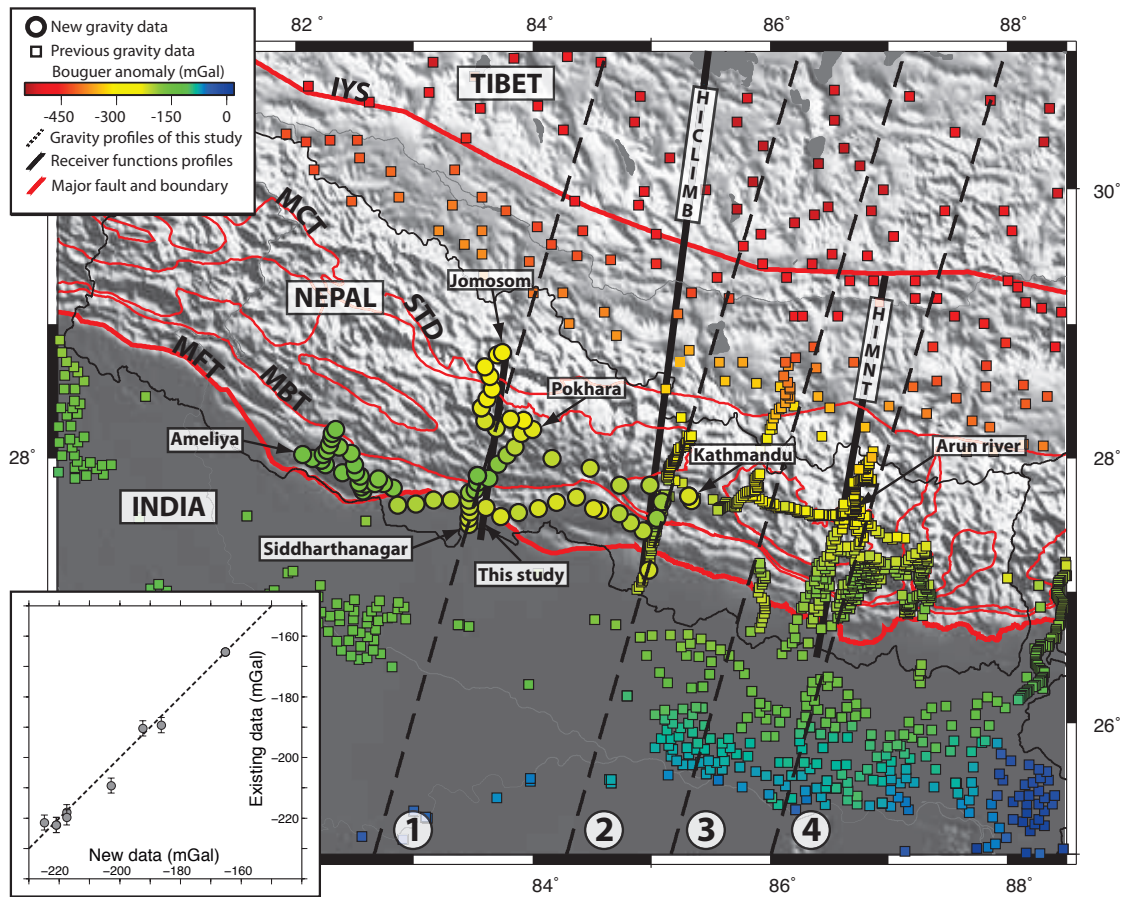


FIGURE 3.1 Map of the study area showing gravity data in Nepal, Northern India and Southern Tibet. Circles represent the location of new gravity measurements and squares the previous dataset (from BGI, *Cattin et al. (2001)*, *Sun (1989)* and *Tiwari et al. (2006)*). The colour scale shows Bouguer anomalies in mGal. Also shown are receiver function profiles obtained in eastern Nepal : HIMNT experiment (*Schulte-Pelkum et al. 2005*) and in the Hi-CLIMB experiment's main array (*Nabelek et al. 2009*) along 85°E and at the longitude of Pokhara (this study, Hi-CLIMB experiment's Nepal-Pokhara line). Black dashed lines show the locations of profiles 1 to 4 used in this study. Thick red lines represent major faults. MFT : Main Frontal Thrust. MBT : Main Boundary Thrust. MCT : Main Central Thrust. STD : South Tibetan Detachment. IYS : Indus-Yarlung Suture. Inset in the lower left corner shows the discrepancy between new measurements and existing ones.

0.5 to 5 mGal. Even though all these datasets are relative, they are consistent with the BGI reference (*Hetényi et al. 2007*). Their compilation allows us to obtain a gravity map from India to Tibet, but with a major data coverage gap in Nepal west of Kathmandu.

3.2.3.2 New gravity data, acquisition and processing

To complete this available dataset and fill the existing data coverage gap in Central Nepal, three Scintrex CG5 relative gravimeters, previously intercalibrated, were used to obtain 80 new gravity observations. Due to survey loops needed to correct instrumental drift, 120 measurements are carried out along two profiles, a 300 km long parallel to the strike of the range from Kathmandu to Ameliya and a 150 km long arc-perpendicular at the longitude of Pokhara extending from Siddharthanagar in the South to Jomosom in the North (Fig. 3.1). For each site, at least five gravity measurements of 90 seconds and fifteen minutes of GPS observations with dual-frequency geodetic receivers are made. The accurate locations of our measurements are obtained by double difference processing with Trimble Geomatic Office and the cGPS stations operated by the CEA/LDG and the National Seismological Center in Kathmandu. We obtain a vertical height accuracy of less than 60 cm for all data points. The EGM08 geoid model is then used to convert ellipsoidal heights to elevations. Gravity measurements are corrected for ocean and earth tides using FES2004 model (*Lyard et al. 2006*) and ETERNA package (*Wenzel 1996*), respectively. We also correct for the linear drift of the gravimeter after daily loops. The free air anomaly is computed using the GRS67 ellipsoid and we estimate its accuracy to be of the order of 0.15 mGal. The complete Bouguer anomaly is calculated with a reduction density of 2670 kg.m^{-3} . Terrain correction are computed using TCQ package (*Hwang et al. 2003*) with SRTM digital elevation model for an inner zone of 20 km and an outer zone of 200 km. The accuracy of the final complete Bouguer anomaly depends on the terrain roughness and ranges between 0.15 and 0.60 mGal, knowing that the main source of error is the accuracy of the vertical position.

3.2.3.3 Consistency of datasets

In order to use the gravity dataset obtained in this study together with the existing Bouguer anomaly dataset, they have to be consistent in terms of reference and corrections. The same reduction density is used to calculate the Bouguer anomalies of both datasets.

The free air anomalies for the new points are computed with the GRS67 instead of GRS80 to be consistent with the existing dataset. All the new measurements are tied to absolute gravity base-stations in Kathmandu, Simra, Siddharthanagar and Tansen. These points have a measured accuracy of less than 0.01 mGal and are defined in the IGSN71 Network as the data provided by BGI. Furthermore, to check the consistency with previous studies, we compare the new measurements with the existing ones at close location. Inset of figure 3.1 shows a very good agreement between the datasets, with a discrepancy of only a few mGal.

3.2.3.4 Bouguer anomaly profiles

Our new measurements together with the available Bouguer anomalies allow constructing four profiles, 1000 km long each, perpendicular to the strike of the orogen (oriented N16° E) and centred on the MFT (Fig. 3.2). The profiles cover a 350 km wide area in Central and Eastern Nepal between the Pokhara valley and the Arun river. Along each profile, the Bouguer anomaly decreases between -25 mGal to the South and -525 mGal to the North. The ~ 500 mGal decrease occurs in about 600 km from South of the Ganga basin above the India plate to the Indus-Yarlung Suture Zone on the Tibetan Plateau. Based on the shape of the Bouguer anomaly variations, these profiles can be divided into three distinct zones :

1. The India section, from -500 km to 0 km distance, exhibits a convex curve with a decrease of ~ 200 mGal. The first part of this section which is nearly flat with a very low gradient corresponds to the signature of the India Plate and the beginning of its flexure. The second part, with a steeper gradient, can be related to the additional effect of the Ganga Basin where low density sediments have been accumulated.
2. The Tibetan Plateau section, from 300 km distance onwards, begins close to the location of the Indus-Yarlung Suture Zone and shows a flattening of the Bouguer anomaly at a value of ~ -525 mGal.
3. In between these two sections lies the Himalaya section. It begins near the MFT with a local minimum corresponding to the maximum thickness of sediments accumulated in the Ganga Basin, then the Bouguer anomaly increases by ~ 50 mGal on a distance that varies somewhat between profiles. After reaching a local maximum, near the MBT, the Bouguer anomaly decreases to reach the aforementioned plateau at the

beginning of the Tibetan section. Despite the average gradient being similar in all the profiles, we observe minor variations. For example, profile 1 exhibits two different gradients in this zone.

By comparing these four profiles we observe that the long-wavelength anomaly, which presumably reflects deepening of the Moho, is quite similar on all four profiles. In contrast, figure 3.2 shows that lateral variations in the Bouguer anomaly signal exist at shorter wavelengths, especially in the Ganga basin and in the Himalaya section.

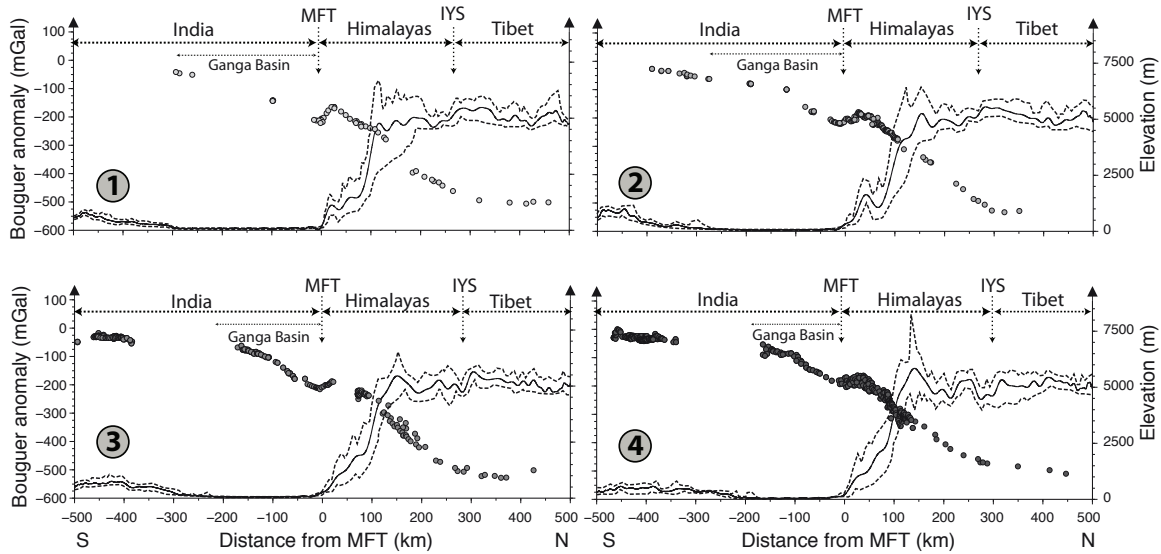


FIGURE 3.2 Bouguer anomaly along the four 1000-km-long N16°E profiles (see Fig. 3.1 for location of the profiles). All the data within a 25 km wide swath on both sides of the profiles were considered and error bars lie within the circles. The profiles are centered at the surface location of the MFT. Location of MFT and IYS as well as extension of the Ganga Basin are also shown. The area between the two dashed lines represent the topography on a 25 km wide swath for each profile from SRTM data with the black line corresponding to the average.

3.2.3.5 Receiver function profiles

The Moho is the main interface within the lithosphere that reflects the geometry of the flexure and mainly controls the location of density contrasts. To complement the constraints on the shape of the plate flexure given by gravity anomalies, we search for available seismological profiles showing the geometry of the Moho. Among passive seismological methods, the Moho is best imaged using receiver functions (RF) : teleseismic P-waves that convert to S-waves beneath the observing station. The delay of the converted S-wave with respect to the direct P-wave contains information on the depth of the Moho and the average velocity structure of the crust (*Langston 1977*).

In Nepal clear images of the Moho are readily available from migrated receiver function images for profile 2 (Hi-CLIMB experiment, main array) as well as for the region of profiles 3 and 4 (HIMNT experiment) (Fig. 3.1). For profile 1 associated with our new gravity data, we process the seismological data from 15 stations of the Hi-CLIMB experiment's Nepal-Pokhara (NP) line (Fig. 3.1), deployed along the road along which

the gravity measurements were made. These stations were operating between 4 and 5 months, and registered teleseismic waveforms from a total of 161 earthquakes of M5.5 or above. The collected data were passed through a quality control procedure to yield 727 high-quality traces. Receiver functions were computed using iterative deconvolution with a maximum frequency of the signal at 1 Hz. These RFs were then migrated from time to depth using the 2D-velocity model of the Hi-CLIMB main array including correction for topography. Complete information on the Hi-CLIMB experiment as well as the receiver function processing is described in *Hetényi (2007)* and *Nabelek et al. (2009)*.

The final receiver function image for profile 1 is shown in figure 3.3. Despite some noise in the Himalayan foothills, already observed along the main array, the Moho of the India plate can be followed along ~ 150 km distance and can be identified with an uncertainty of ~ 5 km. The image shows a flat Moho at about 35 km depth in the south which starts to dip down north of the MFT at an angle of $\sim 12^\circ$ to reach ~ 60 km depth beneath the northernmost station. In comparison to the RF images of profiles 2, 3 and 4, this profile is shorter but shows a similar trend : a flat Moho turning to a ramp at a similar angle. Compared to other profiles, the flat-to-ramp transition here is more abrupt (Fig. 3.6), but considering the error-bars the difference is not significant.

With the processing of seismological data from the Hi-CLIMB experiment's NP line, the flexural shape of the underthrusting India plate within the entire study area can be constrained both by gravity and seismological data.

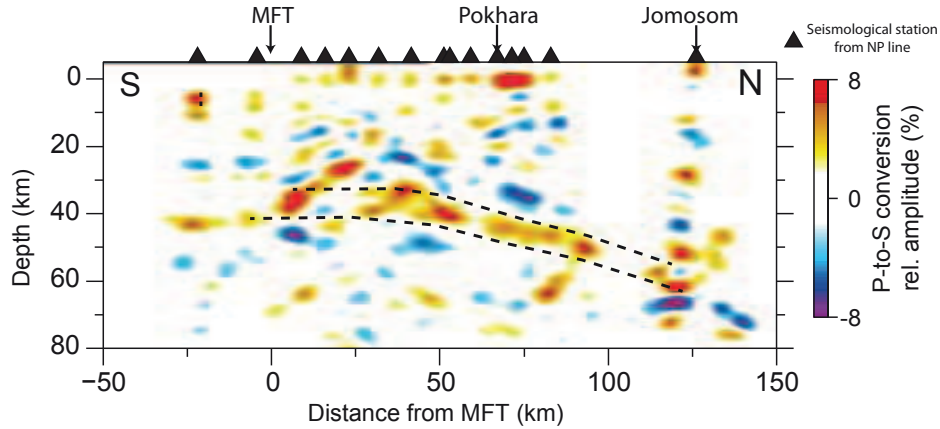


FIGURE 3.3 Migrated receiver function image along profile 1 (see Fig. 1 for location) using 15 stations of the Hi-CLIMB experiment's Nepal-Pokhara line. The Moho appears clearly in warm colours as a positive amplitude conversion of P- to S-waves, and can be followed along the profile between the dashed lines. See text for more details.

3.2.4 Thermomechanical modelling

Deviations of the gravity field across the Ganga Basin and the Himalayas from local Airy isostasy, are commonly interpreted to result from the elastic support provided by the flexed Indian plate (*Cattin et al. 2001*). The flexural shape of the lithosphere depends on its composition, its thermal structure and the thickness of the composing layers. Across the Himalayas the changes of these parameters can be simulated by varying the thickness of an elastic plate overlying an inviscid fluid and loaded by masses above (e.g. *Karner and Watts 1983; Lyon-Caen and Molnar 1985; Jordan and Watts 2005*). In such an approach the dependency of deformation on local temperature and pressure as well as the relative effect of the crust and mantle cannot be dealt with. Thus, following *Cattin et al. (2001)* and *Hetényi et al. (2006)* we use a more comprehensive approach that accounts for the mechanical layering of the lithosphere, the non-Newtonian rheology of rocks and their dependency on temperature and pressure. We use a two-dimensional finite element model, ADELI (*Hassani et al. 1997*), that was modified to incorporate both temporal and spatial variations of the loading.

3.2.4.1 Rheology and thermal structure of the India lithosphere

In our model the India plate is composed of three bodies : upper crust, lower crust and lithospheric mantle (Fig. 3.4). These three layers are deformed in an elastic, brittle or ductile way depending on the local deviatoric stresses, pressure and temperature. The elasticity is expressed by Hooke's law,

$$\varepsilon_{ij} = \frac{1 + \nu}{E} \sigma_{ij} - \frac{\nu}{E} \text{tr}(\sigma) \delta_{ij}, \quad (3.1)$$

with ε the strain tensor, σ the stress tensor, E the Young modulus and ν the Poisson's ratio. The non-Newtonian viscous behaviour, dependent on temperature T , is controlled by the following relationship between differential stress and strain rate $\dot{\varepsilon}$,

$$\dot{\varepsilon} = \gamma_0 (\sigma_1 - \sigma_3)^n e^{\frac{E_a}{RT}}, \quad (3.2)$$

where γ_0 is the standard fluidity, n the power law exponent, E_a the activation energy and $R = 8.314 \text{ J.mol}^{-1}.\text{K}^{-1}$ the universal gas constant. σ_1 and σ_3 are the maximum and minimum principal stress, respectively. The limit between the visco-elastic and the plastic domains is defined by Drucker-Prager failure criterion, according to,

$$\frac{1}{2} (\sigma_1 - \sigma_3) = \left(c (\cot \phi) + \frac{1}{2} (\sigma_1 + \sigma_3) \right) \sin \phi, \quad (3.3)$$

with ϕ and c the internal friction angle and the cohesion, respectively. The laboratory-derived material properties are used under the assumption that they can be extrapolated to geological conditions (Tab. 3.2).

The thermal structure of our model is computed from an analytical approach based on the 1-D approximation of *Royden (1993)*. In this calculation (following *Cattin et al. (2001)*), the temperature is mainly controlled by the geometry of the India plate. The thermal structure in the initial time step and in the final model are shown on Fig.3.4. This thermal field is consistent with the well constrained model presented by *Bollinger et al. (2006)* based on numerous field observations across the Himalayan range in Central Nepal.

3.2.4.2 Geometry and boundary conditions

To simulate the flexure of India, we use a 1000-km-long flat plate centred on the MFT (Fig. 3.4). As an initial condition, we consider 9000 triangular elements to compose

the three layers. The crust is 35 km thick (*Kumar et al. 2001*) and its density contrast with the lithospheric mantle is 370 kg.m^{-3} . The model is submitted to gravitational force ($g = 9.81 \text{ m.s}^{-2}$), supported by hydrostatic pressure at its base with a vertical density profile taken from the PREM (*Dziewonski and Anderson 1981*).

To simulate the bending of the plate, both vertical sides of the model allow for free vertical displacement and the top of the India plate is loaded by the mass of the Himalayas and the Tibetan plateau. Compared to *Cattin et al. (2001)* we have improved the model by applying a more realistic loading which evolves during the numerical simulation. This vertical loading on top of the India plate excludes shear forces and is calculated at each time step by adding the effect of surface load (topography) and subsurface loads (Tibetan crust and of the sediments accumulated in the foreland basin).

We calculate the topographic load by averaging the topography in the four studied profiles and we assume a mean density of 2670 kg.m^{-3} above 0 m (Fig. 3.6a). For the crustal material above the India plate, we use a constant Moho depth of 75 km beneath Tibet, where both HIMNT and Hi-CLIMB experiments (*Schulte-Pelkum et al. 2005; Nabelek et al. 2009*) reveal a flattening of the Moho at the latitude of the IYS ($\sim 300 \text{ km}$ distance on our projected profile). Thus, as we do not consider transfer of material between India and Tibet during the numerical simulation, the pressure applied north of the IYS is constant and equals to the load of a $\sim 45 \text{ km}$ thick Tibetan crust including topography.

From south of the IYS to the MFT, and for the foreland basin, the load depends on the geometry of the bent India plate. It implies that the load ΔP is calculated, at each time step, depending on the northward deepening of the India plate h_{sub} by

$$\Delta P = \rho \cdot g \cdot h_{sub}, \quad (3.4)$$

where ρ is 2500 kg.m^{-3} south of the MFT to reflect the lower density of sediments accumulated in the foreland basin and 2900 kg.m^{-3} north of the MFT, an average for the whole Tibetan Crust.

This loading is applied during 10 Myr on the India plate with a north-to-south movement corresponding to an underthrusting rate of 15 mm.yr^{-1} (*Cattin et al. 2001*). These 10 Myr are divided in 4000 time steps in our experiment and correspond to the order of magnitude for the estimated time of the India plate flexure. We then stop the evolution of the loading and obtain a nearly steady-state geometry which we will consider hereafter.

The main limitation of the used thermomechanical model is that it is two-dimensional

and that it neglects horizontal shortening. Moreover, we do not take into account the rheological behaviour of the eclogite layer in the India lower crust underneath Tibet because of a lack of reliable rheology data at this pressure-temperature range and technical limitations of the modeling tool. In the following we analyze profiles across the Ganga basin, the Himalayas and southern Tibet, perpendicular to the main geological structures.

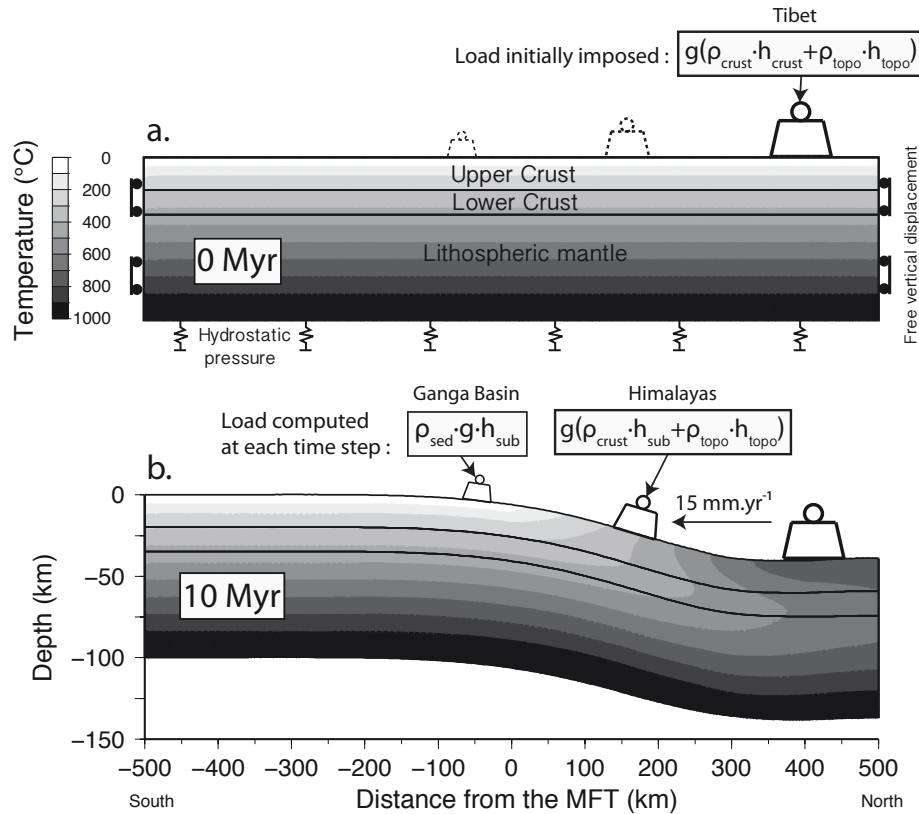


FIGURE 3.4 (a) Geometry, boundary conditions and temperature of the initial model used for the thermo-mechanical modelling with ADELI (*Hassani et al. 1997*). The Indian lithosphere is progressively bent down by the vertical load applied on its top. The loading moves southward at a rate of 15 mm.yr⁻¹ and corresponds to the increasing weight of the sediments in the foreland, Himalayas and the Tibetan plateau. The model is subject to gravity and is sustained by hydrostatic pressure at its base. Free vertical displacements are allowed at both sides of the model. The three layers considered have different rheologies (described in table 3.2). (b) Geometry and temperature field of the model after 10 Myr.

3.2.5 Results

The effects of the thermal structure, the rheological parameters as well as the thickness and the density of the lithospheric layers are tested using a large amount of numerical

experiments with various density contrast at the Moho, crustal and lithospheric thickness, Poisson ratio, Young's modulus, viscous power law parameters, temperature at the basis of the lithosphere and thermal conductivity.

Complexity of the modelling lies in the trade off between parameters which control the flexural rigidity. For instance an increase in the basal temperature gives a warmer thermal structure resulting in a smaller curvature which can be compensated by using stronger rheologies for the lithosphere. It results in the non-uniqueness of the solutions. Due to computation time, here we only test the effect of parameters separately.

Some of these parameters have only a small influence on the final geometry of our model such as the density or the thickness of the layers. Thereby, hereinafter we fix the density contrast at the Conrad discontinuity to 100 kg.m^{-3} and at the Moho to 370 kg.m^{-3} (*Cattin et al. 2001*), the crustal thickness to 35 km (*Kumar et al. 2001*) or the thickness of the lithosphere to 100 km (*Kumar et al. 2007; Devi et al. 2011*).

In contrast, the elastic parameters, the temperature field or the viscous rheology has a greater effect on the final geometry of the India plate. Figure 3.5 shows the results for the exploration of four parameters and their influence on calculated Bouguer anomaly and Moho geometry. To compare models with the data, we calculate a χ^2 misfit following this formula :

$$\chi^2 = \sum_{i=1}^n \left(\frac{G_{data_i} - G_{model_i}}{\sigma G_i} \right)^2 + \alpha \sum_{k=1}^m \left(\frac{M_{data_k} - M_{model_k}}{\sigma M_k} \right)^2 \quad (3.5)$$

where G and M correspond to gravity anomalies and geometry of the Moho, respectively. n and m , the number of data for gravity and receiver functions and σ their associated error. We use a coefficient, $\alpha=20$, in order to balance the weights between gravity and receiver function misfits.

The basal temperature has a strong control on the flexural rigidity of the plate : higher temperatures lower the flexural rigidity. Our results also suggest that the Young modulus has a first order control on the final geometry below the threshold value of $\sim 80 \text{ GPa}$ (Fig. 3.5) .

Using various viscous rheologies proposed by *Ranalli (1995)*, our results demonstrate that, compared to the mantle, the crustal rheology has a minor effect on the shape of the flexure. Furthermore our results suggest that compared to a dry mantle, a wet mantle leads to a smaller curvature.

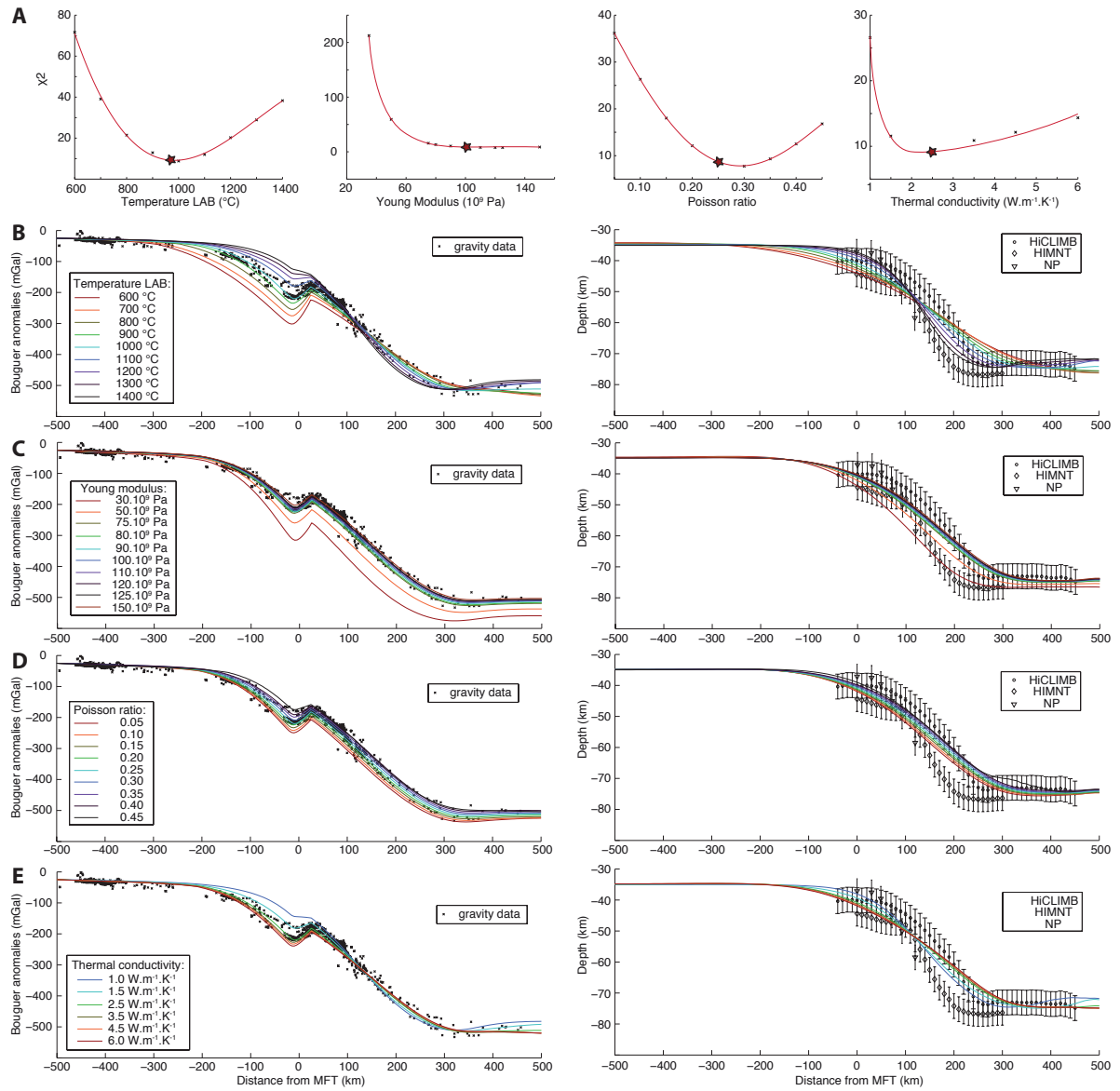


FIGURE 3.5 Effect of temperature at the lithosphere-asthenosphere boundary (LAB), Young modulus, Poisson ratio and thermal conductivity. (A) χ^2 function of the different tested parameters. The red star corresponds to our preferred model. (B) Effect of the LAB temperature on our models. On the left, the colored lines and crosses correspond to the calculated Bouguer anomaly and to the data, respectively. On the right are shown in color the Moho geometries obtained in our models and in black the constrained from receiver function profiles. (C) Same as B. for Young modulus. (D) Same as B. for Poisson ratio. (E) Same as B. for thermal conductivity.

In our preferred model, we use a basal temperature of 977°C, a upper crustal heat production of $2.5 \mu\text{W.m}^{-3}$ and a thermal conductivity of $2.5 \text{W.m}^{-1}.\text{K}^{-1}$. The calculated surface heat flow from this thermal structure is about 60mW.m^{-2} , consistent with the measurements made in cratonic areas of northern India (*Pandey and Agrawal 1999*). We obtain the best fit with the parameters described in Table 3.2, which correspond to a weak upper crust, strong lower crust and strong mantle.

	Upper Crust (Quartz)	Lower Crust (Diabase)	Upper Mantle (Olivine)
ρ (kg.m^{-3})	2800	2900	3270
E (GPa)	100	100	100
ν	0.25	0.25	0.25
c (MPa)	10	10	10
ϕ	30°	30°	30°
γ_0 ($\text{Pa}^{-n}.\text{s}^{-1}$)	6.31×10^{-25}	6.31×10^{-20}	7×10^{-14}
n	2.90	3.05	3.00
E_a (kJ.mol^{-1})	149	276	510

TABLE 3.1 Rheological parameters used for the thermomechanical modelling. ρ , density; E , Young modulus; ν , Poisson’s ratio; c , cohesion; ϕ , internal friction angle; γ_0 , power-law strain rate; n , power-law exponent and E_a , power-law activation energy.

3.2.5.1 Consistency with seismological data

The Moho geometry of our best fit model is consistent with the constraints given by HIMNT and Hi-CLIMB project (*Schulte-Pelkum et al. 2005; Nabelek et al. 2009*) as well as with the new receiver functions profile obtained at the longitude of Pokhara (Fig. 3.6b). This model suggests a smoothly bent geometry for the Moho, which starts to bend down with a low angle at about -250 km distance in our profile. Its dipping angle gradually increases underneath the Ganga basin to reach the steepest dip of 11° northward beneath the High Himalaya. Then the Moho flattens at 300 km.

Between the onset of the flexure of the India plate and the location of the MFT, the top of our model corresponds to the basement of the Ganga basin. It is also in good agreement with the constraints on the depth of the basement from receiver functions (*Hetényi et al. 2006*), from a deep well (*Sastri et al. 1971*) and from a basement depth contour map of

the Gangetic plains (*Raiverman et al. 1983*) showing a maximum depth of 6 km and a mean horizontal extent of ~ 250 km.

More importantly our results suggest that a single lithosphere rheology type explains Moho depth constraints over the lateral distance of 350 km studied here. Minor lateral variations of Moho depth seen on RF constraints could not be explained by numerical models using homogeneous layer properties tested here, which likely points to small along-profile heterogeneities in physical properties. Nevertheless, the gross picture is satisfactory, although better seismological coverage could improve the set of constraints. Next we analyze gravity data to test the hypothesis of lateral similarity of India plate strength from Central to Eastern Nepal.

3.2.5.2 Long-wavelength gravity data and effective elastic thickness of the India plate

We use the final geometry of the modelled India plate layers to calculate the synthetic Bouguer anomalies and compare them to the observations along the four gravity profiles. We assume the same density contrasts as in our thermomechanical model for the India plate : 100 kg.m^{-3} at the Conrad discontinuity and 370 kg.m^{-3} at the Moho. For consistency between thermo-mechanical modeling and gravity modeling, the density of the Tibetan crust and sediment in the foreland basin are the same than those used in the calculation of the load applied on top of the India plate (Fig. 3.6b). The synthetic Bouguer anomalies are computed following the approach proposed by *Won and Bevis (1987)*.

The calculated long wavelength anomalies are in fairly good agreement with the gravity data except beneath the Tibetan plateau, where they are lower than the observed data by ~ 70 mGal (Fig. 3.6). Several studies argue in favour of the eclogitization of the Indian lower crust beneath Tibet (e.g. *Henry et al. 1997*; *Schulte-Pelkum et al. 2005*; *Tiwari et al. 2006*; *Monsalve et al. 2008*) and as proposed by *Hetényi et al. (2007)*, eclogitization presumably occurs as it reaches its maximal depth, *i.e.* at the location where we observe the misfit between calculated and observed gravity data. The fit is improved by increasing the density of the Indian lower crust beneath Tibet for $\Delta\rho = 150 \text{ kg.m}^{-3}$ with respect to the Moho. The final model gives consistent results with the gravity data on a large scale along all four profiles, confirming the seismological result of lateral similarity of the India plate between the longitude of Pokhara and the Arun river (Fig. 3.6).

The estimate of the flexural rigidity of the India plate remains highly variable (e.g.

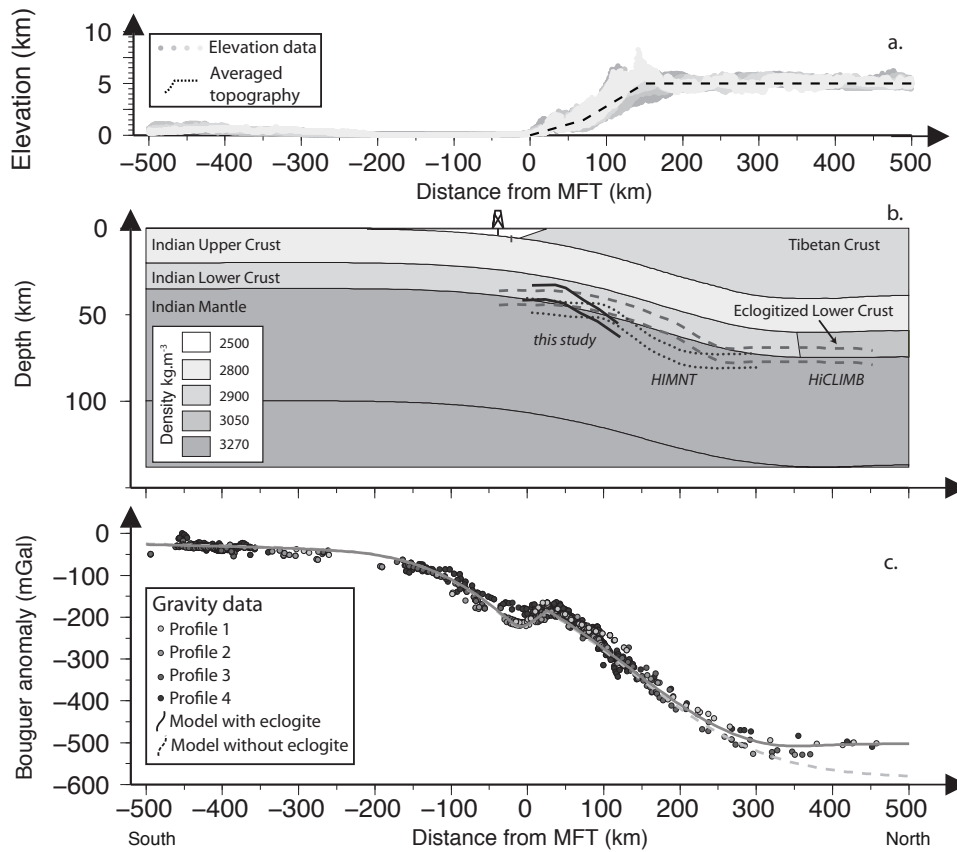


FIGURE 3.6 Topography, preferred flexural model and the associated Bouguer anomaly compared with observed data along the four profiles. (a) Topography on a 25 km wide swath for the four profiles. Dashed black line corresponds to the averaged topography used in the calculation of the subsurface load. (b) Final geometry of the thermo-mechanical model after 10 Myr. To explain the long wavelength anomaly, we increase the density of the Indian lower crust that is underthrusting Tibet where eclogitization occurs. Different geometrical constraints from published receiver functions at different profiles as well as from the depth constraint of the Raxaul deep well are also shown. (c) Circles represent the Bouguer anomaly along the four profiles (see fig. 3.1 for location of the profiles). The thick black line corresponds to the calculated gravity anomaly associated with the flexural model below, the dashed one correspond to the model without eclogite.

Karner and Watts 1983; Lyon-Caen and Molnar 1983; Burov and Watts 2006; Hetényi et al. 2006; Chamoli et al. 2011; Hammer et al. 2013). This variability can be related to the used approach (forward modelling vs. spectral methods), but it can also reflect some real lateral variations. Taking advantage of our new gravity and seismological informations we address the question of the strength of the India lithosphere, *i.e.* its flexural rigidity, in Central and Eastern Nepal. To allow some comparison with thin elastic plate models, our model can be interpreted in terms of the equivalent elastic thickness (EET). The depth-varying rheology leads to the development of weak layers within the lithosphere and decoupling (Fig. 3.7), responsible for variations of the mechanical rigidity of the Indian lithosphere. For a plate decoupled into n layers, the equivalent EET of the plate is given by (e.g. *Burov and Diament 1992*) :

$$EET = \left(\sum_{i=1}^n h_i^3 \right)^{1/3}, \quad (3.6)$$

where h_i is the thickness of the elastic core for the i^{th} layer. The deformation is considered elastic when the relaxation time τ of the material is twice as long as the time of the numerical experiment with :

$$\tau = \frac{\nu}{E} \quad (3.7)$$

where ν is the effective viscosity and E the Young's modulus. Here we assume that a region will be elastic when the effective viscosity is larger than $\sim 2 \cdot 10^{25}$ Pa.s for a modeled deformation age of 10 Myr. The calculated effective viscosity from our thermomechanical model is thus used to assess the EET associated to a layered India plate with a ~ 35 km thick crust and a Conrad discontinuity at 20 km depth.

Our results suggest a flexural rigidity decrease of the India plate from ~ 75 km in the south of the profile at 24°N to ~ 25 km beneath the Tibetan Plateau at 32°N (Fig. 3.7). This decrease has been already suggested by *Karner and Watts (1983)* by implicitly breaking the elastic plate at the suture. The EET of the upper and lower crust decreases beneath the Ganga Basin and vanishes beneath the Himalayas, 100km North of the MFT, at about 27.5°N . The EET of the mantle decreases slowly along the profile from a value of 40 km to 25 km. It suggests that the mantle rheology is the prime control on India plate flexure (*Karner et al. 1983*). Coupling between crust and mantle exists until -250/-350 km distance. North of $\sim 25^\circ\text{N}$, the three layers are decoupled. This results in a non-linear EET decrease along the profile : in the southern part all layers are coupled and the India

plate has an EET of ~ 75 km. Further north the successive decoupling of the base of the upper and lower crust leads to a ~ 100 km long transition zone where the EET drops to ~ 50 km. All the layers are decoupled at about -250 km distance where the EET drops to ~ 35 km, north of which it decreases slowly to reach ~ 25 km beneath Tibet.

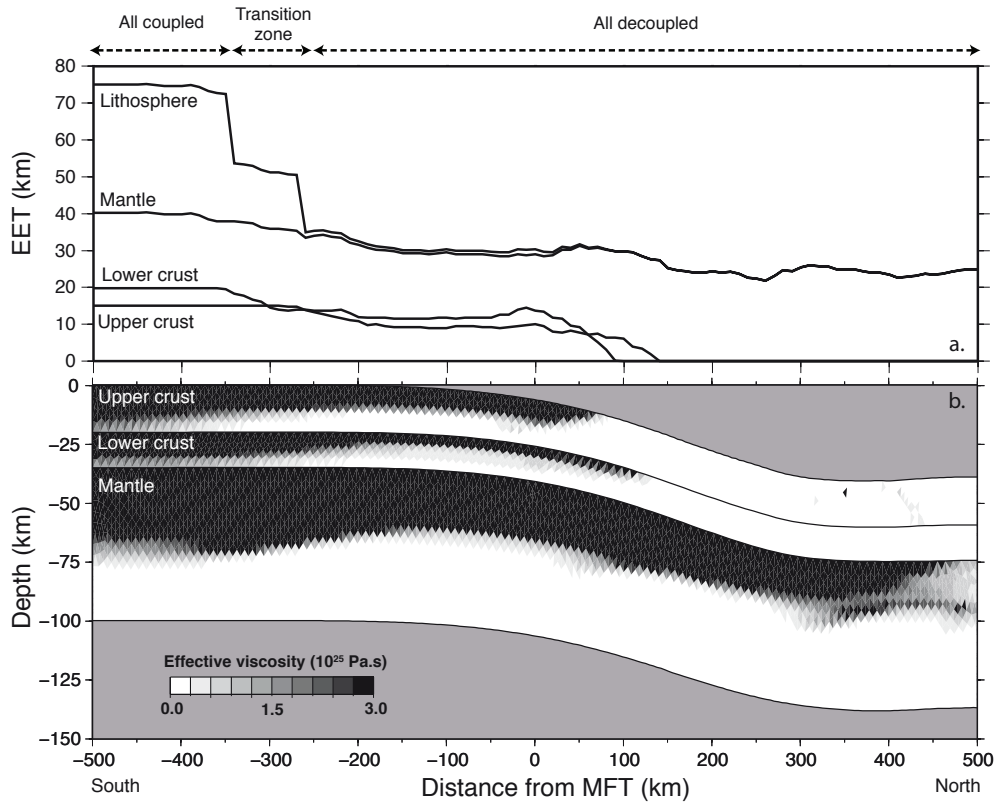


FIGURE 3.7 Effective viscosity of the preferred model and the corresponding Effective Elastic Thickness (EET). (a) EET for each layer and for the whole lithosphere. Depending on the coupling of the layers, EET is ~ 75 km South of our profile and decreases northward because of crustal thickening, temperature increase and decoupling to reach ~ 25 km to the North. (b) Effective viscosity plotted for each layer composing the India plate. Dark areas correspond to the strong and elastic part of the lithosphere whereas light one represents the weak and ductile part.

3.2.5.3 Lateral variations of short wavelength gravity anomalies and subsurface structures

The geometry of the India plate and its associated strength presented in Fig. 3.6 and Fig. 3.7 accounts for the regional trend of the gravity data along the four profiles. However, small differences between these profiles and our forward modeling exists, especially in the Ganga basin and in the Himalayan section (Fig. 3.8a). In this figure, we removed the

gravity effect of the low density sediments in the Ganga Basin from our model to enhance its variability in the residuals of the four profiles. We can notice two main features in these residuals, a gravity low south of the MFT with a maximum value of about -100 mGal and a gravity high of 50 mGal observed at ~ 125 km north from the MFT. The gravity low south of the MFT corresponds to the effect of accumulated sediments in the Ganga Basin (*Lyon-Caen and Molnar 1983*). It exhibits a decrease both in width and amplitude from West to East. This can be easily explained by the rough topography of the basement and especially by the presence of the Munghyr Saharsa ridge near by the easternmost profile 4. Regarding the gravity high, its location lies at the TSS and should be related to the thick sedimentary cover. Here, part of the signal in the Bouguer anomaly could be linked with an overestimated reduction density compared to the rest of the Himalayas.

Short wavelength gravity data along profile 2 have already been interpreted in terms of density contrast of foreland sediments, of Lesser Himalayan meta-sediments and of Palung granites (*Cattin et al. 2001*). Here we focus on profile 1 that corresponds to our new gravity data to show that a simple gravity modeling of subsurface structures fits the two areas of short-wavelength misfit. For the Ganga Basin, as already shown, we assume a density contrast of -420 kg.m^{-3} with respect to the rest of the model and a simplified geometry : a straight south-dipping line starting from the surface location of MBT to the top of the India plate at depth. The TSS is assumed to extend to 8 km depth (*Zhang and Klemperer 2010*) with a low angle northward dip of the South Tibetan Detachment (STD) (*Searle 2010*) and a density contrast of -100 kg.m^{-3} , corresponding to the value used by *Tiwari et al. (2006)* considering the density of 2670 kg.m^{-3} for the rest of our model. The observed and calculated Bouguer anomalies are in very good agreement, with a maximum and an average difference of 10 mGal and 2 mGal, respectively (Fig. 3.8b). It thus turns out that the small lateral gravity variations that remain in our four profiles once the flexural support of the range is taken into account, are linked to variations of geometry and density of subsurface structures.

3.2.6 Discussion

The new thermomechanical modelling carried out in this study, together with a new set of gravity and seismological data, allows us to re-estimate the EET of the India plate in Central and Eastern Nepal. Regarding the similarity in the long wavelength anomalies

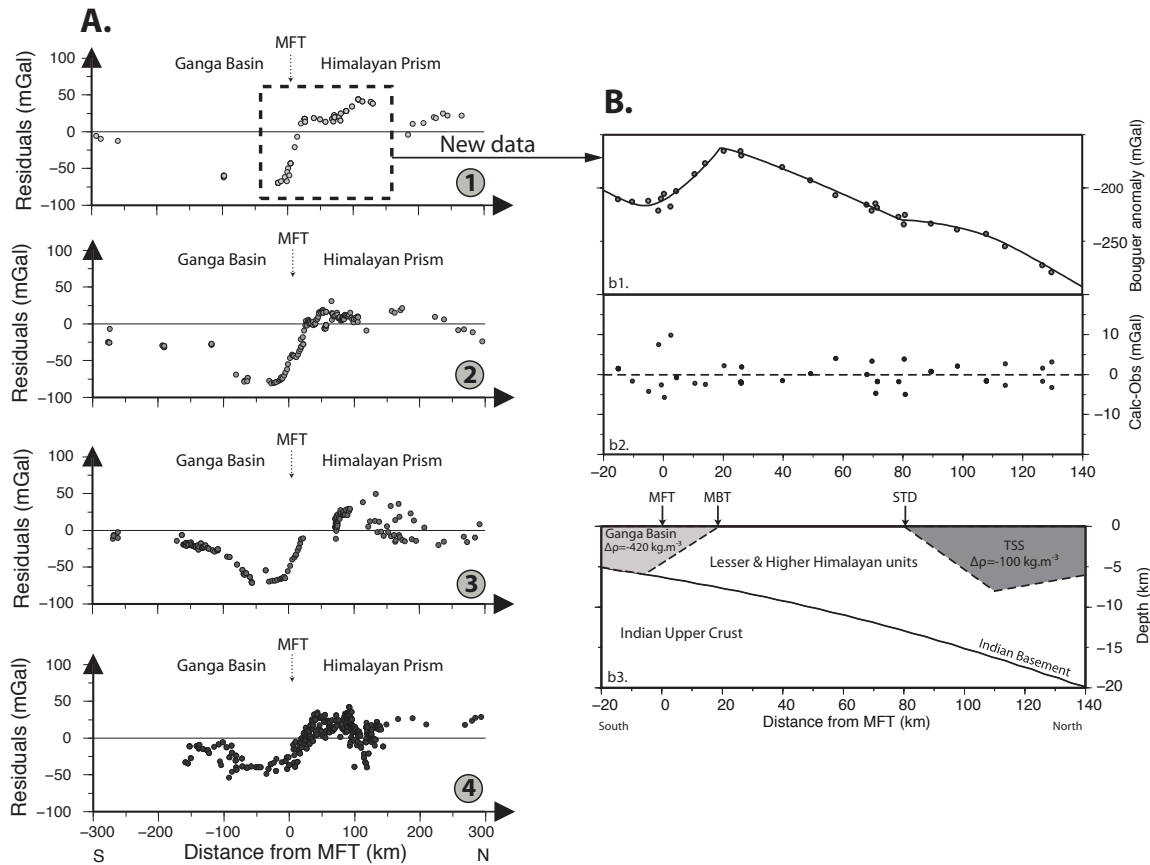


FIGURE 3.8 (A) Bouguer anomaly residuals for the four studied profiles calculated by subtracting the synthetic Bouguer anomaly associated to the thermomechanical model (without the gravity effect of the Ganga basin) from the observed data. (B) Flexural model and associated Bouguer anomaly centred on the new dataset (profile 1) to adjust the residuals at short wavelength. (b1) Calculated (black line) and observed (circles) Bouguer anomalies. (b2) Calculated minus observed gravity anomalies. (b3) Density model with flexural geometry of the final model. The Ganga basin is ascribed a density contrast of -420 kg.m^{-3} , i.e., 2250 kg.m^{-3} considering an upper crustal density of 2670 kg.m^{-3} . The Tethyan Sedimentary Series is assumed to extend to 8 km depth (*Zhang and Klemperer 2010*) with a low angle northward dip of the STD (*Searle 2010*) and a density contrast of -100 kg.m^{-3} in agreement with the 2570 kg.m^{-3} found by *Tiwari et al. (2006)*.

along the four study profiles, it appears that there are no major lateral variations of the India plate strength in this ~ 350 km wide area. The obtained EET profile is in agreement with the result of *Hetényi et al. (2006)* along a profile near Kathmandu. Our result is also consistent with the analytical solutions for the flexure of an elastic plate found by *Lyon-Caen and Molnar (1983)*, *Karner and Watts (1983)*, *Jin et al. (1996)* and *Burov and Watts (2006)*. *Jordan and Watts (2005)* suggest variations in elastic thickness along strike of the Himalaya East of 87°E , our results are in agreement with their findings for the Central Himalayas (western Nepal) which suggests the India plate mechanical behaviour is similar all along Nepal. They also predict a northward decrease of the EET from ~ 100 km in Central India to ~ 25 km beneath Tibet, also in agreement with our findings. In contrast, *McKenzie and Fairhead (1997)* and *Jackson (2002)* suggest a EET of ~ 40 km south of the MFT which is lower than our estimate. However a detailed analysis of their results suggests that the minimum value of EET from their misfit function is poorly constrained.

Even with the simplified geometry used for the northern end of the Ganga basin, the short wavelength anomaly analysis presented in figure 3.8b fits the observed data well. Such a modelling has already been performed along the Kathmandu profile by *Hetényi et al. (2006)* with the same density of sediments. A detailed forward gravity modelling has been realized by *Cattin et al. (2001)* along profile 2. They model the Ganga Basin with a northward dipping end with addition to the LH units but they do not include the TSS even if it lies all along the study area (*Zhang and Klemperer 2010; Zhang et al. 2012*). The explanation is that the STD is further north in profile 2 than in profile 1 and that there are no data at this location on profile 2. Regarding the gravity low associated to the Ganga Basin in the four profiles (Fig. 3.8), we can observe that its width decreases to the East, which is consistent with the isocontour map of the depth of the Ganga basin (*Raiverman et al. 1983; Gahalaut and Kundu 2012*). Finally, the low in profile 4 is not as regular as in the other profiles which can be associated to the highly fractured basement at this location due to the basement high corresponding to the northern prolongation of the Satpura massif.

3.2.7 Conclusion

Using both geological and geophysical information as well as new gravity and seismological datasets we have addressed the question of lateral variations in the India plate strength across the Himalayas over Central and Eastern Nepal. Our new measurements and the available Bouguer anomalies allow obtaining four profiles, each 1000 km long, perpendicular to the strike of the orogen (oriented N16° E) and centred on the MFT. These profiles cover a 350 km wide area in Central and Eastern Nepal between the Pokhara valley and the Arun river. Following *Cattin et al. (2001)* and *Hetényi et al. (2006)* we carry out thermomechanical modelling to interpret jointly these data in terms of flexural rigidity of the Indian lithosphere and crustal structures. Compared to these two previous studies the numerical modelling is improved by considering a more realistic vertical loading including spatial and temporal variations.

Our model fits both seismological and gravity data suggesting that these geophysical observations are compatible with the assumed material properties of crustal and mantle rocks. Our main result is that in the study area no major lateral variations of the strength of the India plate are needed to fit the seismological constraints and gravity anomalies. This suggests that a homogeneous India plate rheology with a strong mantle acts in the India-Eurasia collision in Central and Eastern Nepal. This finding is in agreement with the recent result obtained by *Grandin et al. (2012)* showing no variation in the interseismic loading between the front of the Annapurna mountain range (near Pokhara) and the Kathmandu valley. This is important in terms of seismic hazard assessment, because it promotes lateral continuity of the state of stress along a ~ 350 km wide zone that may control the seismological potential of this area.

The long wavelength variation of the gravity anomaly in the four study profiles is the same. Our result confirms that the main trend of the gravity profile reflects the deepening of the Moho from 35 km to 75 km and a (partially) eclogitized Indian lower crust beneath Tibet (*Hetényi et al. 2007*). Our model also predicts a northward decrease of the EET of the India plate with three successive stages, with an EET of 75 km, 50 km and 35-25 km, respectively associated to an Indian lithosphere that is totally coupled south of the foreland, partially coupled in the foreland and totally decoupled beneath Tibet. At shorter wavelengths, the new dataset can be explained by simply adding density contrasts corresponding to the Ganga Basin and to the Tethyan Sedimentary Sequence. The lateral

variations in the gravity low attributed to the Ganga Basin reflect along-strike variations of basement depth.

The EET of the India plate inferred from Bouguer anomaly profiles acquired in Sikkim (*Tiwari et al. 2006*), in Northwest Himalaya (*Chamoli et al. 2011*), as well as in Bhutan (*Hammer et al. 2013*) seem to be different from those obtained in Nepal. Further geophysical observations along missing portions of the Himalayan arc (Western Nepal, North-Eastern India) are thus now needed to constrain the lateral variations in three dimensions.

Acknowledgments

This work was supported by grants from INSU-ALEAS and CNES-TOSCA. People from the Department of Mines and Geology as well as the drivers are thanked for their help in the field survey. Most of the figures of this article were prepared using the Generic Mapping Tools software (*Wessel and Smith 1995*). We finally thank three anonymous reviewers for their constructive comments on the article.

CHAPITRE 4

COMPORTEMENT FLEXURAL DE LA PLAQUE

INDIENNE SOUS LE BHUTAN

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4.1 Présentation

Grâce aux données gravimétriques acquises au Bhoutan dans le cadre de cette thèse, le chapitre qui suit présente les premières contraintes en terme de géométrie flexurale de la lithosphère Indienne sous le Bhoutan. Le résultat principal est que la longueur d'onde de la flexure est plus courte dans l'Est Himalaya qu'au Népal.

En utilisant le modèle thermo-mécanique décrit dans le chapitre précédent, nous montrons que cette différence nécessite une rigidité flexurale de la plaque indienne plus faible au Bhoutan. Cette variation latérale de la géométrie flexurale de la plaque indienne ne peut être uniquement expliquée par la chargement différentiel lié aux variations latérales de topographie entre le Bhoutan et le Népal.

Nous proposons deux modèles qui permettent d'expliquer les données géophysiques dans cette partie de l'arc Himalayen. Ces deux modèles suggèrent une diminution de la résistance du manteau lithosphérique Indien : dans un cas grâce à l'utilisation d'une rhéologie mantellique hydratée, et dans l'autre à une température à la base de la lithosphère plus élevée que dans le modèle Népalais. Ce chapitre se présente sous la forme d'un article publié à *Geophysical Research Letter*.

4.2 Flexure of the Indian Plate underneath the Bhutan Himalaya

*Paper published in Geophysical Research Letter*¹

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Abstract

We investigate flexural geometry and rheology of the India Plate beneath the Eastern Himalaya from a new gravity dataset acquired in Bhutan. Compared to the well-studied Nepal-Himalaya the obtained Bouguer anomaly profiles across the range show shorter wavelength flexure of the lithosphere with a narrower and shallower foreland basin. This new dataset and seismic Moho constraints are interpreted together in terms of lithospheric flexure using a 2D thermo-mechanical model. Our results suggest that the strongest layer of the continental lithosphere beneath Bhutan is the upper mantle, as it is beneath Nepal. The observed West-to-East decrease in flexural wavelength is associated with weakening mantle rheology. The simulations show that this decrease can be related to ductile mantle behavior : either hydrated wet dunite or dry-and-hot olivine rheology. Both models display decoupled lithospheric layers leading to an eastward decrease of flexural rigidity from ca. 10^{24} N m to ca. $5 \cdot 10^{22}$ N m in Nepal and Bhutan, respectively.

4.2.1 Introduction

The Himalayas stretch over 2400 km between the lowland of the Indian Shield and the highland of the Tibetan Plateau. At depth, the orogen marks the transition between regions of respectively regular and thickened crust. It is now well accepted that the shape of this transition is associated with regional isostasy that reflects the mechanical properties of the India plate bent beneath the mountain belt. Regional isostasy is well imaged and understood at the longitude of Kathmandu, where the upper mantle is the strongest layer of the lithosphere holding up the weight of both the Himalayas and the Tibetan Plateau (*Lyon-Caen and Molnar 1983; Karner and Watts 1983; Cattin et al. 2001; Hetényi et al. 2006; Nabelek et al. 2009*).

Although the tectonic units are remarkably uniform along the 2400 km long shape of the belt (*Schelling 1992; Hodges 2000*), there are hints of lateral heterogeneity. An East-West transition is indicated by different elevation profiles across the range (*Duncan et al. 2003*), a reduction in crustal shortening (*McQuarrie et al. 2008; Long et al. 2011*) and patterns of focal mechanisms (*Hazarika et al. 2010*). At depth, although seismic tomography reveals spatial variation in the mantle structure (*Li et al. 2008*), there is yet no image of the Moho's shape beneath the Eastern Himalaya (East of Sikkim). Here, following the approaches of *Lyon-Caen and Molnar (1985)*, *Duroy et al. (1989)* and *Berthet et al. (2013)* we focus on variations between Nepal and the Eastern Himalayas of Bhutan.

After a brief presentation of the new gravity data collected in Bhutan we discuss the obtained Bouguer anomaly profiles across the Eastern Himalaya spanning from India to Tibet. Next, together with regional Moho depths available from seismology, we interpret this dataset in terms of flexural rigidity using 2D thermo-mechanical models simulating the bending and the rheology of the Indian lithosphere. Finally, from a comparison of the results to those in Nepal, we provide a first assessment of along-strike variations of flexural rigidity over the Central and Eastern Himalayas (Fig .4.1a).

4.2.2 Gravity data

Until now the Kingdom of Bhutan was a blank spot on the gravity data coverage map. To fill this data gap we acquired new gravity data at 214 locations. Measurements were conducted with a *Scintrex CG5* relative gravimeter in 2010-2012 along all major roads in Bhutan.

Following *Berthet et al. (2013)* gravity measurements and dual-frequency GPS recordings are subjected to a standard processing scheme. The resulting Bouguer anomalies are computed using a reduction density of 2670 kg m^{-3} and include terrain corrections ranging from 5 to 80 mGal. The internally consistent but relative dataset's accuracy is ca. 3 mGal. Due to the lack of an absolute reference point the dataset is shifted to match the neighboring Bouguer anomalies in NE-India and S-Tibet. An error bar of ± 10 mGal is thus assigned to the new dataset.

Finally, these new Bouguer anomalies are combined with available Bouguer anomalies compiled by *Hetényi et al. (2007)* to construct continuous profiles across the Himalayan range, covering entirely the flexure of the India plate underneath the Eastern Himalaya from the lowlands to the southern part of the Tibetan Plateau (Fig. 4.1).

4.2.3 Bouguer anomalies across Bhutan

Bouguer anomalies are projected along 1000-km long, arc-perpendicular profiles centered on the Main Frontal Thrust (MFT). Two profiles in E- and W-Bhutan are investigated, and compared to profiles in E-Nepal and in the Sikkim Himalaya to decipher along-strike variations (Fig. 4.1b).

Two distinct wavelengths can be identified on all profiles : the 1st order anomaly is characterized by a long-wavelength step-function of ca. 500 mGal over a distance of 500-600 km as measured between the kink-points of ca. -50 to -20 mGal in the South in India and ca. -520 mGal in the North on the Tibetan Plateau. Hence there is a roughly uniform northward gradient of -0.8 to $-1.0 \text{ mGal.km}^{-1}$, reflecting the increase in crustal thickness. Superimposed on this 1st order anomaly is a 2nd order anomaly that defines an intermediate step near the MFT (0 km horizontal distance on Fig. 4.1b).

The 1st and 2nd order anomalies exhibit along-strike variations between E-Nepal and Bhutan. Most significant is a decrease in wavelength of the step-function from West (ca. 600 km) to East (ca. 500 km) (Fig. 4.1b). Lateral variation is further indicated by the level of the step superimposed on the linear gradient, shallowing West-to-East from -200 mGal to -150 mGal. This is accompanied by a ca. 100-km decrease of the distance between the southern kink-point and the MFT. Hence Bouguer anomalies in Bhutan start to decrease further to the North relative to the position of the MFT, and the step near the MFT that is clearly observable in Nepal and Sikkim is narrower and of lower amplitude. The Sikkim

profile represents a transition between those in Nepal and Bhutan from all perspectives (Fig. 4.1b).

These observed lateral changes in Bouguer anomalies are interpreted as the following :

1. The shorter distance between the kink-points at the longitude of Bhutan is a clear indication of shorter wavelength lithosphere flexure with respect to that in Nepal.
2. In accordance with Bouguer anomaly profiles in Nepal (*Cattin et al. 2001; Hetényi et al. 2006*), where both seismic and borehole data constrain the geometry of the sedimentary basin, the step in the Bouguer anomalies near the MFT south of Bhutan is attributed to a low-density sedimentary infill of the foreland basin. The extent and depth of the foreland basin in the East (Brahmaputra) is however smaller than that in the Central Himalayas (Ganges), as already pointed out by *Lyon-Caen and Molnar (1983)* and *Gansser (1984)* and shown on *Zhang (1983)*.

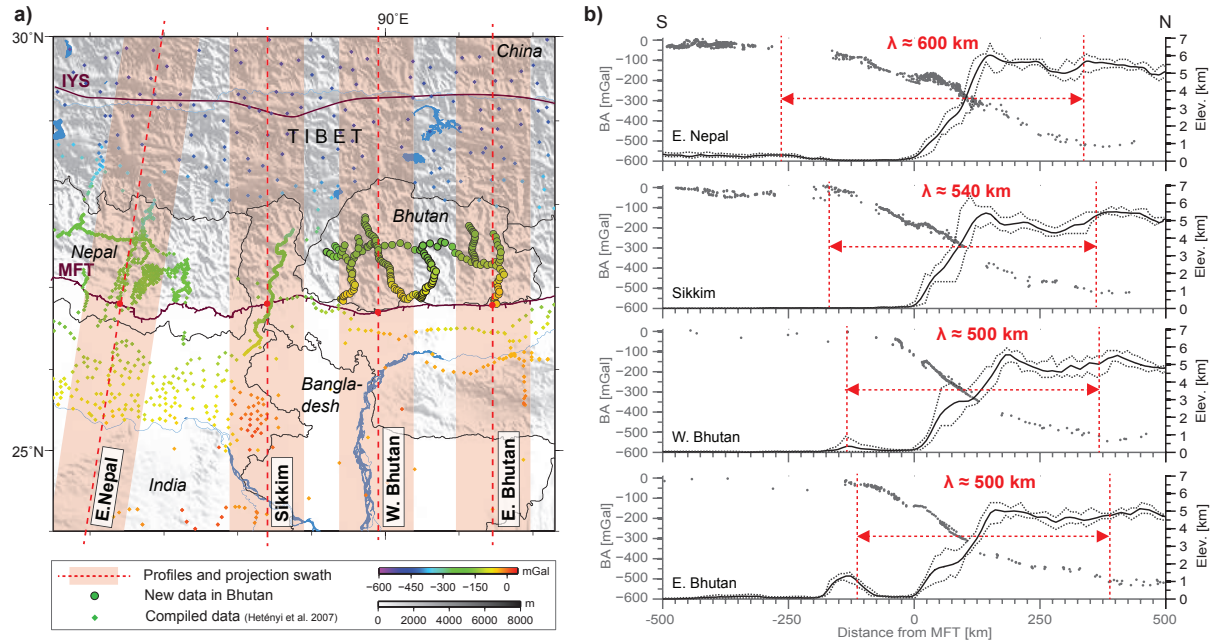


FIGURE 4.1 Setup of the thermo-mechanical model and time evolution of the load. (a) The load is composed of three spatial pressure regimes : a fixed load in Tibet $P_{fix}(x, t) = g(\rho_{crust}h_{crust} + \rho_{topo}h_{topo})$ and two variable loads that are recalculated after each time step to account for the changing thickness of the overlying Tibetan crust $\Delta P_{crust}(x, t) = g(\rho_{crust}h_{crust} + \rho_{topo}h_{topo})$ and of the foreland sediments $\Delta P_{sed}(x, t) = g\rho_{sed}h_{sed}$. Sedimentation in the Brahmaputra Basin is assumed to maintain a flat foreland at a constant zero elevation. The load is progressively moved to the South with an advance rate of 15 mm yr^{-1} . For isostatic balance the model is supported by a hydrostatic pressure at its base. Free vertical displacements are allowed at the sides of the model. The upper crust, the lower crust and the lithospheric mantle have different rheologies. The standard initial temperature distribution is shown. (b) Time evolution of the loads and of the propagation velocity v_{prop} .

4.2.4 Interpretation with lithosphere flexure models

4.2.4.1 Thermo-mechanical modeling

To explore the implications of the shorter flexural wavelength on lithosphere rheology at the longitude of Bhutan, we take advantage of the finite element modeling tool *ADELI* (*Hassani et al. 1997*). This 2D thermo-mechanical approach accounts for the visco-elasto-plastic behaviour of the lithosphere and its dependence on temperature and depth. The thermal structure of the underthrusting India plate is approximated by an analytical expression developed by *Royden (1993)*. A detailed description of the material laws and the thermal setup can be found in *Cattin et al. (2001)*, *Hetényi et al. (2006)* and *Berthet et al. (2013)*, who successfully modeled the flexure of the India plate beneath Nepal using constraints from gravity and seismology.

Following their approach, the initial geometry and density setup of the model was compiled using available controlled source as well as passive seismic results and velocity-density relations (Fig. 4.2). During the simulations, the Indian lithosphere is bent down by the advancing (15 mm.yr^{-1}) load of the Tibetan crust and sedimentation in the foreland basin. The Shillong plateau south of the arc is not included in the model due to the poorly constrained time evolution and its unconstrained deep structure. A parametric study is performed assessing the influence of elastic, ductile and thermal parameters on flexure wavelength. The non-uniqueness of the model is reduced by departing from a well constrained model in central Nepal (*Cattin et al. 2001*) and by developing two contrasting end-member models that define a parameters space within which different combinations of setups likely lead to the same results (shape, flexural rigidity). Geometries produced by the model are evaluated by comparing synthetic and observed Bouguer anomalies and by cross-checking the flexural geometry with available regional seismic Moho constraints. Searching for the best fit models through 90 model simulations results in an evaluation of lithosphere rheologies of the India plate and produces two preferred models. Detailed information on the model setup, the range of values in the rheology parametric study and the compilation of regional Moho constraints is presented in the Supporting Material.

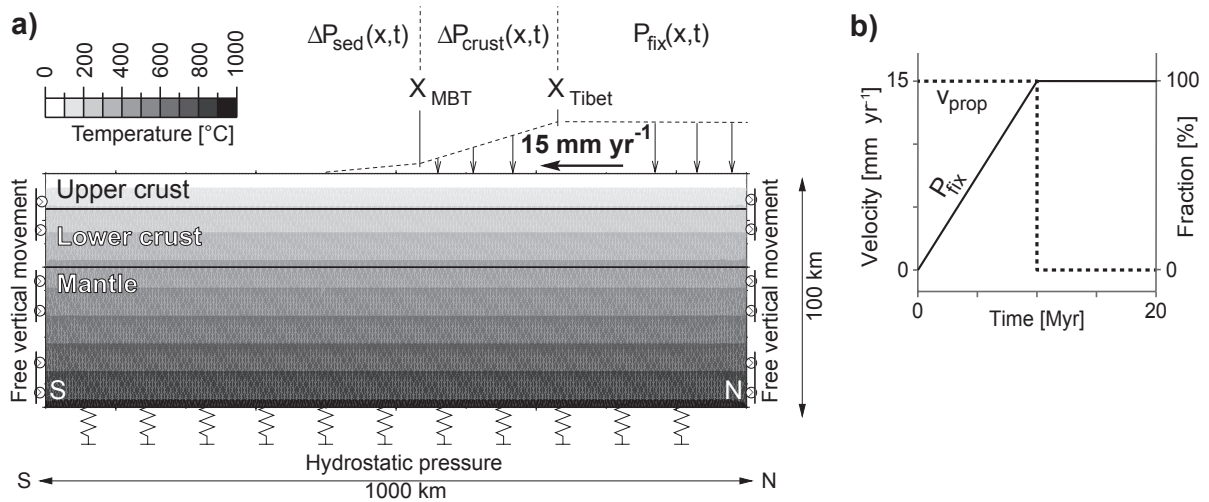


FIGURE 4.2 Setup of the thermo-mechanical model and time evolution of the load. (a) The load is composed of three spatial pressure regimes : a fixed load in Tibet $P_{fix}(x,t) = g(\rho_{crust}h_{crust} + \rho_{topo}h_{topo})$ and two variable loads that are recalculated after each time step to account for the changing thickness of the overlying Tibetan crust $\Delta P_{crust}(x,t) = g(\rho_{crust}h_{crust} + \rho_{topo}h_{topo})$ and of the foreland sediments $\Delta P_{sed}(x,t) = g\rho_{sed}h_{sed}$. Sedimentation in the Brahmaputra Basin is assumed to maintain a flat foreland at a constant zero elevation. The load is progressively moved to the South with an advance rate of 15 mm yr^{-1} . For isostatic balance the model is supported by a hydrostatic pressure at its base. Free vertical displacements are allowed at the sides of the model. The upper crust, the lower crust and the lithospheric mantle have different rheologies. The standard initial temperature distribution is shown. (b) Time evolution of the loads and of the propagation velocity v_{prop} .

4.2.4.2 Preferred models

The parametric study reveals that the creeping strength of the lithospheric mantle and the temperature field (controlled by the input temperature at the lithosphere-asthenosphere boundary (LAB)) are the two key factors that induce a shorter wavelength flexure. The ductile parameters of the crust as well as elastic material properties have a comparatively smaller influence.

The two preferred models respectively use (1) a relatively weaker wet dunite rheology for the lithospheric mantle in a normal thermal regime (1000 °C at the bottom of the model), and (2) a relatively stronger dry olivine rheology in a hotter (+150 °C) thermal regime (see parameters in Tab.4.1). Both of these models match the 1st and 2nd order Bouguer anomaly variations and produce Moho depths coherent with the available seismological constraints. The shape of the intermediate step near the MFT is well fitted by a foreland basin of smaller extent and shallower depth compared to Nepal (Fig. 4.3).

Both preferred models induce a shorter wavelength flexure by altering the rheology of the lithospheric mantle compared to models developed for Nepal (*Cattin et al. 2001*; *Hetényi et al. 2006*). Since the lithospheric mantle has a key role in supporting the weight of the Himalayas and the Tibetan Plateau, such lateral changes from the Central to the Eastern Himalaya therefore mark an important lithospheric boundary between the two regions.

We note that a third type of model setup also produced an acceptable solution. In this setup, the southern edge of the Tibetan Plateau's load (an initial condition) was located 75 km further to the South compared to other models, including those established for Nepal. Although topography alone does not suggest such a difference with respect to Nepal, it is premature to confidently evaluate this model solution before accurate and high-resolution information on the Moho beneath Bhutan becomes available.

4.2.4.3 Flexural strength of India

The thermo-mechanical model provides information on the effective viscosity of the lithosphere. This is employed to assess the effective elastic thickness (EET) of the lithosphere, which is the thickness of an elastic plate having an equivalent flexural rigidity to that of the layered visco-elasto-plastic plate. This calculation is performed by taking the cubic average of lithospheric layer's elastic core thickness (*Burov and Diament 1995*). We

	UC	LC	LM 1	LM 2
γ_0 [$\text{Pa}^{-n}\text{s}^{-1}$]	$6.31 \cdot 10^{-25}$	$6.31 \cdot 10^{-20}$	$7.5 \cdot 10^{-17}$	$7 \cdot 10^{-14}$
n	2.9	3.05	3.3	3
E_a [kJ mol^{-1}]	149	276	444	510

TABLE 4.1 Ductile properties of the two best fit models. The rheology of the upper crust (UC) and lower crust (LC) is the same. The wet dunite lithospheric mantle model (LM1) is run with a temperature at the LAB of $T_m = 1000$ °C. The dry hot olivine model (LM2) is run with $T_m = 1150$ °C. Material laws used are described in *Cattin et al. (2001)* and *Hetényi et al. (2006)*. γ_0 and n are ductile power-law parameters, E_a is activation energy. The following parameters were the same for all three layers : Young’s modulus $E=50$ GPa; Poisson’s ratio $\nu=0.25$; cohesion $c=10$ MPa; internal friction angle $\phi=30^\circ$.

use a threshold value of $1.6 \cdot 10^{25}$ Pa s for the deformation to be considered elastic, which is equivalent to assume a characteristic viscous relaxation time of 10 Myr for a Young modulus of 50 GPa.

The EET estimates of both preferred model types reflect the northward decrease in strength of the lithosphere in response to thermal and flexural weakening under the high range and the foreland basin (Fig. 4.3). The individual elastic cores are decoupled all along the profile which results in a relatively low EET even beneath India. Here the maximum values of 25 km are reached, which is significantly less than in Central Nepal. There the lithospheric layers are coupled on the southern part of the profile yielding EET values reaching 75 km and still at 40 km where the lithosphere layers decouple (*Berthet et al. 2013*). Since these EET values were calculated with a slightly different Young modulus (and therefore threshold viscosity values) a comprehensive comparison can be made using flexural rigidity D (*Burov and Diament 1995*)

$$D = \frac{E \cdot EET^3}{12(1 - \nu^2)} \quad (4.1)$$

where ν is the Poisson’s ratio (0.25) and E is the respective Young modulus. Flexural rigidity of the India plate south of Bhutan is $6.9 \cdot 10^{22}$ N m at most, while it is $3.8 \cdot 10^{24}$ N m south of Nepal where the lithosphere layers are coupled and $5.7 \cdot 10^{23}$ N m as they decouple south of the MFT. Reverting these values to EET with the parameters used in this study yields 25 km (Bhutan), 94 km (Nepal, coupled) and 50 km (Nepal, decoupled), pointing to a significant along-strike difference in flexural strength.

The West-to-East decrease of flexural rigidity (and consequently of EET) from the

Central to the Eastern Himalaya is in agreement with the findings of *Jordan and Watts (2005)* and extends those of *Bilham et al. (2003)*. Finally, both our model solutions for the India plate beneath Bhutan presented here clearly confirm the findings for Central Nepal regarding the key role of the uppermost mantle, which remains the strongest layer of the lithosphere.

4.2.5 Conclusions

New gravity data in Bhutan provide the first constraints on the flexural geometry of the Indian lithosphere underneath this part of the Himalayas. Compared to Nepal, the lithosphere bends down over a shorter distance.

An extensive set of numerical simulations of rheology parameters yields two alternative model types that explain the available geophysical data, including gravity and seismological measurements. Compared to Nepal, both models lower the ductile strength of the lithospheric mantle : one suggests hydration (wet dunite rheology), the other elevated temperature at the base of the lithosphere (hot dry olivine rheology).

The associated effective elastic thickness estimates, displaying a relatively low maximum value (25 km) and a general northward decrease due to thermal and flexural weakening, point to decoupled lithospheric layers. Still, the strongest layer is the uppermost mantle holding up the weight of the orogen, thus demonstrating its main role in continental lithosphere strength.

Acknowledgments

The Authors would like to thank all people helping in realizing the field measurements across Bhutan : officials, drivers, people of the countryside. We are grateful to E. Kissling for advice and discussion during the ETH Master Thesis from which this paper resulted. We are also thankful for the constructive comments by P. Molnar and two anonymous reviewers. We are indebted for the seed funding granted by the North-South Centre and ETH Zurich.

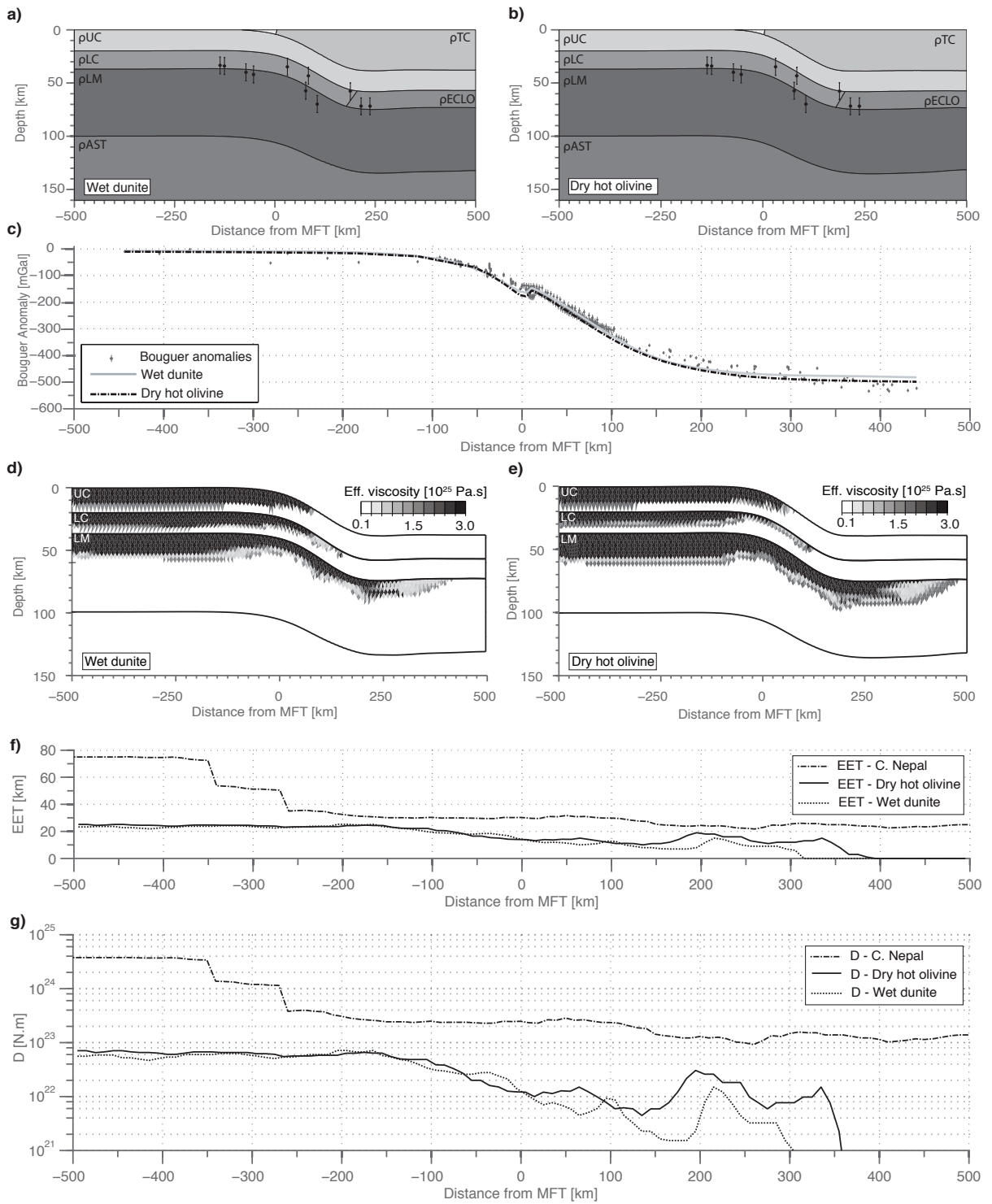


FIGURE 4.3 Preferred model results. (a,b) Resulting geometry with projected seismic Moho depths (dots with error bars) and density structure. Densities as in Table SM1 and $\rho_{ECLO} = 3050 \text{ kg m}^{-3}$. (c) Synthetic (lines) and observed Bouguer anomalies (dots with error bars). (d,e) Effective viscosity profiles. (f,g) Effective elastic thickness and flexural rigidity of the preferred models and of the Nepali model (*Berthet et al. 2013*) for comparison.

	UC	LC	LM	AST	TC	SED
Depth range (<i>km</i>)	0 – 21	21 – 37	37 – 100	≥ 10	$0 \leq 37$	$0 \leq 4$
Density (<i>kg.m⁻³</i>)	2750 ± 70	2938 ± 57	3300	3250	2852 ± 54	2550

TABLE 4.2 Initial layering and density of the Indian lithosphere (UC : upper crust ; LC : lower crust ; LM : lithospheric mantle), the asthenosphere (AST), the Tibetan crust (TC) and the sediments (SED).

4.2.6 Supplementary Material

4.2.6.1 Density structure

The density structure of the Indian lithosphere (Tab. 4.2) is set up using a $v_p - \rho$ relationship for the upper and lower crust rocks with compressional velocities derived from controlled source seismology experiments on the Indian shield (*Kaila et al. 1990, 1992; Sarkar et al. 2001; Behera et al. 2004*). The crustal thickness of the India plate is in agreement with receiver function derived Moho depth estimates in north-eastern India (*Mitra et al. 2005; Ramesh et al. 2005; Acton et al. 2010; Singh et al. 2010*). The density of the Tibetan crust is converted from v_p determined by *Hirn et al. (1984)*. Error bars indicate possible variances of the averaged depth estimates (*Christensen and Mooney 1995*). Standard densities are used for the continental lithospheric mantle and the asthenosphere.

4.2.6.2 Seismic Moho

All available seismic Moho depth constraints for the Eastern Himalaya and neighboring regions were collected and combined in a common reference frame (Fig. 4.4).

Constraints on Moho depth and crustal structure in the SE Tibetan Plateau near Bhutan were produced in the first phase of the INDEPTH project (*Hauck et al. 1998*) where a SSW-NNE-oriented controlled source seismology profile was recorded. In this profile a deep unambiguous reflection horizon that gently dips to the North marks the Moho discontinuity. Further constraints on Moho-topography in the study area come from passive source studies. *Kind et al. (1996)* performed receiver function analysis, Rayleigh wave dispersion and P-waveform modelling with the broadband data of the INDEPTH II campaign up to the Indus-Yarlung Suture. Further receiver function studies followed in the Sikkim Himalaya (*Acton et al. 2010; Singh et al. 2010*) and in NE India (*Ramesh*

et al. 2005; *Mitra et al.* 2005).

Seismic data in Bhutan is available from one experiment only (*Drukpa et al.* 2006; *Velasco et al.* 2007). The data from five stations operated in 2002-2003 is available through IRIS. Using teleseismic events, we computed receiver functions and performed inversions on the stacked waveforms to recover the approximate Moho depth beneath the stations. These have been included in the present Moho depth compilation.

To take into account the different data quality of the compiled Moho depths, two groups were established with respective depth error bars of ± 4 and ± 8 km. Finally, the seismic Moho depth constraints including error bars were projected on profiles oriented perpendicular to the strike of the orogen as defined in Fig. 1 of the main manuscript. Due to the sparseness of seismic Moho constraints in Bhutan and neighboring regions, the three Bhutanese profiles were merged. Future seismological studies covering larger areas with more stations will be able to map lateral variations and avoid projecting Moho topography on a single profile.

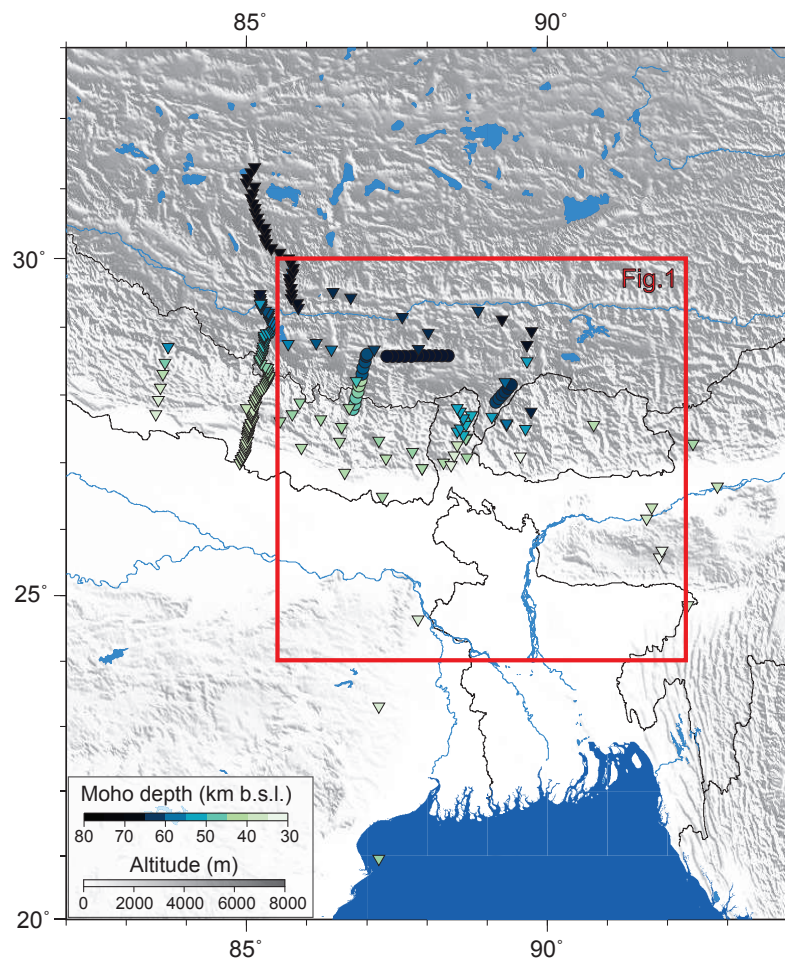


FIGURE 4.4 Available seismic Moho-constraints in the study area. Color according to Moho depth (referenced to sea-level). Triangles indicate constraints from receiver functions, circles mark controlled source seismology results. Red box outlines the area shown in Fig. 4.1 of the manuscript.

4.2.6.3 Rheology parametric study

To explore the parameter space of lithosphere rheology models, a total of 90 numerical simulations were carried out. The Young's moduli E of the different layers were varied from 50 to 125 GPa (*Palmström and Singh 2001*) and Poisson's ratios ν from 0.25 to 0.35 (*Zandt and Ammon 1995*). Both Young moduli and Poisson's ratio have a minor impact on the flexural shape compared to that of rheology and temperature. The Young moduli's impact is however key in the computation of the total EET as it determines the threshold effective viscosity that separates brittle behavior from ductile behavior for a specific relaxation time. As this dependence is linear and viscosity contrasts are evaluated on a logarithmic scale, using higher Young modulus has relatively minor effects on the calculated total EET.

For the case of Bhutan, a choice of $E = 60\text{-}70$ GPa also fits the observed gravity anomalies. Our final choice of $E = 50$ GPa is to demonstrate that a scenario with full mechanical coupling is also possible (the third model in which the load of the Tibetan Plateau is moved further South compared to all other models).

The thermal parameters were explored by altering the radiogenic heat production A and temperature at the lithosphere-asthenosphere boundary (LAB) T_m in the Royden temperature formulation (*Royden 1993*). Radiogenic heat production was altered simultaneously in the upper and lower crust from $A = 2.0 \mu W m^{-3}$ to $A = 4.0 \mu W m^{-3}$. The temperature at the LAB was varied between $T_m = 977$ °C $T_m = 1150$ °C. To test the influence of the ductile parameters different power-laws following *Kirby (1983)*, *Carter and Tsenn (1987)* and *Okaya et al. (1996)* were investigated : ten rheologies were tested for the lithospheric mantle, and four each for the upper and lower crust.

Deuxième partie

Variations latérales de la déformation
actuelle du prisme Himalayen entre le
Népal et le Bhoutan

CHAPITRE 5

DÉFORMATION ACTUELLE DU PRISME HIMALAYEN

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5.1 Introduction

Après avoir mis en évidence des variations latérales de la structure crustale entre le Népal et le Bhoutan, cette partie est consacrée à l'étude des variations latérales de la déformation récente du prisme Himalayen. Ce chapitre, qui fait l'état des lieux des connaissances actuelles dans ce domaine, permet de donner une base à la lecture des chapitres suivants. La première section présente les résultats préliminaires de l'application à l'Himalaya d'une méthode proposant une corrélation entre anomalies gravimétriques et comportement sismogénique en zone de subduction (*Song and Simons 2003*). Dans la suite de ce chapitre, je présente les données disponibles en termes de couplage intersismique et de sismicité instrumentale. Enfin, je ferai un rapide résumé des séismes majeurs ayant eu lieu au cours du dernier millénaire entre l'ouest Népal et le Bhoutan.

5.2 Comportement sismogénique et gravimétrie

5.2.1 Observations dans les zones de subduction océaniques

Dans ce paragraphe, je présente quelques usages des données gravimétriques pour l'étude du comportement sismogénique dans les zones de subduction. Je détaille plus précisément les travaux de *Song and Simons (2003)* qui, en utilisant les anomalies gravimétriques à l'air libre en mer, proposent une corrélation entre les variations latérales des anomalies gravimétriques et le comportement sismogénique des zones de subduction.

Pour ce faire, *Song and Simons (2003)* soustraient un profil gravimétrique régional moyen qui permet à la fois de faire ressortir la segmentation le long des zones de subduction étudiées et de les normaliser entre elles. Ils étudient ensuite les corrélations entre les variations latérales de gravimétrie et de topographie et l'occurrence des séismes dans les zones de subduction. Leurs observations suggèrent que les régions présentant des anomalies gravimétriques et topographiques parallèles à l'arc (AGPA et APTA) négatives par rapport à la moyenne sont sujettes à des séismes de plus forte magnitude et plus nombreux, contrairement aux régions avec AGPA et APTA positive qui sont relativement asismiques. La figure 5.1 présente les résultats obtenus dans la zone de subduction des Aléoutiennes/Alaska. À l'ouest de 160°W, on observe $AGPA < 0$ près de la fosse puis $AGPA > 0$ plus loin de la fosse, alors qu'à l'est c'est l'inverse (Fig. 5.1B). Dans les deux

cas, les majorité des séismes et les surfaces de rupture associées au séismes majeurs de 1957 ($M_w=8,6$), 1938 ($M_w=8.2$) et 1964 ($M_w=9.2$) ont lieu dans les régions où $AGPA < 0$.

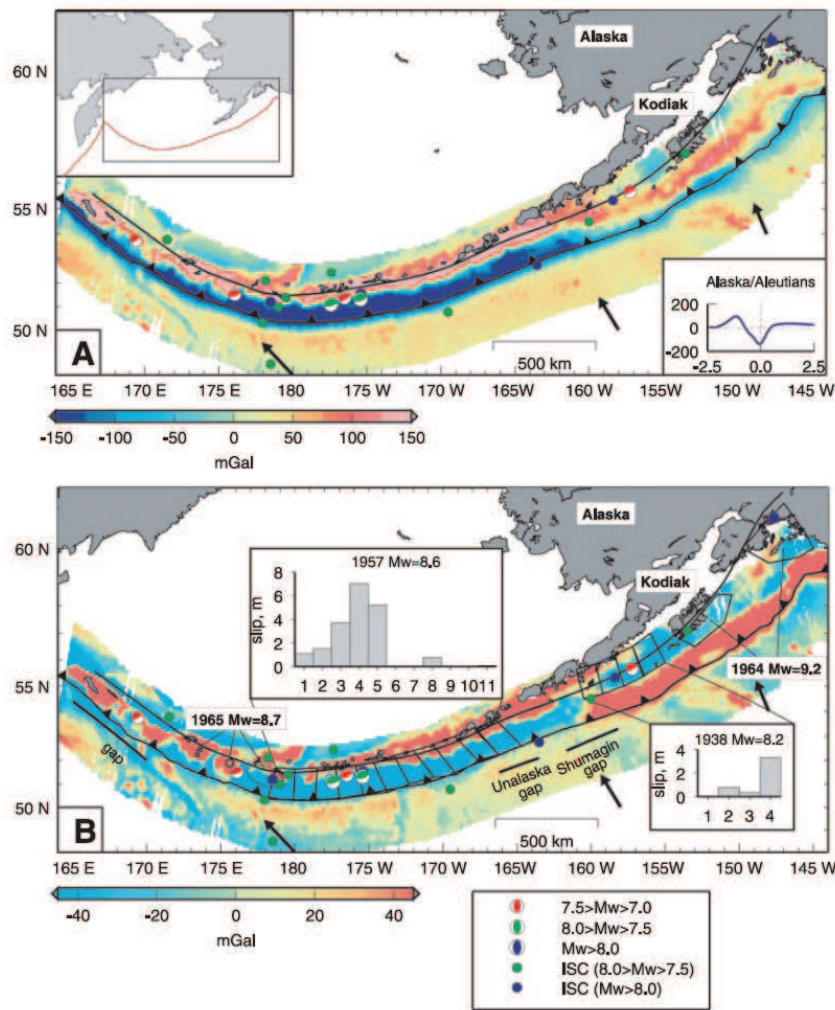


FIGURE 5.1 (A) Carte des anomalies de gravité (anomalies à l'air libre) de la zone de subduction des Aléoutiennes/Alaska. Le profil moyen de gravité perpendiculaire à la fosse calculé pour la région est représenté en bas à droite et est utilisé pour calculer les variations des anomalies de gravité le long de la fosse. (B) Carte des variations latérales des anomalies gravimétriques le long de la fosse calculées en soustrayant le profil moyen à la carte des anomalies de gravité en (A). Les séismes de $M_w > 7$ sont représentés ainsi que les surfaces de rupture des séismes majeurs (1938, 1957, 1964 et 1965)

Suite au calcul des APTA, ils observent que dans la majorité des zones de subductions, les segments où APGA & APTA sont fortement négatives sont les plus actives en terme de sismicité. Afin d'expliquer la corrélation entre AGPA/APTA et le comportement sismogénique, *Song and Simons (2003)* proposent un modèle où les variations de contrainte cisailante sur l'interface (qui dépendent de la contrainte normale et du coefficient de friction) impliquent des variations de topographie sur l'avant-arc sur le long terme (Fig. 5.2A).

Si un tel mécanisme a déjà été mis en évidence par des modèles visco-élastiques (Fig. 5.2B ; *Cattin et al. (1997)*), justifiant la possible corrélation entre comportement sismogénique et topographie, on peut se demander ce qu'apportent les anomalies gravimétriques à l'air libre dans cette étude qui sont intrinsèquement dépendantes de la bathymétrie.

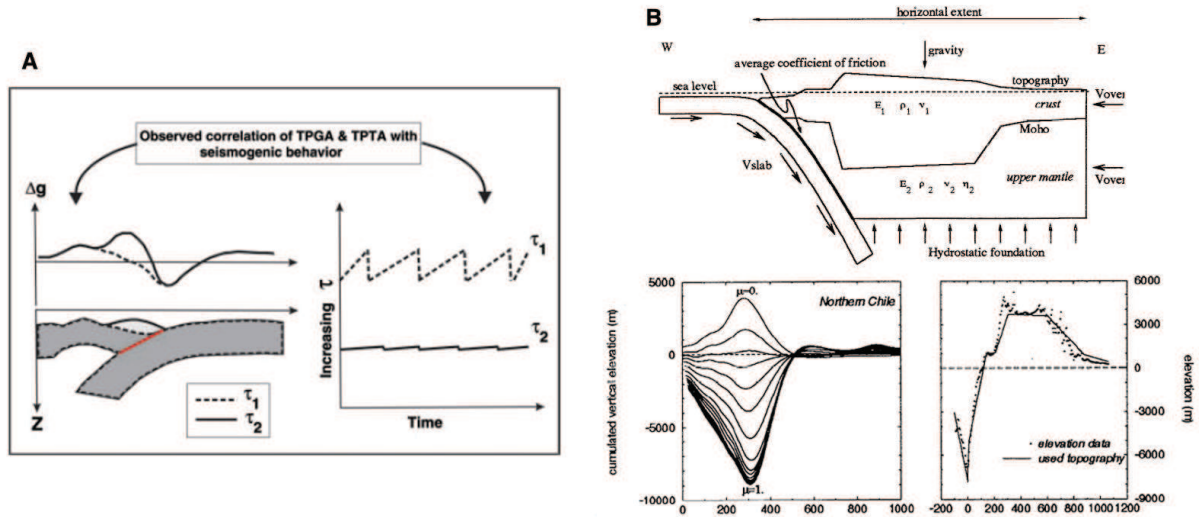


FIGURE 5.2 Principe mécanique sur lequel repose la méthode de *Song and Simons (2003)*. (A) Schéma conceptuel illustrant les variations de topographie et de gravité impliquées par une variation de la contrainte cisailante sur l'interface de subduction (ligne rouge), d'après *Song and Simons (2003)*. Les lignes noires en pointillées correspondent au modèle où le cisaillement est fort et les lignes en trait plein, un modèle où le cisaillement est relativement plus faible. (B) Résultats de modélisation visco-élastique permettant de mettre en évidence une corrélation entre la friction sur l'interplaque et la topographie (*Cattin et al. 1997*).

L'utilisation que font *Wells et al. (2003)* des anomalies gravimétriques dans ces mêmes zones de subduction me semble plus pertinente. En effet, ils utilisent à la fois le signal gravimétrique, la bathymétrie et les contraintes géophysiques pour montrer une corrélation entre les structures de l'avant-arc (notamment les bassins) et les glissements cosismiques. Ils proposent que ce lien existe aussi sur le long-terme grâce au phénomène de subsidence associé au comportement érosif de l'interplaque. Les récentes observations d'une corrélation entre le couplage intersismique et les structures de la plaque supérieure Andine par *Bejar-Pizarro et al. (2013)* vont dans ce sens.

Pour finir sur les liens possibles entre anomalies gravimétriques dans les zones de subduction et comportement sismogénique de l'interplaque, je citerai un dernier exemple qui illustre l'intérêt des anomalies de Bouguer. *Grevemeyer and Tiwari (2006)* corrélient les variations des anomalies de Bouguer entre la zone de Sumatra et de Java (relativement asismique par rapport à Sumatra) à une différence de comportement sismogénique liée à

l'extension de la zone couplée. Ils proposent que l'anomalie positive au sud de l'île de Java soit liée à la présence d'un coin mantellique dans la plaque supérieure qui limite l'extension de la zone bloquée en profondeur et donc l'occurrence de séismes majeurs. Dans ce cas, les anomalies gravimétriques (Bouguer) sont utilisées pour mettre en évidence une variation latérale de structure profonde pouvant conduire à une segmentation sismogénique.

5.2.2 Application à l'arc Himalayen

À partir des anomalies gravimétriques obtenues au cours de cette thèse, nous avons effectué le calcul de APTA et APGA le long de l'arc Himalayen entre l'ouest Népal et le Bhoutan. Les données topographiques utilisées proviennent du MNT SRTM3 dégradé à 1' d'arc afin de lisser le signal haute fréquence lié à l'érosion. Pour définir le profil moyen perpendiculaire à l'arc, une régression polynomiale à partir de tous les points de données et de leur distance au MFT a été réalisée. Le profil moyen est ensuite soustrait aux données. J'interpole le résultat pour obtenir les cartes de APTA et APGA représentées sur la figure 5.3.

La carte des variations d'anomalies topographiques montre une nette transition entre le Népal et le Bhoutan. Au nord du MFT, on observe une anomalie négative au Népal sur une centaine de km alors qu'elle est positive au Bhoutan. Au delà de cette zone, on observe une anomalie positive au centre et à l'est Népal qui correspond aux sommets du Haut Himalaya. L'ouest Népal semble différent du reste du Népal à ce niveau de l'arc mais dans une moindre mesure que le Bhoutan. Les APGA montrent aussi une variation latérale importante entre le Népal et le Bhoutan qui s'opère au niveau du Sikkim. Du sud au nord, les APGA évoluent du négatif au positif au Népal alors qu'elles évoluent du positif vers le négatif au Bhoutan. La transition entre $APGA > 0$ et $APGA < 0$ au Bhoutan est située au centre du pays mais pourrait être légèrement décalée en latitude une fois le réseau gravimétrique Bhoutanais calé en absolu.

La variation latérale de topographie observée entre le Népal et le Bhoutan (Fig. 5.3A) est en accord avec les observations de *Duncan et al. (2003)* faites à partir de profils perpendiculaires à la chaîne (Fig. 1.3CD). Les variations latérales de topographie observées au Népal semblent être corrélées avec la géométrie du MHT qui présente une rampe plus prononcée à l'est (*Mugnier et al. 2013*). En ce qui concerne les variations d'anomalies gravimétriques de Bouguer, aucune corrélation n'est observée avec APTA. En effet, le signal

dû à la topographie dans les mesures gravimétriques est corrigé pour obtenir les anomalies de Bouguer. Au Népal, *Ader et al.* (2012) montrent que le MHT est complètement bloqué pendant la phase intersismique sur une centaine de km à partir du MFT. Les séismes majeurs en Himalaya ont lieu sur cette portion du MHT. Si les APTA montrent une anomalie négative dans cette zone, les APGA sont positives à cet endroit ce qui montre l'absence de corrélation directe entre APGA, APTA et le comportement sismogénique. La variation latérale des APGA observée est liée à la structure crustale et aux variations du comportement flexurale de la plaque Indienne analysée dans les chapitres 3 et 4. Il est difficile d'établir un lien direct entre cette variation structurale et le comportement sismogénique du MHT. Cependant la température joue un rôle prépondérant sur les mécanismes de déformation en contexte continental et plus particulièrement la zone de transition entre la zone couplée et le glissement stable (*Dominguez et al.* 2003; *Hsu et al.* 2010). En effet, *Cattin and Avouac* (2000) montrent qu'au Népal la transition entre la zone couplée et la zone qui flue en intersismique est située au niveau de l'isotherme 350°C. Nous avons montré, dans le chapitre 4, qu'un des modèles permettant de diminuer la rigidité flexurale de la plaque Indienne au Bhoutan nécessitait un champ de température plus élevé, ce qui pourrait augmenter le champ de température au niveau du MHT et réduire la largeur de la zone bloquée par rapport au Népal.

5.3 Couplage intersismique en Himalaya

Par rapport aux zones de subduction océanique, l'arc Himalayen offre une meilleure couverture GPS et donc une mesure plus complète des déformations. Si nous avons montré dans le paragraphe précédent l'existence de variations latérales de topographie et de gravité qui pourrait suggérer des variations de couplage long terme, seule l'analyse des données GPS au travers de l'arc Himalayen permet une quantification du couplage intersismique. À partir d'un réseau de 30 stations GPS au Népal et au sud Tibet, *Ader et al.* (2012) montrent que le MHT est bloqué à partir de la surface sur une distance horizontale de 100 km (Fig. 5.4). Au-delà, la déformation est accommodée de manière ductile le long de la prolongation en profondeur du MHT. La zone de transition entre la partie bloquée du MHT et la partie qui flue est inférieure à 30 km et se situe à des profondeurs entre 15 et 20 km où les modélisations thermo-cinématiques prédisent des températures $>350^{\circ}\text{C}$

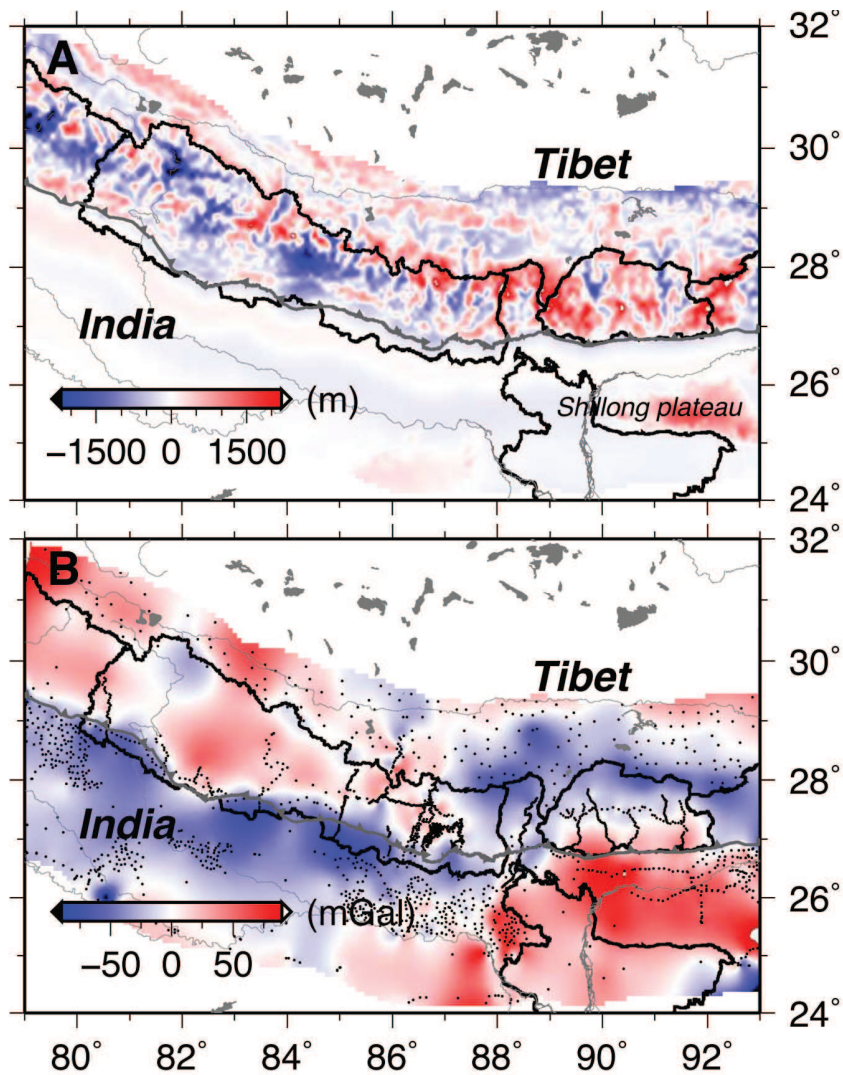


FIGURE 5.3 Variations latérales topographiques et gravimétriques le long de l'arc Himalayen. (A) Carte des variations latérales des anomalies topographiques. Elles sont calculées à partir du MNT SRTM3 dégradée à 1' d'arc (afin de limiter le signal haute fréquence lié à l'érosion en domaine continental). La moyenne retranchée est calculée à partir de tous les points du MNT située à $\pm 2.5^\circ$ du MFT. (B) Carte interpolée des variations latérales des anomalies gravimétriques calculées à partir des anomalies de Bouguer acquises au cours de cette thèse et des pré-existantes (points noirs).

(Fig. 5.4B). Le raccourcissement estimé entre l'Inde et le sud Tibet est de $20,5 \pm 1 \text{ mm/an}$ pour l'ouest Népal et $17,8 \pm 0,5 \text{ mm/an}$ à l'est Népal. À l'est du Népal, les estimations

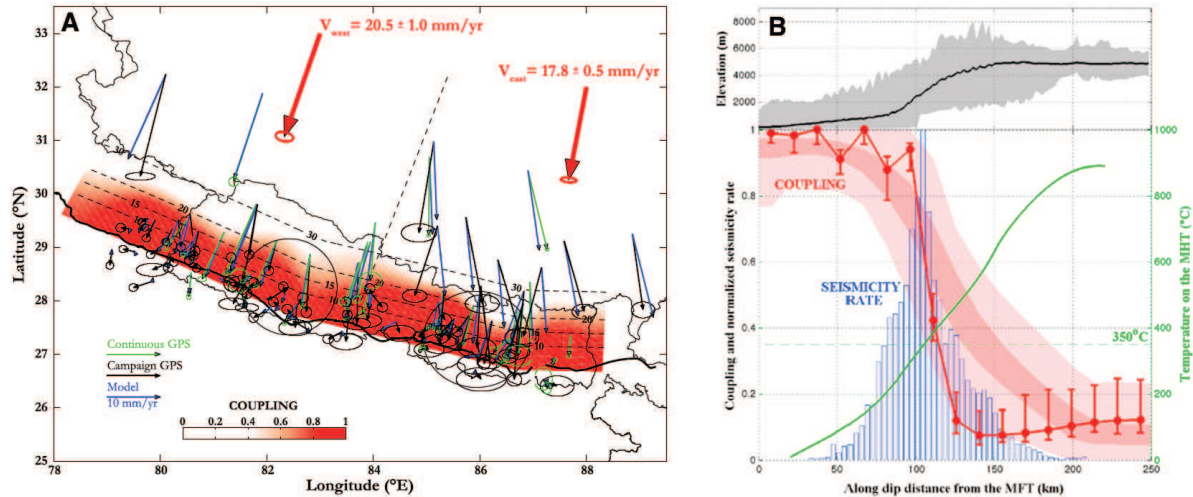


FIGURE 5.4 Couplage intersismique au Népal, d'après *Ader et al.* (2012). (A) Le couplage intersismique est représenté en ombrage de rouge sur la faille. Comparaison des vitesses horizontales observées (GPS continu en vert et campagnes GPS en noir) et prédites par le modèle de couplage (flèches bleues). Les flèches rouges indiquent le chargement séculaire. (B) Comparaison, en coupe, du couplage intersismique (en rouge), du taux de sismicité (en bleu) et de la température calculée sur l'interface (en vert). La ligne rouge correspond à la valeur du couplage calculée au niveau du profil de nivellement disponible au Népal central. La topographie moyenne est représentée au dessus.

de raccourcissement sont du même ordre de grandeur : $15 - 20 \text{ mm/yr}$ et l'implication du plateau du Shillong ne semble pas majeure ($\sim 15\%$ tout au plus) (*Jade et al.* 2007; *Mukul et al.* 2010). Les récentes remesures d'un réseau GPS installé en 2003 à l'ouest Bhoutan semblent montrer des caractéristiques similaires au Népal. En effet la modélisation de la déformation intersismique à l'ouest Bhoutan présentée à la figure 5.5B montrent une dislocation situé à $\sim 100 \text{ km}$ du front Himalayan (*Vernant et al.* 2013). Lors de ces missions, des profils N/S au centre et à l'est Bhoutan ont été installés et devraient nous renseigner sur le comportement du MHT dans cette zone (Fig. 5.5A). En attendant de meilleures contraintes au centre et à l'est Bhoutan, l'apparente absence de variations latérales de couplage intersismique jusqu'à l'ouest Bhoutan est un résultat intéressant en cela qu'il diffère des zones de subduction où il existe des variations latérales de couplages (eg *Chlieh et al.* 2008).

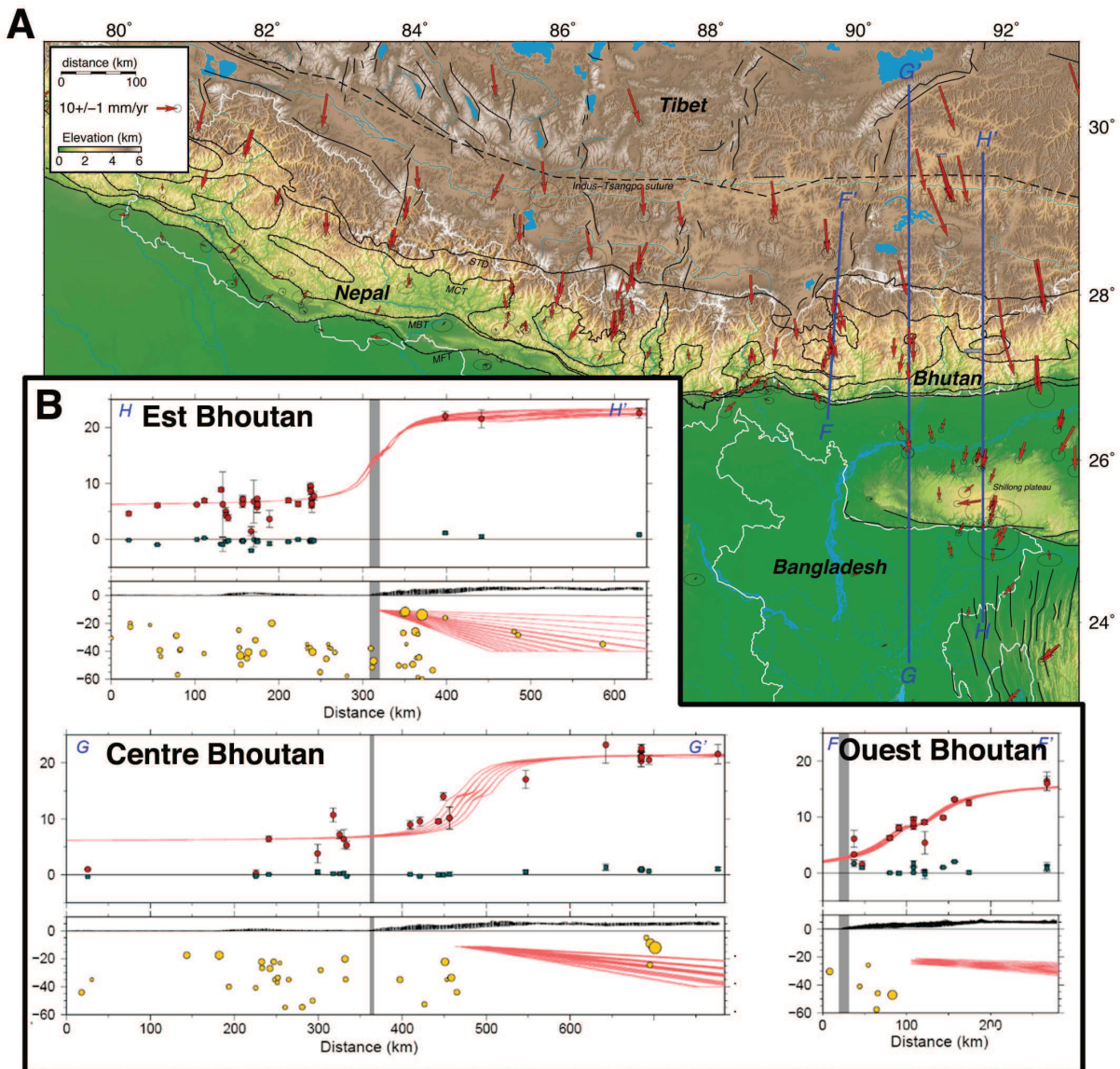


FIGURE 5.5 (A) Déformation intersismique mesurée par GPS incluant les récentes mesures effectuées au Bhoutan, d'après *Vernant et al. (2013)*. Les flèches rouges indiquent les vitesses GPS dans le référentiel Inde fixe. (B) Modélisation de la déformation intersismique avec une approche permettant d'inverser les déplacements de surface selon 4 paramètres : la localisation de la dislocation (distance et profondeur), le pendage du MHT et le taux de raccourcissement. En haut, vitesses GPS observées (en rouge la composante normale, en vert la composante décrochante) et les 20 meilleurs modèles issus de l'inversion (lignes rouges). En bas, géométries du MHT pour les 20 meilleurs modèles issus de l'inversion. Les ronds jaunes représentent la sismicité.

5.4 Sismicité instrumentale

L'installation d'un réseau sismologique permanent au Népal en 1994 a révélé une microsismicité intense (Fig. 5.6A). Cette microsismicité se concentre ~ 100 km au nord du MFT, le long de la haute chaîne Himalayenne. Elle est en revanche beaucoup plus faible au niveau du moyen pays et des Siwaliks. La concentration de ces événements surimposée aux coupes géologiques à l'ouest et au centre Népal montre que la microsismicité se concentre au voisinage de la rampe mi-crustale du MHT (Fig. 5.6B). Cette zone correspond à la transition entre la portion bloquée du MHT et la portion profonde qui flue. Ce mécanisme conduit à une accumulation de contraintes dans cette zone à l'origine de la microsismicité. Cependant le moment cumulé associé à cette microsismicité ne correspond qu'à un faible pourcentage de la déformation intersismique mesurée ($< 2\%$) (*Avouac et al. 2001*). La vitesse de raccourcissement déduite à partir de l'étude des terrasses Holocènes au niveau du MFT (*Lavé and Avouac 2000*) est semblable à celle estimée sur le MHT pendant l'intersismique. Cela montre non seulement que le prisme Himalayen subit peu de déformation interne mais aussi que la déformation accumulée durant l'intersismique est élastique et transférée au niveau du front Himalayen. La plupart de ce transfert se fait lors des séismes majeurs ($M_w > 8$) qui nucléent près de la zone de transition et se propagent le long du MHT jusqu'à son émergence en surface (*Avouac et al. 2001*).

L'analyse de la sismicité enregistrée dans les catalogues globaux montre une différence notable à $\sim 89^\circ\text{E}$ (Fig. 5.7). En effet, la sismicité est concentrée dans la chaîne Himalayenne à l'ouest de 89°E alors qu'il existe aussi de la sismicité au sud du MFT, dans la plaque Indienne, à l'est de 89°E . Cette observation suggère que la plaque indienne subit une déformation plus intense qu'au Népal avant son passage en subduction sous le Bhoutan.

Les mécanismes au foyer compilés par *Kayal (2001)* montrent que le régime chevauchant associé au MHT est présent sur toute la longueur de l'arc (Fig. 5.8A). Les pendages de ces mécanismes au foyer sont faibles ($\sim 10\text{-}20^\circ$ vers le nord) et ne présentent pas de variations latérales marquées. Cependant, l'analyse de la sismicité instrumentale ($4.5 < M_b < 6.5$) au Bhoutan entre 1937 et 2003 montre l'existence d'un régime décrochant dans cette partie de l'arc (Fig. 5.8C ; *Drukpa et al. (2006)*). Ce régime décrochant n'existe pas à l'ouest du Sikkim. La coupe présentée sur la figure 5.8B est située au niveau de cette zone de transition et montre la concentration des mécanismes au foyer dans la plaque Indienne (*Baur 2007*). Ceci indique que la plaque Indienne continue à se déformer lors de

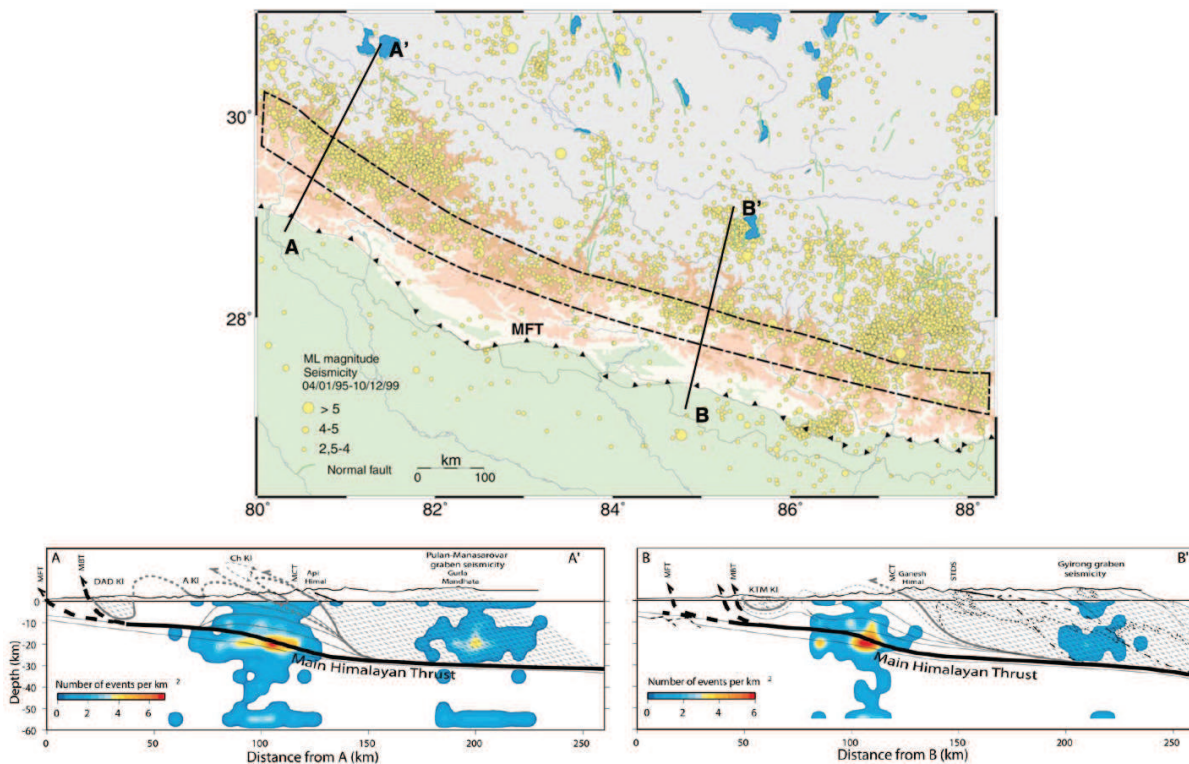


FIGURE 5.6 En haut, microsismicité du Népal enregistrée et localisée par le réseau du National Seismological Center entre 1995 et 1999 (*Avouac et al. 2001*). Le tiret noir délimite la zone où se concentre la microsismicité et qui correspond à la zone de transition fragile-ductile au niveau du MHT. En bas, coupes géologiques et densité de la microsismicité d'après *Bollinger et al. (2004)*. La microsismicité est concentrée au niveau de la rampe du MHT.

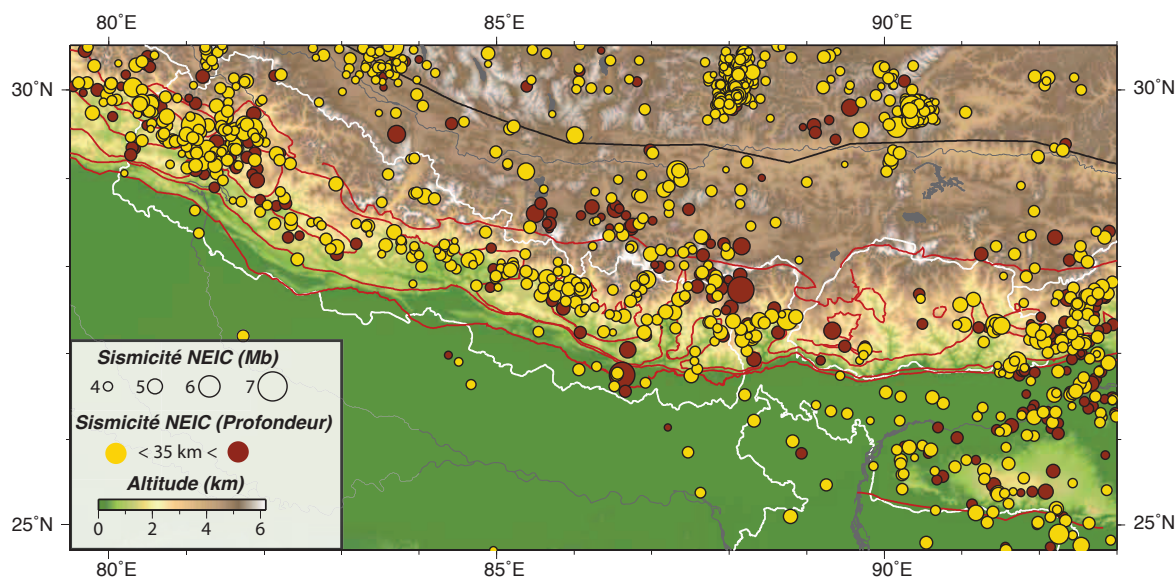


FIGURE 5.7 Sismicité instrumentale du catalogue NEIC (1973-2013). Les cercles jaunes correspondent à des hypocentres situés au dessus de 35 km de profondeur et les marrons en dessous. Les lignes rouges représentent les principales failles du système Himalayen.

son passage en subduction sous le Bhoutan.

5.5 Séismes majeurs du dernier millénaire

Comme nous venons de le voir, le moment sismique libéré lors des séismes majeurs ($M_w > 8$) en Himalaya permet de relâcher l'essentiel des contraintes élastiques accumulées durant l'intersismique. À l'est de 80°E, trois séismes majeurs ont eu lieu au cours du XX^e siècle : Bihar-Népal en 1934 (M_w 8,1), Assam 1950 (M_w 8,4) et au sud du front Himalayen au niveau du plateau du Shillong en 1897 (M_w 8.4) (Fig. 5.9A). Les récentes mises en évidence de ruptures de surface à proximité du front Himalayen associées aux séismes instrumentaux de 1934 et 1950 renforcent l'hypothèse d'une déformation long-terme principalement accumulée au niveau du MFT (*Jayangondaperumal et al. 2011; Sapkota et al. 2013*). Cela n'écarte pas pour autant la possible occurrence de séismes sur des chevauchements hors-séquences (Cachemire 2005 - $M_w \sim 7.6$) ou de séismes majeurs n'atteignant pas la surface. Les aires de ruptures des séismes majeurs du siècle dernier laissent apparaître des zones non rompues (ouest Nepal, Sikkim-Bhoutan).

La compréhension du comportement sismogénique du MHT le long de l'arc nécessite

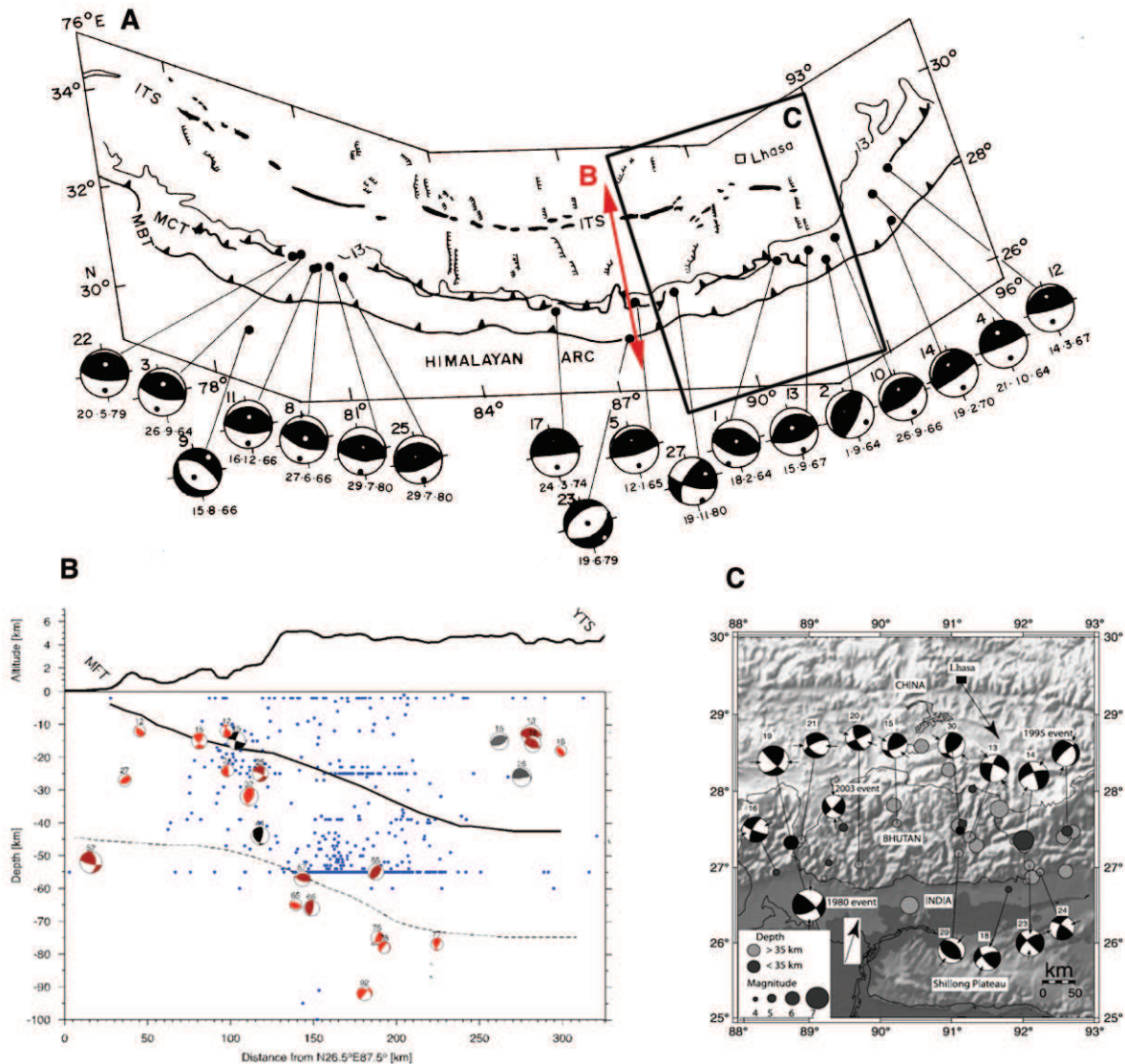


FIGURE 5.8 Mécanismes au foyer dans la région de l'Himalaya. (A) Carte séismotectonique de l'arc Himalayen montrant le régime chevauchant présent sur toute la longueur de l'arc (*Kayal 2001*). Les mécanismes au foyer montrent systématiquement un faible pendage ($\sim 10\text{-}20^\circ$ vers le nord) mais pas de variations latérales évidentes. (B) Mécanismes au foyer en coupe au niveau du Sikkim, d'après *Baur (2007)*. La plupart de ces mécanismes sont situées dans la plaque Indienne. Cette zone correspond à une transition qui marque l'apparition d'un régime avec une composante décrochante à l'est. (C) Mécanismes au foyer au Bhoutan qui mettent en évidence l'existence d'une composante décrochante au régime de déformation de cette région, d'après *Drukpa et al. (2006)*.

donc d'étendre les catalogues de sismicité. Les études paléosismologiques constituent un bon outil pour documenter les séismes ayant produit des déformations en surface sur le dernier millénaire. Elles sont basées sur l'observation et la datation des déformations associées aux ruptures de surface dans l'enregistrement sédimentaire (e.g. *Sieh 1978*). Depuis une quinzaine d'années, et la première mise en évidence de rupture de surface en Himalaya (Fig. 5.10 (*Nakata et al. 1998*)), de nombreuses tranchées ont été ouvertes, permettant de mieux contraindre l'âge et la magnitude des séismes majeurs. La localisation des différentes études paléosismologiques entre l'ouest Népal et l'Arunachal Pradesh est présentée dans la figure 5.9. Les méthodes de datation utilisées en tranchée (principalement ^{14}C et OSL) ne permettent pas d'obtenir des barres d'erreurs inférieures à la centaine d'années (Fig. 5.9D). La corrélation de ces contraintes d'âge avec les écrits historiques permet une meilleure datation des événements. Par exemple la rupture de surface observée par *Yule et al. (2006)* à l'ouest Népal (Fig. 5.11) est datée entre 1610 et 1410 a été attribué au séisme de 1505 décrit par *Ambraseys and Douglas (2004)*. En plus de la rupture associée au séisme de 1934 au site de Sir Khola à l'Est Népal, *Sapkota et al. (2013)* ont mis en évidence un autre séisme dont les contraintes d'âge sont compatibles avec le séisme historique de 1255 ayant tué 1/3 de la population de Katmandou dont Abhaya Malla, le roi Népalais d'alors. Ce séisme est aussi compatible avec les contraintes d'âge obtenues par *Mugnier et al. (2005)* à 82,20°E de longitude (ouest Népal). Sur ce site, *Mugnier et al. (2005)* montrent que les ruptures atteignent la surface à différents endroits : au niveau de la plaine alluviale, au niveau du front topographique (contact entre les sédiments Quaternaires et les Siwaliks) et aussi au sein même des Siwaliks. Au centre Népal en rive droite de la rivière Marha, *Lavé et al. (2005)* ont mis en évidence un escarpement tectonique de 7.5 m de haut et estimé un déplacement cosismique de ~ 17 m qui suggère l'occurrence d'un très fort séisme vers 1100. Par la suite, des études paléosismologiques de part et d'autre du Bhoutan ont montré des événements avec des glissements cosismiques comparables à celui calculé par *Lavé et al. (2005)* et dont les contraintes d'âge sont compatibles avec l'événement de ~ 1100 . Ainsi, *Kumar et al. (2010)* propose que la rupture associée au séisme de ~ 1100 s'étende du centre Népal jusqu'en Arunachal Pradesh couvrant un segment de 800 km le long de l'arc Himalayen incluant le Bhoutan. Une telle surface de rupture correspondrait à des magnitudes approchant 9, encore jamais observées en domaine continental. Cependant, les incertitudes sur les datations des événements passés et sur leur extension spatiale permettent d'envisager des scénarii alternatifs impliquant plusieurs

séismes majeurs au lieu d'un seul et unique événement. En combinant l'étude des sismites dans le bassin de Katmandou, les écrits historiques et les résultats de paléosismologie, *Mugnier et al.* (2013) proposent un modèle différent, impliquant deux événements de part et d'autre du Bhoutan plutôt qu'un seul méga-seisme.

5.6 Conclusion

La segmentation structurale au niveau du comportement de la plaque Indienne entre le Népal et le Bhoutan est aussi marquée par une sismicité intraplaque Indienne et l'apparition d'une composante de déformation décrochante à l'est de 89°E. Les mesures GPS à l'ouest Bhoutan semblent montrer un couplage intersismique similaire à celui du Népal mais pour l'instant, l'absence de données au centre et à l'est du Bhoutan ne permettent pas de confirmer cette tendance. Les nombreuses études paléosismologiques le long de l'arc Himalayan menées depuis 15 ans, ont permis d'étendre le catalogue des événements majeurs sur ~ 1000 ans avec la possibilité d'un méga-événement en ~ 1100 . Dans le chapitre 6 nous étudierons les variations de contraintes accumulées sur le MHT durant les 900 dernières années en prenant en compte le chargement séculaire et les séismes majeurs. Dans le chapitre 7, nous présenterons les résultats de la première étude morphotectonique au Bhoutan permettant d'apporter de nouvelles contraintes en termes de paléosismologie et de raccourcissement Holocène au front de la chaîne.

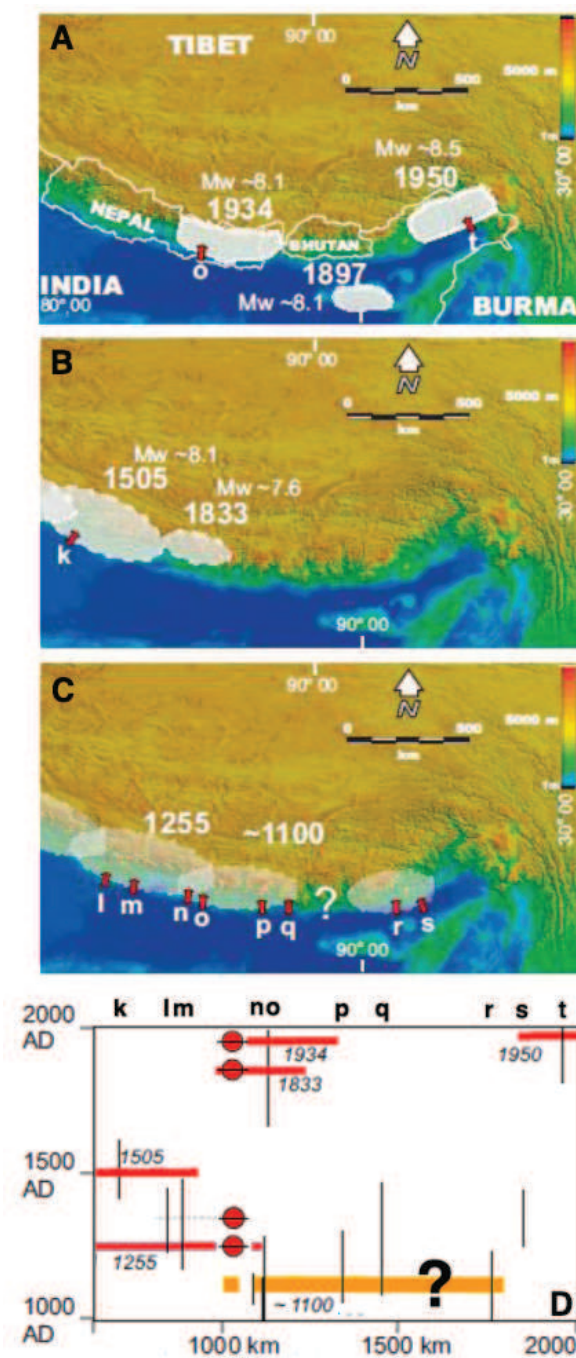


FIGURE 5.9 Les séismes majeurs du dernier millénaire en Himalaya, adapté de *Mugnier et al. (2013)*. Les rectangles rouges représentent les évidences de ruptures de surface : (k) (*Yule et al. 2006*); (l) (*Mugnier et al. 2005*); (m) (*Upreti et al. 2008*); (n) (*Lavé et al. 2005*); (o) (*Sapkota et al. 2013*); (p) (*Nakata et al. 1998*); (q,r,s) (*Kumar et al. 2010*); (t) (*Jayangondaperumal et al. 2011*). (A) Séismes instrumentaux du dernier siècle. (B) Séismes majeurs entre 1890 et 1500. (C) Séismes majeurs avant 1500. (D) Diagramme présentant les contraintes d'âges obtenues par la paléosismologie (barre verticales noires), les séismes de fortes intensités décrits à Katmandou (rond rouges; (*Pant 2002*)) et l'extension supposée des ruptures des séismes majeurs (barres horizontales).

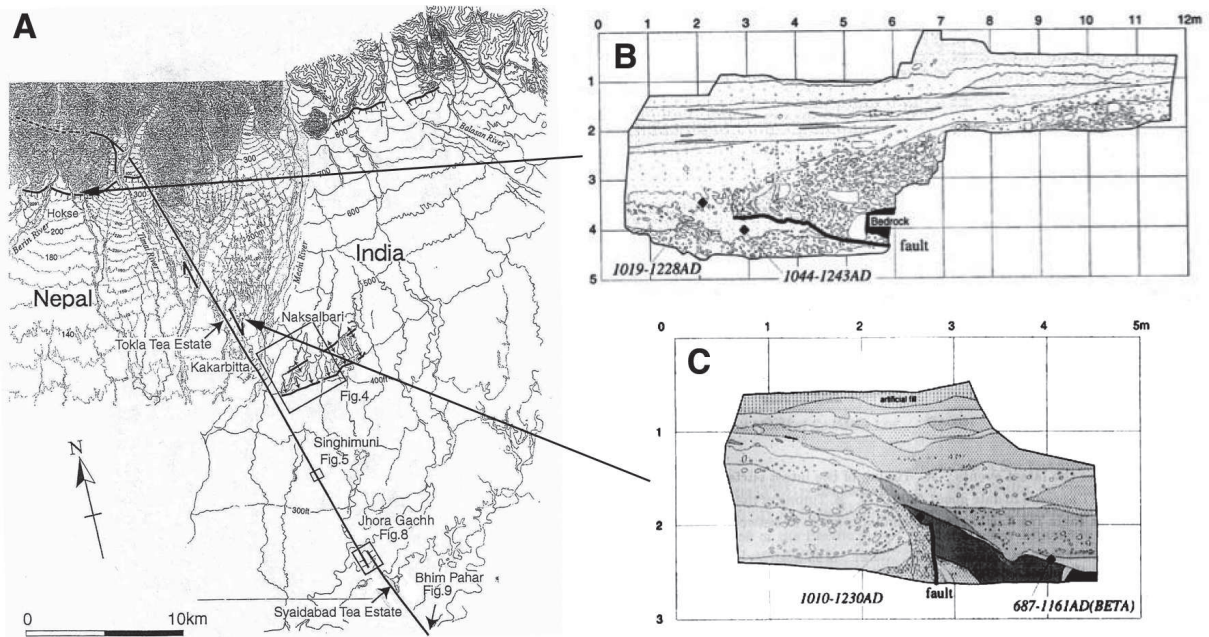


FIGURE 5.10 Premières évidences de ruptures de surface en Himalaya (*Nakata et al. 1998*). (A) Carte de localisation des deux tranchées. (B) Log de la tranchée du site Hoske. (C) Log de la tranchée du site de Tea Estate.

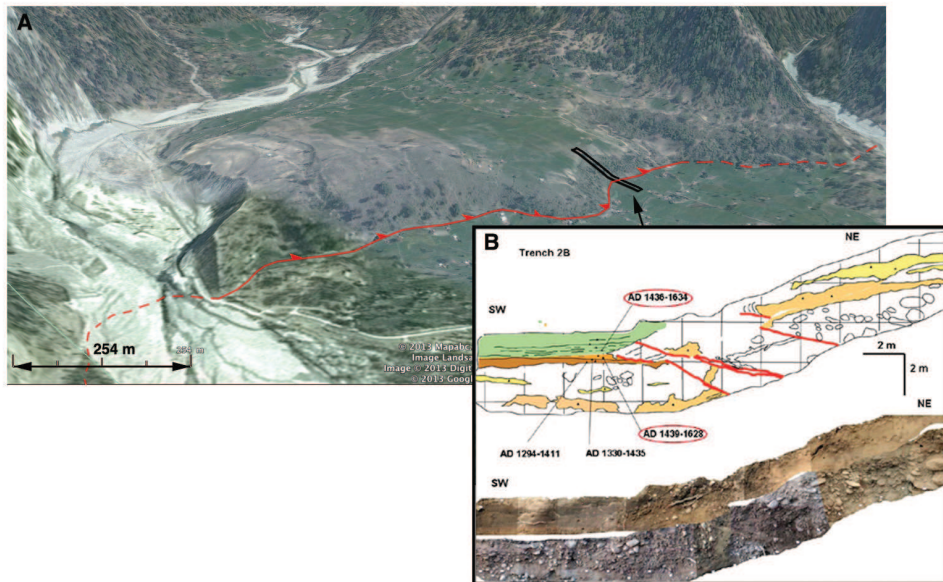


FIGURE 5.11 (A) Vue GoogleEarth de l'escarpement du MFT à l'ouest Nepal associé aux travaux de *Yule et al. (2006)*. (B) photo-mosaïque de la tranchée et schéma interprétatif ayant permis d'associer cette rupture de surface au séisme historique de 1505, d'après *Sapkota (2011)*.

CHAPITRE 6

VARIATIONS DE CONTRAINTE SUR LE

CHEVAUCHEMENT HIMALAYEN AU COURS DU

DERNIER MILLÉNAIRE

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6.1 Présentation

Comme nous venons de le voir, les études paléosismologiques menées depuis 15 ans en Himalaya ont permis d'étendre le catalogue des événements majeures en Himalaya sur le dernier millénaire. En intégrant à la fois les contraintes disponibles sur ces séismes majeurs et le chargement séculaire, nous étudions les variations spatiales et temporelles de l'état de contrainte du MHT le long de l'arc Himalayen.

Dans ce chapitre, nous utilisons le code développé par *Toda et al. (2011)*¹ afin de calculer l'évolution des contraintes de Coulomb sur le MHT entre 79°E et 93°E. Plusieurs modèles sont testés pour prendre en compte les incertitudes sur le chargement séculaire à l'Est Bhoutan et sur l'occurrence d'un séisme de magnitude > 9 qui aurait rompu une zone de plus de 900 km entre le Centre Népal et l'Est Bhoutan.

Contrairement aux résultats obtenus par *Gahalaut et al. (2011)*, les modèles présentés dans ce chapitre montrent que les contraintes induites par le séismes du Shillong en 1897 ne jouent pas un rôle majeur sur l'état de contrainte du MHT au Bhoutan. De plus, les modèles ne montrent pas d'évidences nettes de déclenchement de séismes majeures liés aux variations de contraintes de Coulomb associées au séisme antérieur. Cependant, cette hypothèse ne peut pas complètement être écartée, notamment à cause des incertitudes sur les surfaces de ruptures et les glissements moyens associés ainsi que sur les contraintes temporelles des séismes majeurs. Cette étude montre aussi que même dans le cas où un méga-séisme s'est produit en ~ 1100 en Himalaya, le chargement séculaire a quasiment compensé la chute de contrainte associée. Enfin, les résultats obtenus suggèrent que la distribution spatiale des séismes majeurs passés suffit à expliquer la sismicité actuelle. Ce résultat suggère une différence de comportement avec les zones de subduction où la distribution de la sismicité est aussi liée aux variations latérales de taux de convergence ou de couplage sismique. Ce chapitre est présenté sous la forme d'un article soumis à *Geophysical Journal International*.

1. available at <http://pubs.usgs.gov/of/2011/1060/>

6.2 Stress change over the last 900 years along the Main Himalayan Thrust fault from western Nepal to eastern Bhutan

Paper submitted to Geophysical Journal International

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Abstract

Although the mountain building processes have been widely studied in Himalayas, the seismic behaviour of this ~ 2500 km long area remains poorly understood. For instance, the possibility of a $M > 9$ earthquake occurring on the Main Himalayan Thrust (MHT) is still a matter of debate. Here, taking advantage of available information on the timing and location of major earthquakes, we assess the stress change along the MHT over the last nine centuries. In our approach we calculate the Coulomb stress change due to both the secular loading and the co-seismic rupture of historic events in the region of Nepal and Bhutan Himalayas. Due to uncertainties in the co-seismic slip distribution for some earthquakes as well as in the inter-seismic coupling of the MHT in Bhutan many scenarios are tested. For distances to fault rupture edges greater than 100 km our calculations show that the effect of a seismic event is lower than 0.1 bar. Hence our results suggest no obvious evidence for major stress-triggered earthquakes over a few decades or less, even though a link cannot be completely ruled out. Comparison with present-day deformation field reveals that the pattern of seismic activity along the Himalayan arc from Nepal to Bhutan is most likely related to past events rather than major longitudinal variations in the convergence rate as commonly proposed or in the inter-seismic coupling of the MHT. More importantly, our calculation indicates that inter-seismic loading has now nearly overcome the Coulomb stress decrease caused by the ~ 1100 medieval earthquake(s). This suggests that one or more major earthquakes along the Himalayan arc are highly plausible in the near future.

Keywords : Inter-seismic coupling, Stress triggering, Historic earthquakes, Present-day seismicity, Himalaya.

6.2.1 Introduction

The frequent magnitude 7-8 earthquakes occurring along the Himalayan arc are one of the most dramatic manifestations of the shortening between India and Eurasia (Fig. 6.1). Over geological time scales, it is now well-established that this frontal deformation has been accommodated along three crustal-scale north verging thrust faults : the Main Central Thrust (MCT), the Main Boundary Thrust (MBT) and the Main Frontal Thrust (MFT) that strike the entire length of the Himalayan arc and merge into a common décollement, the Main Himalayan Thrust (MHT) (*Gansser 1964*). Furthermore both river terrace observations and GPS measurements in Nepal indicate that the present-day shortening rate of 15-20 mm/yr is taken up by thrusting on the MFT (e.g. *Lavé and Avouac 2000*; *Ader et al. 2012*). The inter-seismic loading is thus suggested to be released through major earthquakes initiated beneath the Higher Himalaya and propagating southward up to the MFT (eg *Seeber and Armbruster 1981*; *Cattin and Avouac 2000*). Recent evidences of surface ruptures associated with major earthquakes argue in favour of this hypothesis (e.g. *Lavé et al. 2005*; *Kumar et al. 2010*; *Sapkota et al. 2013*). For consistency with previous terminology, we refer to this fault as the seismogenic MHT hereinafter.

Over the last four decades many studies including historic and instrumental information as well as paleoseismological investigations have provided estimates of the timing, magnitude and rupture parameters for these major earthquakes (e.g. *Chen and Molnar 1977*; *Molnar and Pandey 1989*; *Chander 1989*; *Kumar et al. 2001*; *Bilham 2004*; *Rajendran and Rajendran 2005*; *Lavé et al. 2005*; *Szeliga et al. 2010*; *Kumar et al. 2010*). However, except for the area or particular earthquakes like the 2005 Kashmir or 1897 Shillong events (e.g. *Cattin and Avouac 2000*; *Gahalaut 2008*; *Gahalaut et al. 2011*; *Arora et al. 2012*), relationship between major events, inter-seismic loading, stress interaction between them as well as their effect on the current seismicity rate have been poorly studied (*Bollinger et al. 2004*).

Here using available information on past earthquakes, we focus on the central part of the Himalayas between 79-93°E and 25-31°N, *i.e.* from western Nepal to eastern Bhutan (Fig. 6.2) to calculate the Coulomb stress change since the great historical rupture(s) reported in Nepal and in Assam region around A.D. 1100 (*Lavé et al. 2005*; *Kumar et al. 2010*). We first summarize constraints on both major seismic rupture parameters and inter-seismic coupling of the seismogenic MHT. We pay a particular attention to the Bhutan

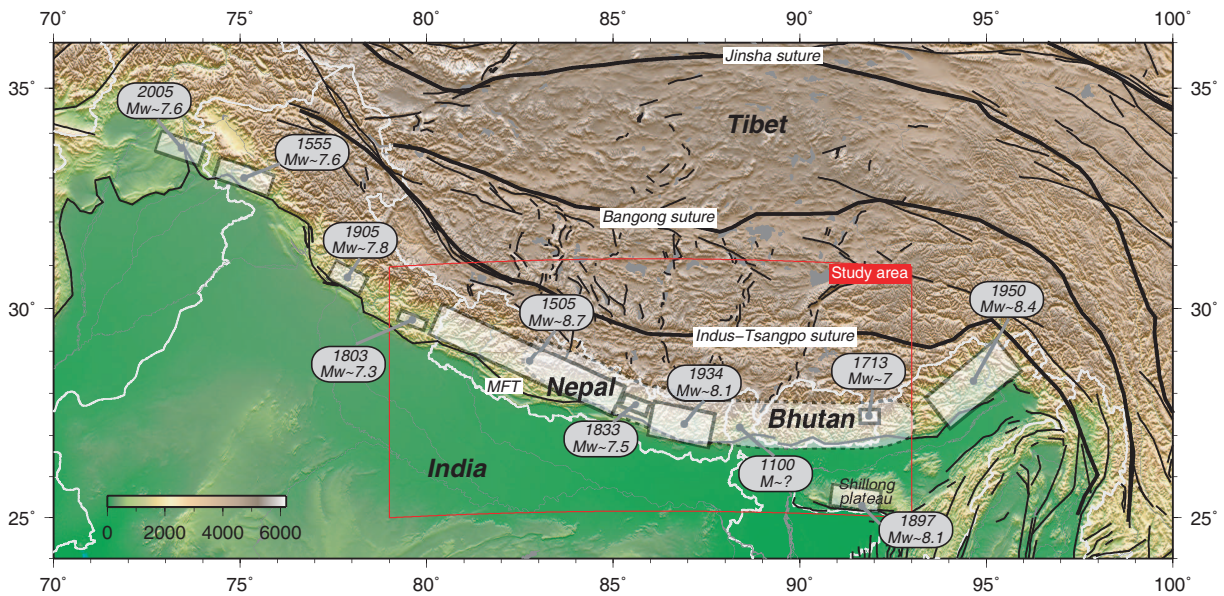


FIGURE 6.1 Simplified map of active faults in northern India, Nepal, and Tibet (adapted from *Replumaz and Tapponnier (2003)*). The Main Frontal Thrust (MFT) and inferred rupture extent (shaded and labelled with year) of major historical earthquakes along the ~2500 km Himalayan arc are shown. The rupture extent of the 1505 central Himalayan earthquake (Mw~8.7), the 1555 Kashmir earthquake (Mw~7.6), the 1803 Kumaon-Garhwal earthquake (Mw ~7.3), the 1833 Nepal earthquake (Mw~7.5), the 1905 Kangra earthquake (Mw~7.8), the 1934 Bihar-Nepal earthquake (Mw~8.1), the 1950 Assam earthquake (Mw~8.4), and the 2005 Kashmir earthquake (Mw~7.6) are adapted from published literature (*Pandey and Molnar 1988; Molnar and Pandey 1989; Chander 1989; Ambraseys and Bilham 2000; Ambraseys and Jackson 2003; Ambraseys and Douglas 2004; Bilham 1995, 2004; Wallace et al. 2005; Szeliga et al. 2010*). The rupture zone of the 1100 earthquake is speculative (dotted contour) and discussed in detail in the text. The colour scale gives the elevation in meter. The red rectangle indicates the extent of the study area.

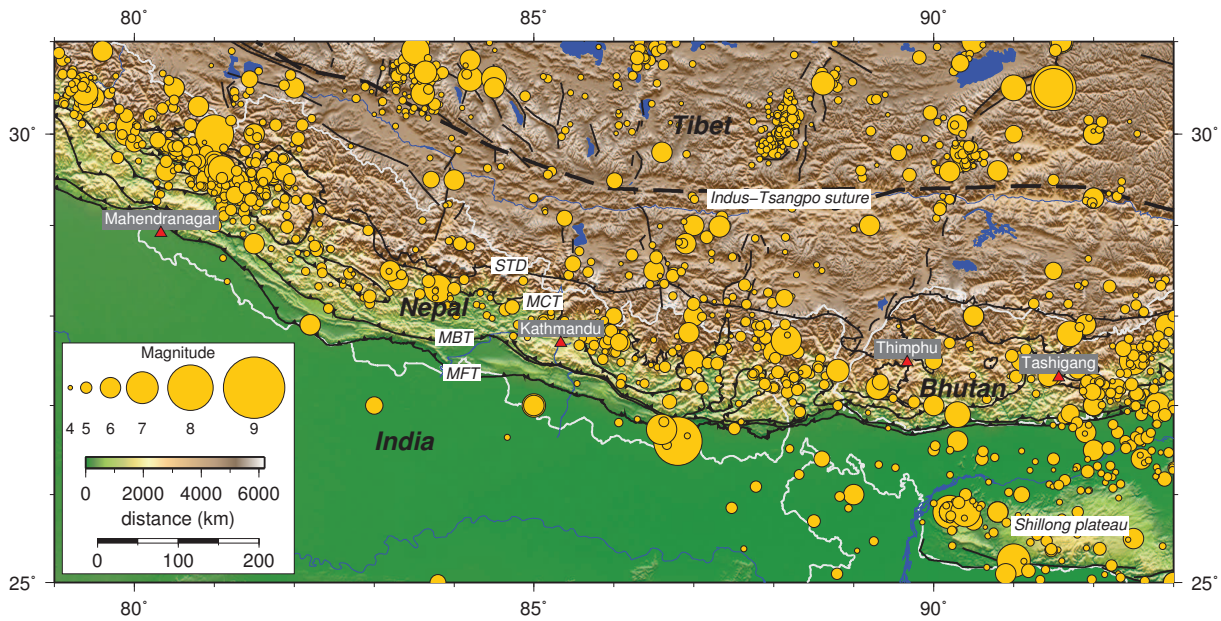


FIGURE 6.2 Topographic map of the study area showing instrumental seismicity recorded between 1913 and 2013. Circles indicate location and magnitude of earthquakes from National Earthquake Information Center -NEIC catalogue (<http://earthquake.usgs.gov/regional/neic/>). The colour scale gives the elevation in meter. Black lines represent major faults : MFT, Main Frontal Thrust. MBT, Main Boundary Thrust. MCT, Main Central Thrust. STD, South Tibetan Detachment.

Himalaya, for which the properties of the seismogenic MHT are mostly unknown. Next we present the assumptions and the limitations of the modelling approach. We discuss the relative effect of both secular loading and co-seismic ruptures on the stress change along the seismogenic MHT. Finally the resulting Coulomb failure stress in Bhutan is compared with the present-day seismicity rate.

6.2.2 Large historical Himalayan earthquakes

Many large historical earthquakes that occurred along the Himalayan arc are now documented. However there are still many uncertainties in the rupture parameters associated with these events. Hence in the following we will only focus on great $M > 7.3$ earthquakes occurring inside or adjacent to the study area between longitude 76°E and 96°E and for which co-seismic slip and rupture extents are available (Tab. 6.1). In the following we first present the instrumentally observed earthquakes and next discuss the rupture parameters estimated for major events that have taken place along the Himalayan arc before 1897.

6.2.2.1 Instrumental earthquakes

The 1950 Assam earthquake The 1950 Assam earthquake is the largest intra-continental earthquake recorded in history (*Triep and Sykes 1997*). It occurred on August 15, 1950 and its epicentre coordinates are 28°23'N, 96°41'E. Unfortunately, due to the lack of local seismic array, the location and the size of the rupture zone as well as the mechanism type are poorly determined for this giant event. Using teleseismic records *Ben-Menahem et al. (1974)* propose a strike-slip rupture of ~35 m on 250 km long and 80 km wide fault striking 330-337°N and dipping 55-60° to ENE. In contrast *Chen and Molnar (1977)* relocated aftershocks and prefer a low-angle fault dipping to the north. They propose a complex pattern of a thrust faulting that implies a rupture zone ~250 km long and ~100 km wide, with an increase of the strike-slip component at its eastern end. These two different solutions for the Assam earthquake will be tested in the following.

The 1934 Bihar-Nepal earthquake With more than 10 000 deaths, this event is one of the most disastrous earthquakes in India's history. Using both teleseismic records and felt intensities maps *Chen and Molnar (1977)* propose an average slip of ~5.4 m on a plane of 130 km × 50 km, corresponding to $M_w=8.0$. *Ambraseys and Douglas (2004)* calculate from detailed isoseismal maps M_w 8.11 for this event corresponding to dimensions of 150 km × 80 km and a slip of 5 m. Here, following *Hough and Bilham (2008)* and the recent finding of *Sapkota et al. (2013)* we assume a thrust faulting rupture of ~5 m on a fault 150 km long and 90 km wide reaching the surface and corresponding to the upper part of the MHT.

The 1930 Dhubri earthquake This earthquake has been poorly studied and most of what we know about this event comes from the report of *Gee (1934)*. Located near the Garo Hills in the western part of the Shillong plateau, this earthquake was felt over a wide area including northeastern India, Bangladesh, eastern Nepal and Bhutan. Both isoseismal map and the low casualty rate associated with this event suggest a relatively deep focus. A depth of 35 km is reported in the centennial earthquake catalogue (*Engdahl and Villaseñor 2002*) whereas *Kayal (2008)* proposed a depth of ~60 km. To our knowledge the geometry of the rupture is also unknown. Based on local geology *Kayal (2008)* proposes that this event occurred along the Dhubri-Jamuna fault, but additional studies are required to confirm this hypothesis. Given our ignorance concerning the ruptures parameters of this

earthquake and the distance from our study area, the effect of the Dhubri earthquake will be neglected in the following.

The 1905 Kangra earthquake The 1905 Kangra earthquake, in terms of death toll is one of the most severe events that occurred in Himalayas over the last two centuries. *Ambraseys and Bilham (2000)* estimate a magnitude of $M_S \sim 7.8$ from teleseismic records. Furthermore using re-measurements of historic triangulation points, the location of inter-seismic locking zone and global scaling laws, *Wallace et al. (2005)* suggest for this earthquake a $100 \text{ km} \times 55 \text{ km}$ NE dipping rupture with a uniform slip of $\sim 4 \text{ m}$ that does not reach the surface. They also propose a small strike-slip component. Thus, in the following, we assume a rake of 80° , leading to a reverse and sinistral slip of 3.9 m and 0.7 m , respectively.

The 1897 Shillong earthquake The re-evaluated instrumental data including geodetic observations and seismographs indicate that the great 1897 Shillong earthquake was a $M_S = 8 \pm 0.1$ event which occurred beneath the Shillong Plateau (*Ambraseys and Bilham 2000; Bilham and England 2001*). Two-end member rupture models have been proposed for this event. Based on the qualitative observations of the earthquake effects, the first model assumes a $170 \text{ km} \times 100 \text{ km}$ rupture occurring on a gently north-dipping thrust fault between 15 and 23 km depth (*Gahalaut and Chander 1992*). The reverse and sinistral slip is estimated to be 4 m and 1 m , respectively. In contrast, the second model based on geodetic data suggests that the earthquake occurred on a steep south dipping thrust fault beneath the northern edge of the Shillong Plateau (*Bilham and England 2001*). The 1897 rupture is inferred to have extended from 9 km to 45 km depth on a 110 km long fault that slipped of $\sim 16 \pm 5 \text{ m}$ with a rake of 76° . Following the approach of *Gahalaut et al. (2011)* these two different rupture models will be considered in this study.

6.2.2.2 19th century earthquakes

Little is known about Himalayan earthquakes before the 1897 Shillong event. A step-by-step approach is thus used to estimate rupture parameters of these great earthquakes. First : following published studies based on isoseismal maps we assign the epicentre location and the magnitude of each event (e.g. *Bilham 1995; Ambraseys and Jackson 2003; Rajendran and Rajendran 2005; Szeliga et al. 2010*). Second : using scaling laws we de-

fine the length and width of the fault rupture as well as the average co-seismic slip from the magnitude (e.g. *Kanamori and Anderson 1975*; *Wells and Coppersmith 1994*; *Blaser et al. 2010*). We then assume that all these events occur in purely reverse ruptures. Finally, using our knowledge of the deep geometry of the MHT we fix the depth, the dip angle and the strike of the rupture plane.

The 1833 and 1866 Kathmandu earthquakes Based on felt intensities and epicentral locations *Bilham (1995)*, *Rajendran and Rajendran (2005)* and *Szeliga et al. (2010)* infer two earthquakes in 1833 and 1866 with similar locations nearly 80 km east-southeast of Kathmandu and similar magnitudes of $\sim 7.5 \pm 0.2$. Using scaling laws (*Kanamori and Anderson 1975*; *Blaser et al. 2010*), we assume for these two earthquakes a similar 80 km \times 40 km NE dipping rupture with a uniform slip of ~ 2 m.

The 1803 Kumaon-Garhwal earthquake *Szeliga et al. (2010)* assign for this event a magnitude of 7.3 with an epicentre location of 30.656°N and 78.784°E. We assign from scaling laws (*Kanamori and Anderson 1975*; *Blaser et al. 2010*) a 70 km \times 35 km NE dipping rupture with a uniform slip of ~ 1.9 m. Knowing the geometry of the seismogenic MHT, the epicentre location 50 km south from the Himalayan front suggests a deep focus. In the following we assume that the rupture have extended from 13 km to 20 km depth.

6.2.2.3 Prior 19th century earthquakes

The 1713 Bhutan earthquake The accurate location as well as the magnitude of the great earthquake that occurred in eastern Bhutan or in Arunachal Pradesh in 1713 is mostly unknown. Based on one Tibetan eyewitness (*Jackson 2002*), this event is reported in numerous contemporary Bhutanese sources. The only estimate suggests a magnitude $M_S = 7$ earthquake (*Ambraseys and Jackson 2003*). As for the 19th century earthquakes, we use scaling laws to assess both surface rupture geometry and displacement (Tab. 6.1).

The 1505 Central Himalaya earthquake Historic accounts of the 1505 earthquake describe strong shaking over a 600-km-long stretch of the central Himalaya from Kathmandu in the east to Almora in the west (*Ambraseys and Jackson 2003*). In his Memoir Emperor Baber, who was camping near Kabul at that time recorded this event. Sikandar Shah established his new capital at Agra in 1504. In the very next year an earthquake da-

amaged the new city of Agra. A magnitude for this event is estimated to be $8.5 < M_w < 8.8$ based on the maximum extent of felt strong shaking, the 70-100 km down-dip width of the locked portion of the basal detachment, and an average slip of 7-15 m (e.g. *Yule et al. 2007*).

The 1255 Eastern Nepal earthquake This event is the first recorded earthquake in history of Nepal. It took place on June 7, 1255. One third of the total population of Kathmandu was killed including Abhaya Malla, the King of Kathmandu valley. Numerous buildings and temples of the valley were entirely destroyed while many of them were severely damaged. Location as well as size of the rupture is unknown. *Sapkota et al. (2013)* assume rupture parameters similar to those of the 1934 Bihar-Nepal earthquake. However this assumption cannot be tested because their study is restricted to the Sir Valley area only. Here we rather favor the use of a model based on additional informations on river terraces and seismites in western Nepal (*Rajendran and Rajendran 2011; Mugnier et al. 2011*), in which the 1255 event is related to a co-seismic slip of 8 m on a 500 km long rupture.

The 1100 medieval earthquake(s) This event is documented from paleo-seismic trench studies only but at several different locations. In east-central Nepal in the Marha Khola region *Lavé et al. (2005)* indicate a large earthquake that ruptured the MFT at ~ 1100 , with a surface displacement of $\sim 17 (+5/-3)$ m. Trenching on the MFT nearby Kakarbhitta at the eastern border of Nepal suggests surface displacement of 4-8 m associated to a large earthquake that occurred between 1050 and ~ 1300 (*Nakata et al. 1998*). More eastward two trench sites immediately west and east of the Kingdom of Bhutan reveal large displacement reaching as much as ~ 18 m that occurred in 1049-1435 and 1025-1234, respectively (*Kumar et al. 2010*). Based on the large observed displacements and the consistency of the estimated timing it is conceivable that all these ruptures are related to one megaquake. In that case the estimated rupture length would approach ~ 1000 km, comparable to the length of the greatest oceanic subduction earthquakes. However, due to dating uncertainties we cannot rule out a more complex scenario of two large ruptures spanning ~ 200 years and occurring on either side of the Kingdom of Bhutan. In the following these two end-member models will be tested.

TABLE 6.1 Rupture parameters of major earthquakes following *Chen and Molnar (1977)*; *Ben-Menahem et al. (1974)*; *Gahalaut and Chander (1992)*; *Bilham (1995)*; *Bilham and England (2001)*; *Ambraseys and Jackson (2003)*; *Ambraseys and Douglas (2004)*; *Wallace et al. (2005)*; *Bilham and Ambraseys (2004)*; *Lavé et al. (2005)*; *Kumar et al. (2006)*; *Hough and Bilham (2008)*; *Szeliga et al. (2010)*; *Kumar et al. (2010)*; *Gahalaut et al. (2011)*; *Mugnier et al. (2011)*. M, magnitude; Lon, longitude of the leading edge of rupture plane projected vertically to surface; Lat, latitude of the leading edge of rupture plane projected vertically to surface; L , rupture plane length; W , rupture plane width; S , rupture plane strike; D , rupture plane dip; U reverse slip; V right lateral slip; z_t , depth of the top of the rupture plane; z_b , depth of the bottom of the rupture plane. (1) from *Chen and Molnar (1977)*. (2) from *Ben-Menahem et al. (1974)*. (3) from *Gahalaut and Chander (1992)*. (4) from *Bilham and England (2001)*.

Location	Date	M	Lon (°N)	Lat (°E)	L (km)	W (km)	S (°)	D (°)	U (m)	V (m)	z_t (km)	z_b (km)
Assam ⁽¹⁾	15 Aug. 1950	8.4	96.61	27.9	250	100	317	10	20	0	~ 0	17
Assam ⁽²⁾	15 Aug. 1950	8.4	97.5	27.83	250	80	330	55	0.642	30	~ 0	65
Bihar-Nepal	15 Jan. 1934	8.1	87	26.8	150	90	285	11.3	5	0	~ 0	20
Kangra	4 Apr. 1905	7.8	76.65	31.3	100	55	335	7	3.9	-0.7	5	11.7
Shillong ⁽³⁾	12 Jun. 1897	8.1	90.12	25.27	170	100	280	5	4	-1	15	23
Shillong ⁽⁴⁾	12 Jun. 1897	8.1	90.87	25.91	110	43	112.5	57	16	-2.679	9	45
Nepal	23 May 1866	7.5	85.11	27.55	80	40	280	11.3	2	0	5	13
Nepal	25 Aug. 1833	7.5	85.11	27.55	80	40	280	11.3	2	0	5	13
Garhwal	01 Sep. 1803	7.4	79.21	29.96	70	35	300	11.3	1.9	0	13	20
Bhutan	spring 1713	7	92	27.4	40	22	270	11.3	1.2	0	15	20
Central Himalaya	6 Jun. 1505	8.7	84.72	27.2	600	100	295	11.3	10	0	~ 0	20
Nepal	7 Jun. 1255	8.5	87	26.8	500	90	285	11.3	8	0	~ 0	20
Nepal	~ 1100	8.8	88.45	26.75	400	100	280	11.3	17	0	~ 0	20
Assam	~ 1100	8.4	93.05	27	100	100	260	11.3	17	0	~ 0	20
Nepal-Bhutan	~ 1100	9.1	93.05	26.7	1000	100	275	11.3	17	0	~ 0	20

6.2.3 Secular loading

Several previous studies have demonstrated from GPS measurements and levelling data that the seismogenic MHT is fully coupled in both Nepal and Sikkim (e.g. *Bilham et al. 1997*; *Cattin and Avouac 2000*; *Mullick et al. 2009*; *Ader et al. 2012*). In these regions inter-seismic loading is thus related to aseismic slip of 17-22 mm/yr along the deep portion of the MHT beneath the Higher Himalaya and southern Tibet, while the seismogenic MHT remains locked to the south. In contrast, our knowledge of secular loading in Bhutan is still low. *Drukpa et al. (2012)* proposed a convergence rate of ~ 20 mm/yr across Bhutan, but the inter-seismic coupling of the seismogenic MHT in this area remains unconstrained. Compared to the Nepal and Sikkim Himalayas, we have shown above that until now no major historical earthquake has been documented in Bhutan. At shorter time scales, *Gahalaut et al. (2011)* show that Bhutan is also characterized by a low present-day seismicity rate. Together these two observations may reflect two opposite fault behaviours in terms of coupling : a creeping zone or a region of stress accumulation for future great earthquakes. New GPS observations in Bhutan measured between 2003 and 2013 suggest that the seismogenic MHT is fully locked in the western part of Bhutan (*Drukpa et al. 2012*; *Vernant et al. 2013*). In central and east Bhutan the GPS network has been densified in March 2013 only. The associated time period is thus too short to assess inter-seismic coupling. One can note that the region near the MFT in Bhutan exhibits higher elevation and gravity anomaly with respect to Nepal and Sikkim (*Duncan et al. 2003*; *Hammer et al. 2013*). At geological time-scales, this may reflect longitudinal variations in the frictional properties of the plate interface as proposed by *Song and Simons (2003)* in subduction zones. Testing this hypothesis requires a systematic study of lateral variation in both topography and gravity anomaly, which is out of the scope of the present paper. In the following the effect of longitudinal variation of inter-seismic coupling in Bhutan will be tested by assuming two end-member models : (1) a fully locked seismogenic MHT along the Himalayan arc and (2) a model with a creeping zone in central and eastern Bhutan.

6.2.4 Modelling approach

To assess the stress change pattern along the Himalayan arc over the last 900 years we develop a 3D modelling approach to calculate the stress change associated to both past

events and secular tectonic loading.

6.2.4.1 Modelling of earthquakes and secular loading

Co-seismic slips due to past events are imposed using dislocations embedded into an isotropic homogeneous elastic half-space (*Okada 1992*). As previously mentioned the rupture parameters of these events are partly poorly constrained. For some earthquakes several scenarios have been proposed and will be tested.

The geometry of our model is defined with a great circle centre at 89°E, 40.2°N with an outer radius of 1508 km, spanning from 76°E to 96°E. The inter-seismic loading is modelled considering that most of the convergence is accommodated by a constant reverse slip along the aseismic portion of the MHT below 20 km depth. Following *Vergne et al. (2001)* the inter-seismic stress is then calculated from 75 buried creeping dislocations dipping 11.3° northward. For simplicity we will assume a constant reverse slip of 20 mm/yr along the Himalayan arc with no temporal or lateral variations. The co-seismic ruptures as well as the creeping zones in Bhutan are applied on the seismogenic MHT, for which we assume a constant dip angle of 11.3° between 0 and 20 km depth and a surface trace associated with the MFT (e.g. *Schelling and Arita 1991; Cattin and Avouac 2000; Nabelek et al. 2009; Ader et al. 2012*). The rupture parameters are defined in table 6.1. The arcuate shape of the MFT is taken into account using 260 patches for which the size increases with depth from 1 to 40 km to better assess stress change at shallow depth.

6.2.4.2 Coulomb stress change calculation

To calculate static stress change we assume an elastic half-space with a Young modulus and a Poisson ratio of 80 GPa and 0.25, respectively. Using the Coulomb 3.3 software (*Lin and Stein 2004; Toda et al. 2005*), we then calculate the associated Coulomb stress change

$$\Delta CFF = \Delta\tau + \mu' \Delta\sigma, \quad (6.1)$$

where $\Delta\tau$, $\Delta\sigma$ and μ' are the shear stress change on the receiver fault, the normal stress change on the receiver fault and the effective friction coefficient, respectively. The variation of friction coefficient between 0.4 and 0.8 does not affect significantly our results. Thus, in the following we will use an effective friction coefficient of 0.4, which is a commonly used

value in many stress-triggering studies (*King et al. 1994*). We consider the seismogenic MHT as the receiver fault.

As many stress triggering studies, the proposed modelling approach has several limitations : inelastic behaviour, post-seismic fluid flow, spatial variations in rheological parameters and dynamic stress changes are neglected. Given all these limitations and the uncertainties on ruptures parameters, here we do not attempt to obtain a unique distribution of the Coulomb stress change along the Himalayan arc. Thus we will rather discuss the main features of the Coulomb stress change distribution by focusing on the relative effects of co- and inter-seismic stress change associated with the different scenarios proposed above.

6.2.5 Results

6.2.5.1 Inter- and co-seismic stress change

Assuming a fully locked seismogenic MHT along the Himalayan arc our results suggest a Coulomb stress increase rate ranging from 0.002 bar/yr at shallow depth to 0.05 bar/yr at 20 km depth (Fig 6.3a). These values are in agreement with previous results, which, in the Himalaya of Central Nepal, give a ΔCFF increase rate between 0.008 bar/yr and 0.06 bar/yr on the shallow and the deep portion of the seismogenic MHT, respectively (*Cattin and Avouac 2000; Bollinger et al. 2004*). Taken into account a convergence rate between 17 and 22 mm/yr leads to a ΔCFF rate increase of 0.002 ± 0.0002 bar/yr at shallow depth and 0.049 ± 0.006 bar/yr at 20 km depth.

A creeping zone in central-eastern Bhutan has a major effect on the state of stress on the seismogenic MHT in this region. It leads to a ΔCFF decrease rate that can reach -0.5 bar/yr at the surface (Fig 6.3b). In the central part of the unlocked zone our results give a ΔCFF decrease rate of 0.007 ± 0.002 bar/yr. One can also note that the creeping zone leads to an additional loading of 0.005 ± 0.004 bar/yr in western Bhutan. In the other regions of the Himalayan arc the effect of a creeping zone in Bhutan is very low ($< 10^{-4}$ bar/yr).

For each historic event, the rupture zone is associated to a ΔCFF decrease in the range of 10-100 bar, whereas ΔCFF increases by up to 10 bar in the neighbouring patches (Fig. 6.4–6.7). For distances greater than ~ 100 km from the rupture fault plane the effect of

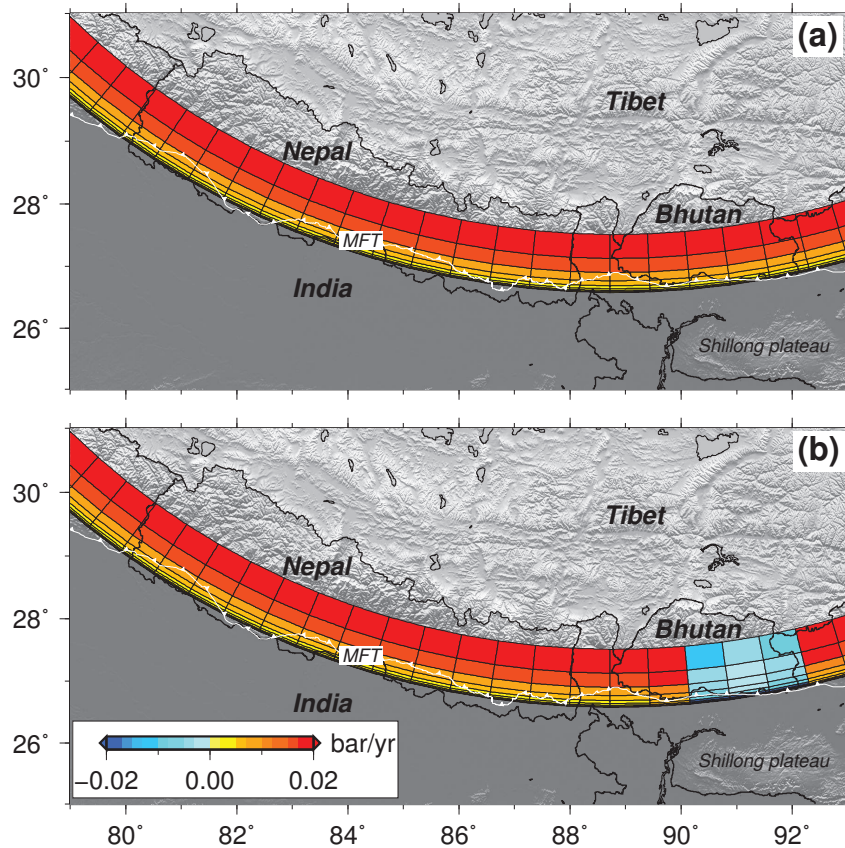


FIGURE 6.3 Rate of the coulomb stress change ΔCFF along the Main Himalayan Thrust fault (MHT) due to inter-seismic loading of reverse faulting failure planes. Inter-seismic loading is associated to a reverse slip of 20 mm/yr along dislocations dipping 11.3° northward. (a) Map showing the pattern of ΔCFF rate assuming that the MHT is fully locked along the Himalayan arc. The location of the Main Frontal Thrust (MFT) is also given. (b) Same as (a) except that the central-eastern part of the seismogenic MHT in Bhutan is unlocked.

a co-seismic stress change is < 0.1 bar. Therefore, the 1905 rupture and the 1950 event have very small effect on ΔCFF within our study area, regardless of the model used for the Assam earthquake.

Our results suggest no obvious evidence for stress-triggered earthquakes considering a time scale of a few decades. For instance the 1255 and 1934 earthquakes cannot be explained by stress transfer due to prior events. However given all the uncertainties on ruptures parameters, we cannot rule out some relationship between the 1100 and the 1505 events or between the 1255 ruptures and 1833-1866 earthquakes.

6.2.5.2 Cumulated stress change

In a first time two end-member models have been tested to study the cumulative Coulomb stress change since the 1100 event. In the first one, model 1 hereinafter, we assume (1) no lateral variation in the coupling along the seismogenic MHT and (2) a single rupture for the 1100 earthquake. In that case the behaviour of the Bhutan Himalaya is similar to Nepal and Sikkim (Fig. 6.4, 6.5). The present-day cumulated ΔCFF change along the seismogenic MHT is mostly controlled by the medieval earthquakes and the inter-seismic loading over the 900 years. This stress change is heterogeneous with three distinct zones : one in the western part of our study area in Uttarakhand and western Nepal with $\Delta CFF > 10$ bar, one in central with a $\Delta CFF < -5$ bar and one in Far-East Nepal, Sikkim and Bhutan with low ΔCFF between -2 and 2 bar.

In the second model, model 2, we assume lateral variations along the Himalayan arc with (1) a creeping zone in central-eastern Bhutan and (2) two large ruptures occurring on either sides of Bhutan around 1100. In that case the cumulative stress change distribution associated to both inter-seismic loading and historical earthquakes over the last 900 years is more complex (Fig. 6.6, 6.7). Although the present-day state of stress in the western part of our study area differs slightly in the two models, the corresponding Coulomb stress increase remains high. In contrast, in the central and eastern part of the study area the two scenarios give very different results. In the second model, the region between 84° and 93° of longitude is associated to a Coulomb stress decrease area with several patches of positive ΔCFF in western Bhutan and on the sides of the 1713, 1833, 1866 and 1934 ruptures.

Two additional models can be tested to unravel the effect of both the 1100 earthquake

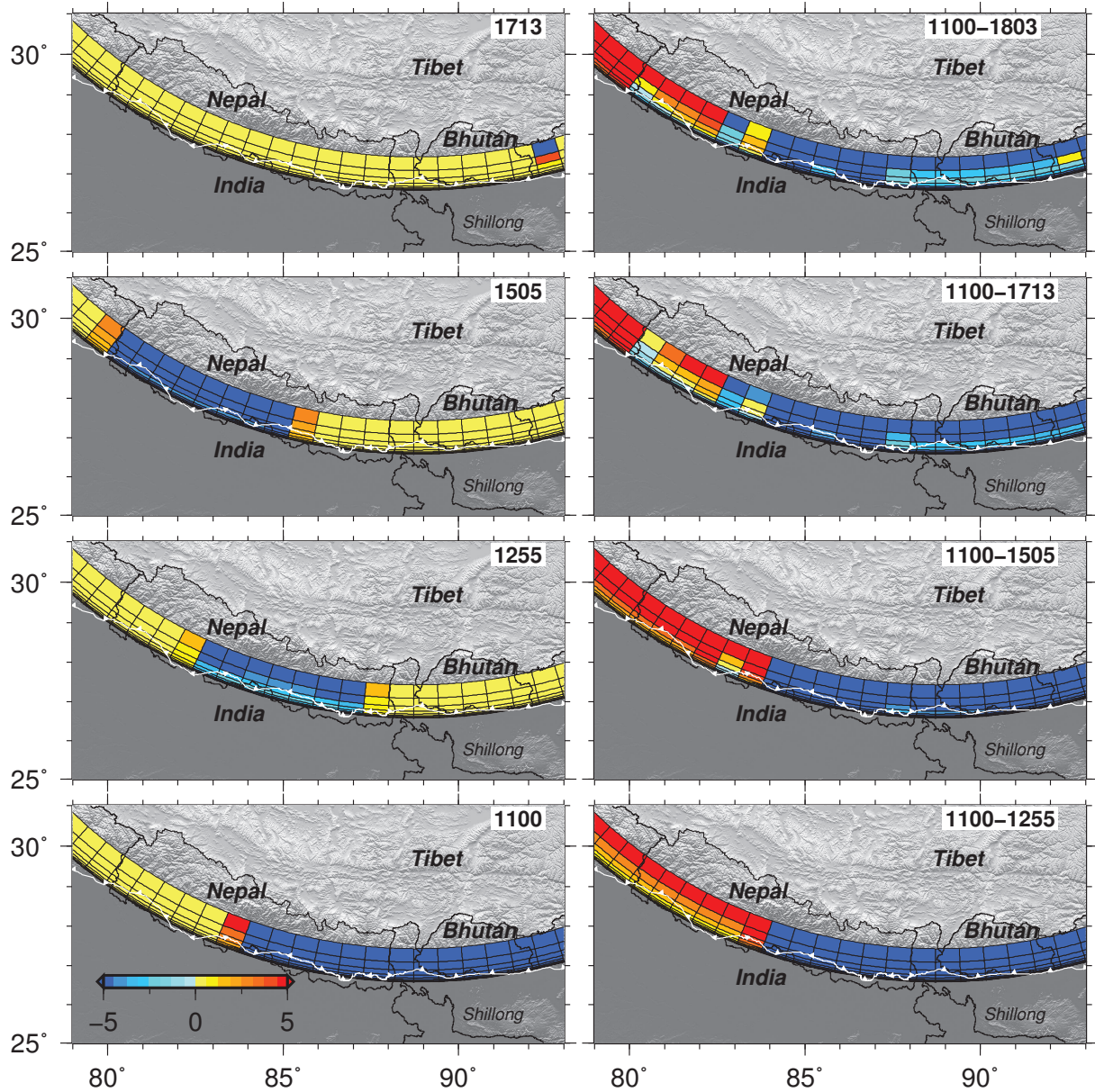


FIGURE 6.4 Coulomb stress change ΔCFF in bar along the MHT associated to both secular loading and the largest earthquakes before the 19th century. We assume that (1) the MHT is fully locked during inter-seismic loading and (2) the 1100 event is related to only one rupture. Date and time are indicated in the white box. Colour scale gives ΔCFF in bar. On the left, ΔCFF due to each earthquake. On the right, cumulated ΔCFF over the indicated time period.

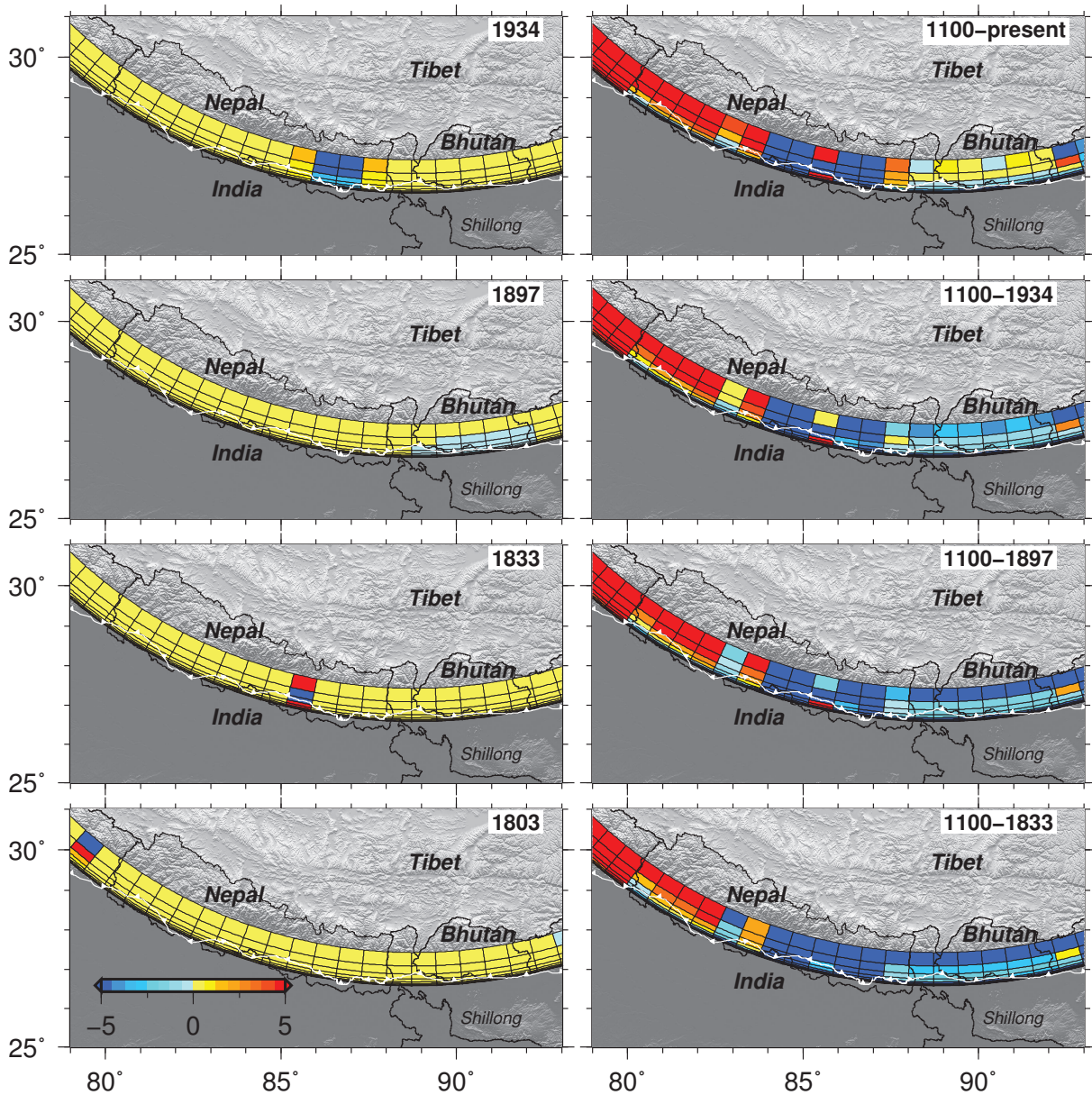


FIGURE 6.5 Same as figure 6.4 for the last two centuries. The 1866, 1905 and 1950 earthquakes are taken into account in the calculation of the cumulated ΔCFF .

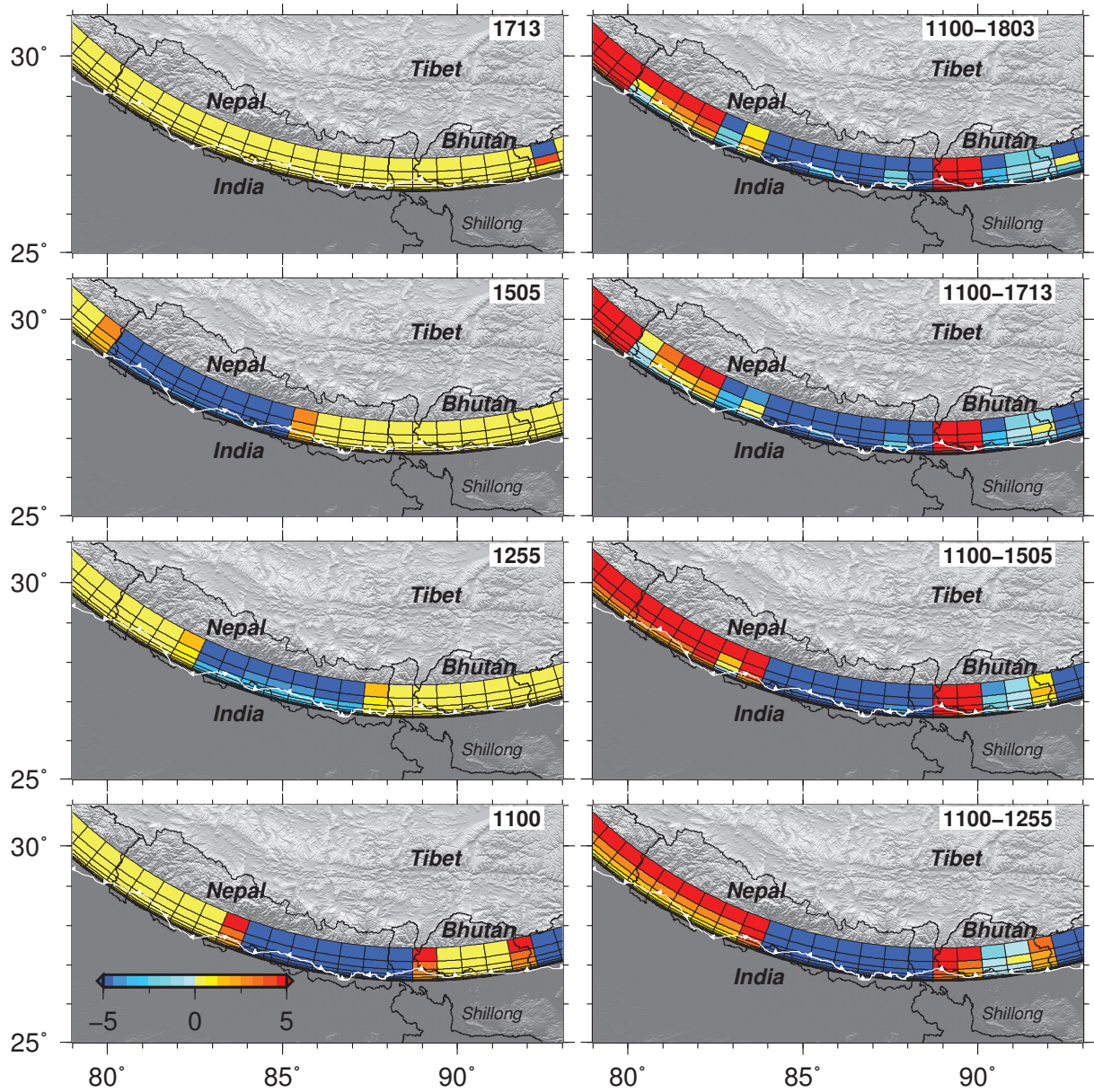


FIGURE 6.6 Same as figure 6.4 except that (1) the MHT is unlocked during inter-seismic loading in western Bhutan and (2) the 1100 event is modelled with two separate rupture zones.

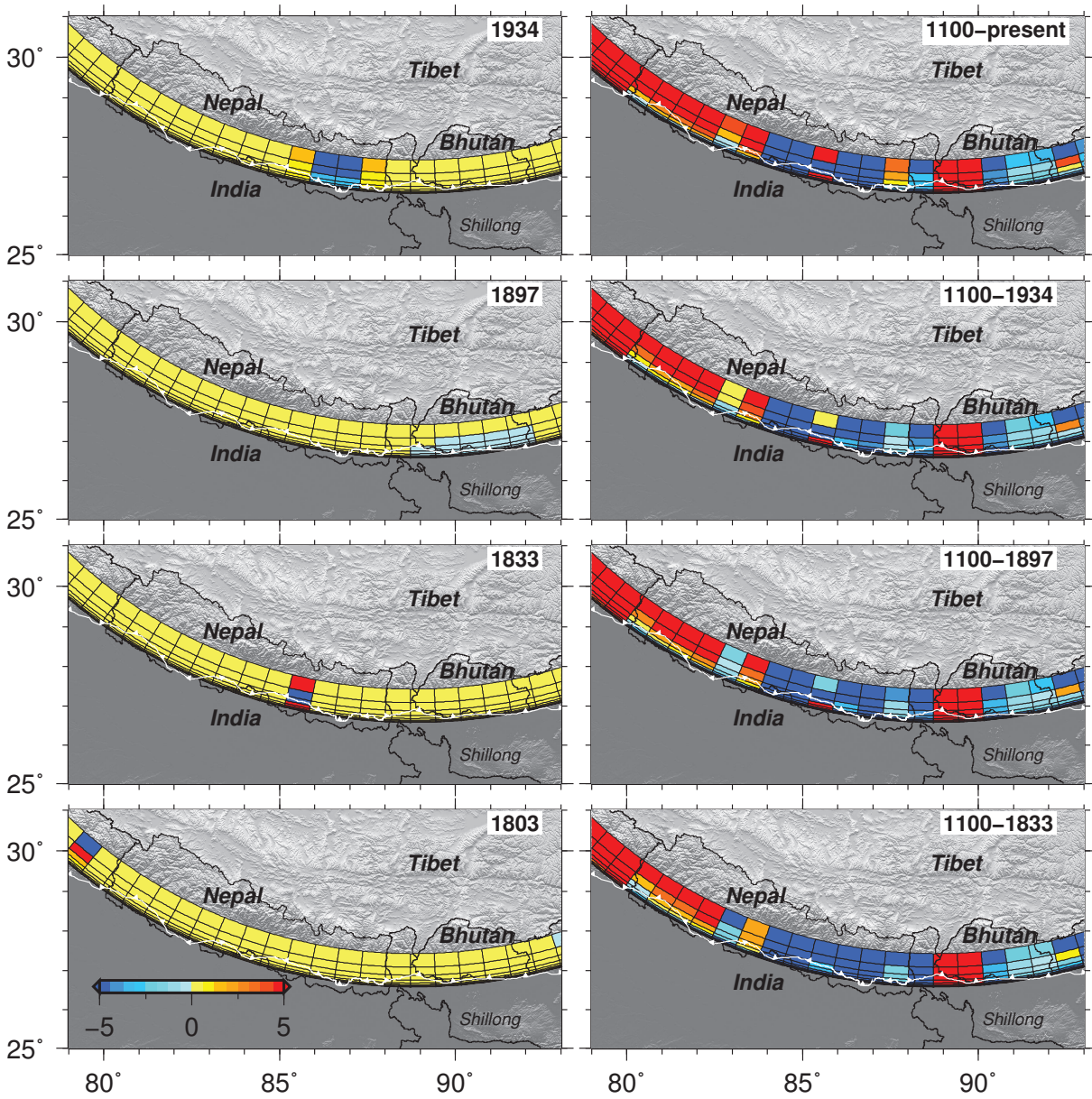


FIGURE 6.7 Same as figure 6.6 for the last two centuries. The 1866, 1905 and 1950 earthquakes are taken into account in the calculation of the cumulated ΔCFF .

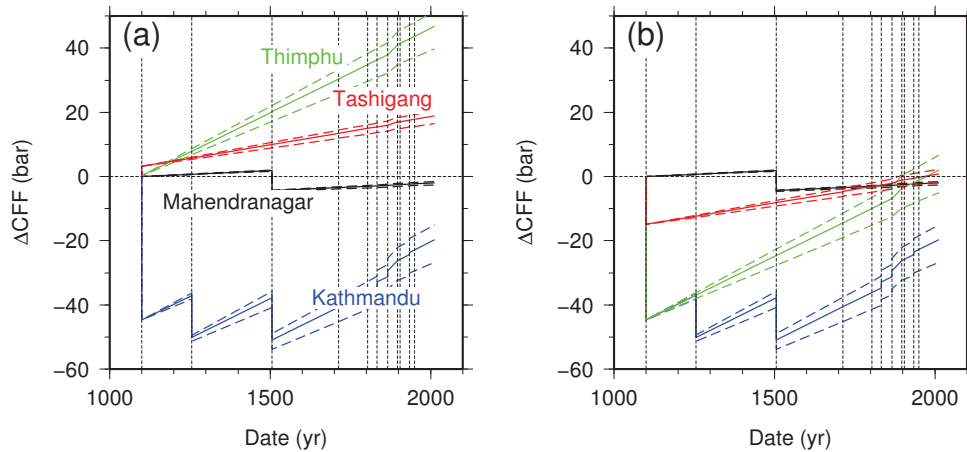


FIGURE 6.8 Time evolution of ΔCFF over the last nine centuries near major Nepalese and Bhutanese cities (see location on figure 6.2), assuming that the MHT is fully locked during inter-seismic loading. Solid and dashed lines are related to a convergence rate of 20 mm/yr and 17-22 mm/yr, respectively. (a) and (b) show the results obtained with the 1100 earthquake modelled with two and one rupture zone, respectively. Vertical dashed lines indicate the time of major earthquakes considered in this study.

rupture and the unlocked MHT in Bhutan. We assume a fully locked MHT and two events around 1100 in model 3, and a creeping zone and a single rupture in 1100 in model 4. The study of the temporal variations of ΔCFF near Mahendranagar, Kathmandu, Thimphu and Tashigang associated with model 1 and model 3 reveals the key role played by the slip distribution of the 1100 event(s). The lack of medieval rupture causes a Coulomb stress accumulation of > 20 bar, leading to a very high seismic hazard in Bhutan (Fig. 6.8). In contrast the Coulomb stress decrease due to a unique rupture in 1100 leads to low present-day ΔCFF nearby these major Nepalese and Bhutanese cities. Finally, in Bhutan the effect of a creeping zone can be studied from models 1–4. Our results suggest that the variation of coupling has a maximal effect near the frontal thrust (Fig. 6.9). Beneath Thimphu the lack of a creeping zone leads to a decrease of ~ 5 bar in the cumulated CFF, whereas beneath Tashigang it causes an increase of ~ 20 bar.

For variations between 17 and 22 mm/yr the effect of the convergence rate can be significant in the lesser Himalaya (± 5 bar). However it does not modify the major trends given in the previous results (Fig. 6.8).

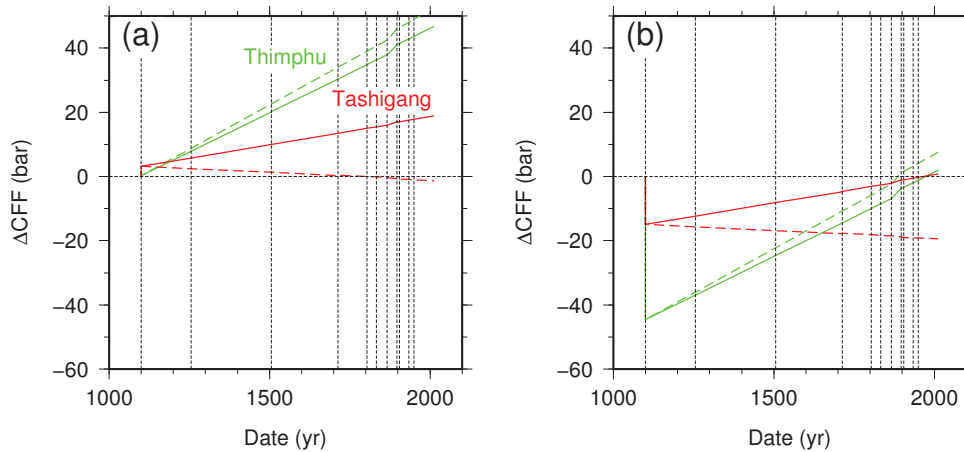


FIGURE 6.9 Time evolution of ΔCFF over the last nine centuries near major Bhutanese cities (see location on figure 6.2), assuming that the MHT is fully locked (solid line) or unlocked (dashed line) in western Bhutan during inter-seismic loading. (a) and (b) show the results obtained with the 1100 earthquake modelled with two and one rupture zone, respectively.

6.2.6 Discussion

6.2.6.1 Present-day seismicity along the Himalayan arc

The study of instrumentally recorded seismicity over the past four decades shows lateral variations in seismic activity along the Himalayan arc (Fig. 6.10). Compared to central Nepal and Sikkim, the regions of western Nepal as well as the western part of Arunachal Pradesh appear to be seismically more active, whereas Bhutan is a low seismicity zone. Nonetheless this observation already done by *Gahalaut et al. (2011)* must be qualified in the light of temporal variations. In fact Bhutan appears as a low seismicity zone in the time period between 1973 and 2003. However over the last decade only central Bhutan (between 90° and 91°) was not affected by earthquakes, whereas western and eastern Bhutan appear as zones of moderate and high seismicity rate, respectively. This high seismicity in eastern Bhutan is mostly due to the 21 September 2009 M_w 6.1 earthquake, which, together with the 2011 Nepal-Sikkim earthquake, is one of the biggest Himalayan earthquakes since 2003.

Compared to our results, part of these lateral variations in the seismicity rate can be interpreted in terms of Coulomb stress change (Fig. 6.11). High seismicity in western Nepal can be associated to an absence of major earthquakes since 1505 leading to a $\Delta CFF > 10$ bar in this area. The cluster of seismicity observed nearby 85.5° may be associated to the stress change of the 1833, 1866 and 1934 earthquakes. The 2011 Sikkim-Nepal earthquake

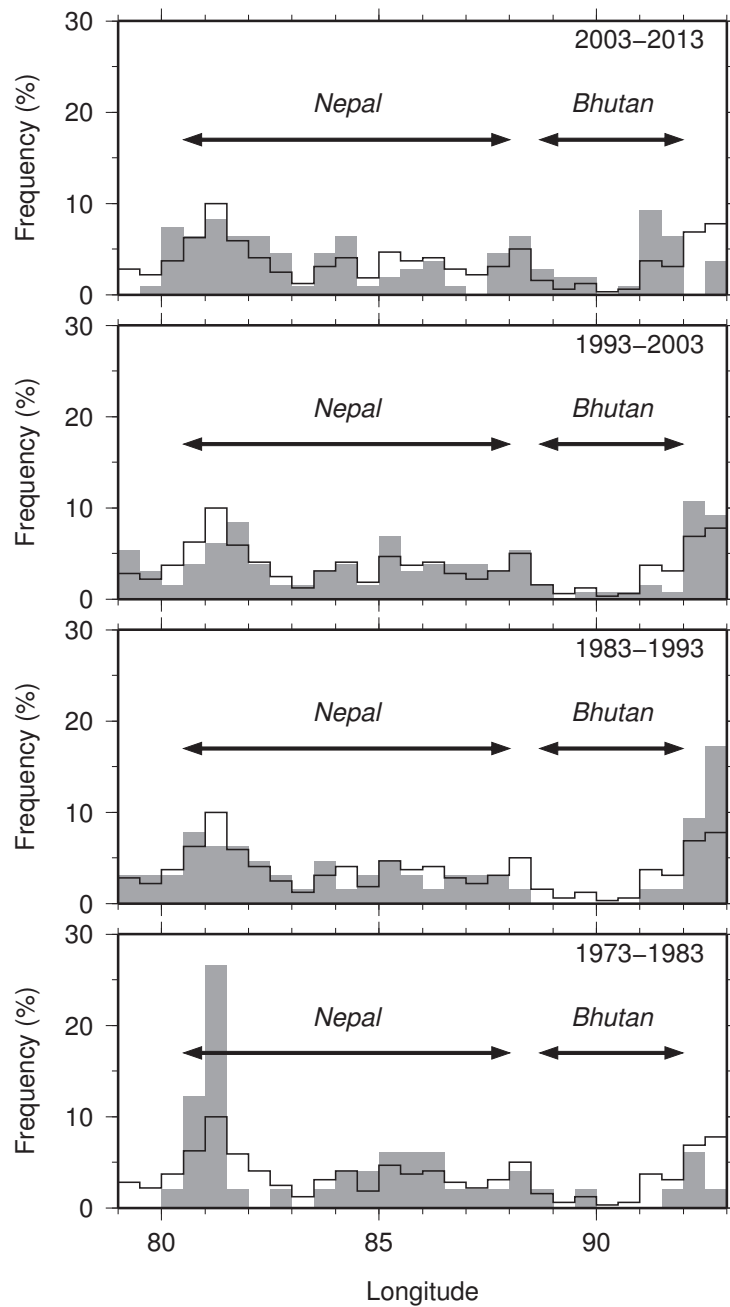


FIGURE 6.10 Longitudinal variation of seismicity over the past four decades along the Main Himalayan Thrust fault that extends from the MFT at the surface to beneath the front of the high range, over a width of 100 km. Gray histogram and black line show the relative frequency of shallow earthquakes (<30 km depth) over the indicated decade and since 1973, respectively. Source : NEIC

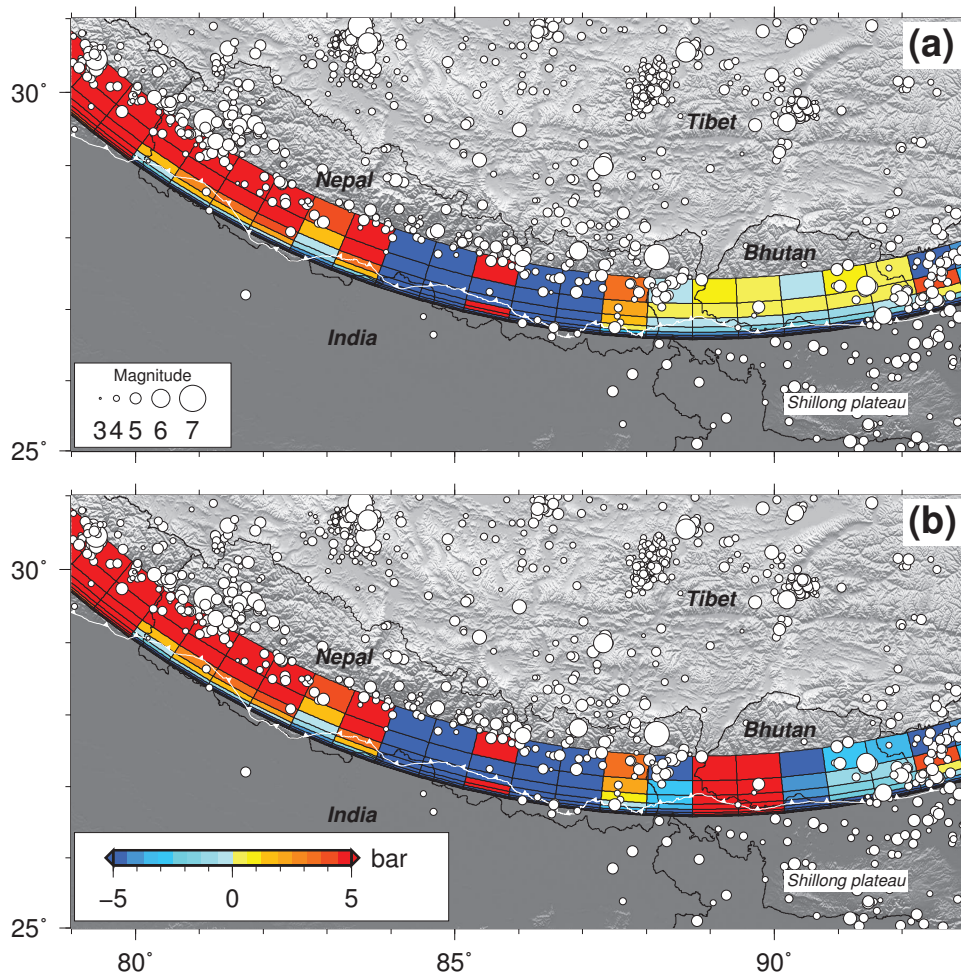


FIGURE 6.11 Topographic map showing the location and magnitude of shallow earthquakes (< 30 km depth) from the NEIC catalogue over the past four decades and the distribution of the cumulated Coulomb stress change over the last 900 years. (a) Model 1 : the MHT is fully locked during inter-seismic loading and the 1100 event is modelled with a single rupture. (b) Model 2 : a creeping zone exists in central-eastern Bhutan and the 1100 event is modelled with two rupture zones on both sides of Bhutan.

may be related to ΔCFF increase after the 1934 Bihar-Nepal earthquake. Even though we cannot rule out partial decoupling of the MHT, the recent seismicity in Bhutan favours a model with no creeping zone in this area.

In contrast, regardless of model assumptions, our results fail to explain the present-day seismicity in central Nepal. The calculated ΔCFF is controlled by the inter-seismic loading and the co-seismic slip distribution of past earthquakes. Both inter-seismic coupling and convergence rate are well-constrained in this segment (*Ader et al. 2012*). Hence the inconsistency of our results with the observed seismicity may be due to missing historical earthquakes in our catalogue or to the simplicity of the slip distribution assumed for the past events : uniform slip and simple fault geometry.

6.2.6.2 Effect of the 1897 Shillong earthquake on the seismicity in Bhutan

Many interpretations, including low convergence rate or transcurrent deformation, have been proposed to interpret the low seismicity in Bhutan (e.g. *Drukpa et al. 2006*; *Velasco et al. 2007*). More recently *Gahalaut et al. (2011)* suggest that this low activity is due to the occurrence the 1897 Shillong earthquake, which leads to a Coulomb stress decrease along the seismogenic MHT in Bhutan. Here, we use our approach to test this last interpretation by calculating the stress change due to both the secular loading and the major earthquakes since 1897.

Following *Gahalaut et al. (2011)*, we test the two models proposed by *Gahalaut and Chander (1992)* and *Bilham and England (2001)* for the 1897 Shillong earthquake. Although the two models differ greatly for the two possible fault plane solutions, the corresponding Coulomb stress change is similar in amplitude (Fig. 6.12a,d). In both cases the 1897 Shillong earthquake causes a Coulomb stress decrease of up to 1 bar with a mean ΔCFF of ~ -0.2 bar. The stress patterns are however slightly different : ΔCFF decrease is localized in the southern Bhutan with the *Gahalaut and Chander's* model, whereas it is localized in the eastern part with the *Bilham and England's* model. Although these results are in agreement with those obtained by *Gahalaut et al. (2011)*, our conclusion differs. We have shown in paragraph 6.2.5.1 that both the 1905 and the 1950 earthquakes have very low effect on the MHT in Bhutan. The Bihar-Nepal 1934 event affects the stress in eastern Nepal only (Fig. 6.12b,c,e,f). However the inter-seismic ΔCFF between 0.002 ± 0.0002 bar/yr at the front and 0.049 ± 0.006 bar/yr at 20 km depth cannot be neglected. Hence, except at shallow depth, our results obtained from model 1 (Fig. 6.12b,e) or model 2 (Fig. 6.12c,f) suggest that the 1897 Shillong plateau earthquake does not have a major role on the stress that accumulated since 1897.

As previously mentioned above for the last decade (Fig. 6.10), the low seismic activity in Bhutan can be apparent due and to the short time scale of the available instrumental earthquake catalogue and the lack of paleo-seismic studies in Bhutan.

6.2.6.3 Slip deficit in the Himalayas

Our results suggest lateral variations in the stress change distribution along the Himalayas. The area with a positive ΔCFF in western Nepal can be interpreted as zones of slip deficit. Several explanations have been already proposed (*Bilham and Ambraseys*

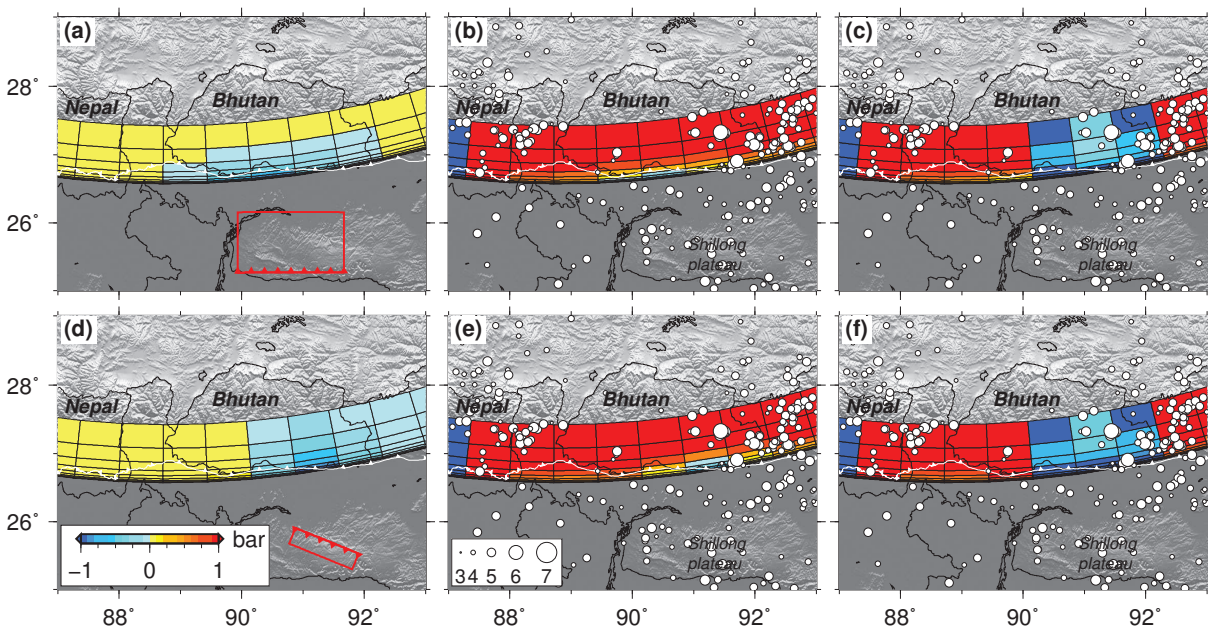


FIGURE 6.12 Coulomb stress change ΔCFF along the MHT of Bhutan. White circles give the magnitude and the location of earthquakes over the past four decades from the NEIC catalogue. Red lines give the projected surface geometry of the 1897 rupture proposed by *Gahalaut and Chander (1992)* (top row) and *Bilham and England (2001)* (bottom row). (a,d) ΔCFF caused by the 1897 earthquake. (b,e) ΔCFF since 1897 taking into account the effect of the 1897 Shillong event, the 1934 Bihar-Nepal earthquake and the 1950 Assam rupture as well as the inter-seismic loading on a locked MHT. (c,f) same as (b,e) except that a creeping zone is assumed in eastern Bhutan.

2004). First : this deficit can be due to the incompleteness of earthquake catalogues and to the uncertainties in the rupture parameters. Second : it can be related to aseismic creep or to slow-slip events. However geodetic measurements in the Nepal Himalaya over the two last decades suggest a homogeneous coupling and no evidence of large slow-slip event has been found (*Ader et al. 2012*).

An alternative explanation can be found when considering the time scale. Based on the comparison of convergence rates estimated from Himalayan earthquakes and GPS observations, *Bilham and Ambraseys (2004)* underline a slip deficit along the Himalayan arc since 1500, whereas our result gives several zones with a negative ΔCFF . This disagreement is only apparent. Ignoring pre-1505 earthquakes leads to $\Delta CFF > 30$ bar in Nepal and Bhutan with the model 1. As mentioned by these authors this slip deficit can be apparent and due to an earthquake cycle duration larger than 500 years in this area. This is corroborated by *Ader et al. (2012)*'s conclusion, who, from the rate of moment deficit accumulation, estimate a return period of large earthquakes exceeding 1000 years in Nepal and propose a maximum magnitude of 9.2 for these events. Hence, as recently proposed by *Goldfinger et al. (2013)*, this suggests long-term seismic cycles for the MHT and emphasizes the need to extend the study of past events to several millennia.

6.2.7 Conclusions

The assessment of stress related to both numerous historical Himalayan earthquakes as well as inter-seismic loading has been used to quantify the temporal evolution of Coulomb stress change distribution along the seismogenic MHT in the past 900 years.

Our results suggest no obvious evidence for major stress-triggered earthquakes. For instance the 1255 and 1934 earthquakes cannot be explained by stress transfer due to prior events. This can be related to over-simplified models used in our approach that are based on elastic dislocation theory. Furthermore we assume a simple geometry for the MHT with a constant dip angle rather than a flat and ramp geometry with lateral variations as previously proposed for the Nepal Himalaya (e.g. *Cattin and Avouac 2000*; *Robert et al. 2011*). Last but not least, many uncertainties regarding past events still exist due to partly poorly constrained slip distribution of ruptures and due to the incompleteness and the limited time window of earthquake catalogues. Hence a relationship between the 1100 and the 1505 events as well as between the 1255 rupture and 1833-1866 earthquakes cannot

be ruled out.

In spite of these uncertainties, our simple approach accounts for the main features of the present-day seismic activity in Himalayas. High seismicity in western Nepal may be related to the lack of major earthquakes since the 1505 events. Furthermore the recent seismicity increase at Nepal-Sikkim border and in eastern Bhutan may be interpreted as the result of the joint effect of the 1934 Bihar-Nepal earthquake and inter-seismic loading of a mostly coupled MHT. More generally, our results suggest that the main patterns of present-day seismicity can be related to the location of past events rather than to major longitudinal variations in the convergence rate or in the inter-seismic coupling of the MHT. This differs significantly from the current understanding of oceanic subduction zone behaviour, in which the seismic coupling is segmented and correlates with the extent of co-seismic ruptures.

This finding is in agreement with the results obtained from GPS measurements in Nepal, Sikkim and western Bhutan (e.g. *Mullick et al. 2009*; *Ader et al. 2012*; *Vernant et al. 2013*). In central-eastern Bhutan new observations suitable to characterize inter-seismic convergence rate and coupling are needed. As previously proposed (e.g. *Kumar et al. 2006*; *Ader et al. 2012*), our results suggest that the lateral stress variations along the MHT are mostly controlled by the occurrence of megaquakes with return periods of 1000-3000 years. This points out the importance of denser spatial paleo-seismological sampling to capture such events and to extend the time-window of our knowledge of the earthquakes along the Himalayan arc.

Acknowledgment

This work was supported by INSU, CNES and Swiss National Science Foundation grants.

CHAPITRE 7

CINÉMATIQUE HOLOCENE DU CHEVAUCHEMENT

FRONTAL BHOUTANAIS

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7.1 Présentation

Suite au repérage d'évidences géomorphologiques de déformations actives au niveau du front Himalayen Bhoutanais pendant les campagnes gravimétriques de 2010 et 2011, nous avons effectué une première étude morphotectonique du 26 Novembre au 16 Décembre 2012. Notre objectif était à la fois de déterminer les taux de déformation moyens, vertical et horizontal, pour la période Holocène au niveau du front Bhoutanais, et de rechercher les traces de ruptures de surface dans cette zone.

Nous avons concentré notre étude dans la partie centrale du Sud Bhoutan où plusieurs niveaux de terrasses alluviales soulevées le long du front chevauchant ont été cartographiés. Afin de dater l'âge de mise en place de ces terrasses, nous les avons échantillonnées pour effectuer des datations à l'aide d'isotopes cosmogéniques (charbons pour ^{14}C et amalgames de galets riches en quartz le long de profils en profondeur pour ^{10}Be). De retour du terrain, j'ai traité les amalgames de galets du broyage à la mesure de concentration ^{10}Be à Aix-en-Provence, en collaboration avec le CEREGE. J'ai ensuite utilisé le code développé par *Hidy et al. (2010)* pour estimer les âges d'exposition à partir des profils en profondeur de concentration de ^{10}Be . Les datations ^{14}C ont été effectuées par le *Poznan Radiocarbon Laboratory* et *Beta Analytic*.

Les principaux résultats de cette étude suggèrent qu'une part importante de la déformation Holocène est accommodée au niveau du front Himalayen, comme au Népal. En effet, nous déterminons un soulèvement tectonique de l'ordre de $8,7 \pm 2,0 \text{ mm/an}$ au niveau du chevauchement affectant les terrasses alluviales étudiées, ce qui est cohérent avec les résultats obtenus au Népal (*Lavé and Avouac 2000*). Nous montrons également des évidences de ruptures de surface avec des décalages verticaux de plusieurs mètres associés à des séismes majeurs, encore une fois comparables à ceux observées au Népal (eg *Sapkota et al. 2013*). Les datations effectuées indiquent que ces deux événements sont postérieurs à 955 ± 65 et ont donc eu lieu durant le dernier millénaire. Ce chapitre est présenté sous la forme d'un article soumis à *Geology*.

7.2 Active tectonics in eastern Himalaya : new insights from a first morphotectonic study in central Bhutan

Paper submitted to Geology

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Abstract

The Bhutan region is considered as a seismic gap. It corresponds to the only remnant portion of the Himalayan arc where no reliable historical records or surface ruptures of major earthquakes have been reported until now. This can be due either to the presence of the Shillong plateau 150 km southwards of the Himalayan front or to the fact that no active tectonic studies were done so far in this area. In order to characterize active tectonics in eastern Himalaya we carried out a first morphotectonics analysis in south-central Bhutan. We show that the same amount of Holocene deformation is accommodated on the Bhutanese frontal thrust than on the neighboring portions along the Himalayan arc ($\sim 20\text{-}25$ mm/yr), and that two major earthquakes ($M > 8$) broke this thrust in the Bhutan Himalayas during the last millennium. Our results therefore show that Bhutan cannot be considered as a seismic gap. They also challenge the interpretation of one single medieval mega-event that would have broken the Main Frontal Thrust (MFT) from Central Nepal to Eastern Bhutan in the past 1000 years, and the role of the Shillong plateau in the India-Asia convergence.

7.2.1 Introduction and tectonic setting

The 2500 km long Himalayan belt is one of the few collision zones where great earthquakes with sizes comparable to those of oceanic subduction zones can occur. Over the last centuries several major earthquakes ($M > 8$) have been documented in the central and eastern Himalayas : in western Nepal (1505 $M \sim 8.7$), in Bihar-Nepal (1934 $M 8.1$), in Assam (1950 $M \sim 8.4$) and near the Shillong plateau (1897 $M \sim 8.1$) (*Kumar et al. 2010; Lavé et al. 2005; Yule et al. 2006; Mugnier et al. 2013*) (Fig. 7.1). In this sequence of large historical earthquakes the Bhutan Himalayas appear as a ~ 350 km long seismic gap (*Bilham and England 2001*). Only one historical account reports an earthquake in 1713 but its magnitude and accurate location are unknown (*Ambraseys and Jackson 2003*). The occurrence of the 1897 event in the Shillong plateau have been proposed as a mechanism to accommodate some amount of the Himalayan convergence at the longitude of Bhutan, and also as an explanation for the low seismicity currently observed in this area (*Bilham and England 2001; Gahalaut et al. 2011*).

In terms of paleoseismological investigations the study of *Lavé et al. (2005)* along the MFT in Central Nepal suggested the occurrence of a great medieval earthquake around AD 1100 with an estimated vertical slip component of 7-7.5 m. This event would be the last major earthquake to have occurred in this region. More recently, *Kumar et al. (2010)* described large co-seismic displacements with contemporaneous age constraints – and with large uncertainties – both East and West of Bhutan, which suggests that this great medieval surface-rupturing event could have broken the MFT all along the Bhutan Himalayas too.

In central Nepal and Arunachal Pradesh (East of Bhutan) morphotectonic studies have determined Holocene shortening rates on the MFT of 21 ± 1.5 mm/yr (*Lavé and Avouac 2000*) and 23 ± 6.2 mm/yr (*Burgess et al. 2012*), respectively. In Nepal, the shortening rate of $\sim 18-20$ mm/yr measured by GPS (e.g. *Ader et al. 2012*) is consistent with this Holocene shortening rate. This suggests that the interseismic loading accumulated beneath the High Himalayas on the Main Himalayan Thrust (MHT, the down-dip continuation of the MFT and the main interface along which India underthrusts the Himalayas) is relaxed during large events which transfer the shortening to the emergent MFT (*Lavé and Avouac 2000; Cattin and Avouac 2000; Mugnier et al. 2013*). This hypothesis is strengthened by the recent observations of surface ruptures along the MFT associated with great earthquakes

in Nepal (eg *Sapkota et al. 2013*). The current tectonic setting in Bhutan associated to an apparent seismic gap and the presence of the Shillong plateau to the south raises a question of whether the Nepalese model can be used in this part of the Himalayan arc. To answer this issue, we carried out a morphotectonic study in south-central Bhutan that yields to the first estimate of the Holocene slip rate along the Bhutanese frontal thrust and the first insights about the age and magnitude of major seismic events in Bhutan.

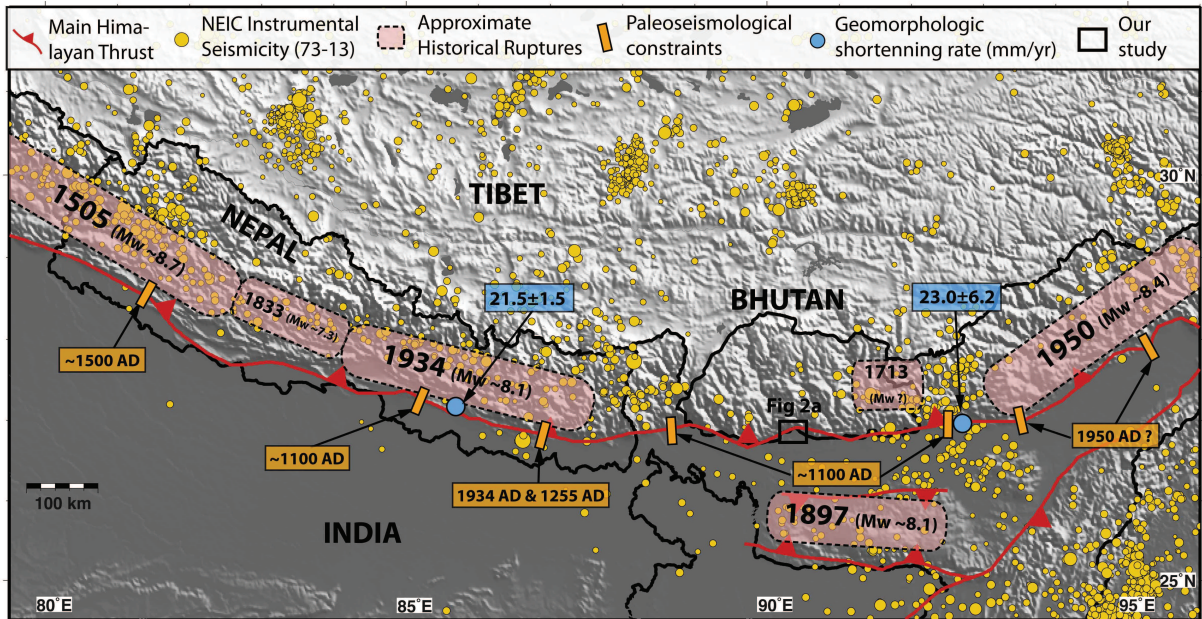


FIGURE 7.1 Seismotectonic map of the Central and Eastern Himalaya. Red rectangles show historical events. Orange rectangles indicate the ages of major surface rupturing events (*Lavé et al. 2005*; *Yule et al. 2006*; *Sapkota et al. 2013*; *Kumar et al. 2010*; *Jayan-gondaperumal et al. 2011*). Blue rectangles give the Holocene shortening rates (*Lavé and Avouac 2000*; *Burgess et al. 2012*). The black frame indicates the location of our study.

7.2.2 Morphotectonic analysis

The E-W trending Bhutan Himalaya corresponds to a ~350 km long segment of the Himalayan arc (Fig. 7.1). We focused our work on the south-central part of Bhutan within the Sarpang re-entrant (Fig. 7.2A) where well preserved morphotectonic markers have been observed from satellite imagery and earlier fieldwork. In this area, the topographic front of the Bhutan Himalayas is 15 km further north compared to the rest of the country where the active front remains roughly linear (*Long et al. 2011*). South of the front, two large southwards flowing rivers (the Sarpang Khola and the Mao Khola), covered the re-entrant with alluvium. A “foreberg” structure (few kilometers wide ridge of hills

rising above the surrounding alluvial fans corresponding to a thrust-related structure; *Bayasgalan et al. (1999)*) affects the alluvial plain 15 km southwards of the topographic front (Fig. 7.2A) attesting of the propagation of the active deformation to the South. This structure extends over 30 km laterally and ends at longitude 90.33°E, uplifting and deflecting the drainage network.

The analysis of a 5 m resolution Digital Elevation Model (built from Pleiades satellite images) and 1/25,000 aerial photographs allowed us to select two sites along the topographic front at the outlets of the Sarpang Khola and the Mao Khola (Fig. ??B). In both areas, sequences of alluvial uplifted terraces are preserved attesting for the accumulation of vertical deformation through time. The first site is located on the western part of the re-entrant within the town of Sarpang; the second one on the eastern part immediately northwards of the town of Gelephu. In the Sarpang area (Fig. 7.2C), abandoned terraces are preserved on both banks of the river. Two main terraces (T1 and T2) can be distinguished by the fact that both are affected to the south by an east-west trending fault scarp perpendicular to the drainage. In the Gelephu area (Fig. 7.2C), we observe the same sequence of fluvial terraces T1 and T2 plus an additional older one T3, preserved on the left bank of the river. Within the right bank of the river, the E-W trending scarp bounds the uplifted terraces T1 and T2. Within the left bank of the river, the scarp is found 1.5 km further North and trends also E-W cutting through the T1 terrace. There we suppose that the two scarps are connected through a NW-SE section within the river bed. The fault trace then turns abruptly N-S and subdivides into a reverse fault scarp uplifting the fluvial terraces deposited on the left bank of the river and a strike slip fault located backwards few hundred meters further East, suggesting that the deformation is partitioned at the surface.

7.2.2.1 Sarpang area

Figure 7.3A shows a perspective view of the left bank of the Sarpang Khola where we focus our morphotectonic analysis. The lower and youngest affected terrace T1 is preserved along the riverbed of the Sarpang Khola (Fig. 7.3A). Figure 7.3B shows the riser of terrace T1 made of cobble and boulder layers topped with a sandy-silty unit. The terrace is affected by a ~4 m high fault scarp perpendicular to the flow of the river. Within the hangingwall, immediately North of the scarp, one observes the slightly folded

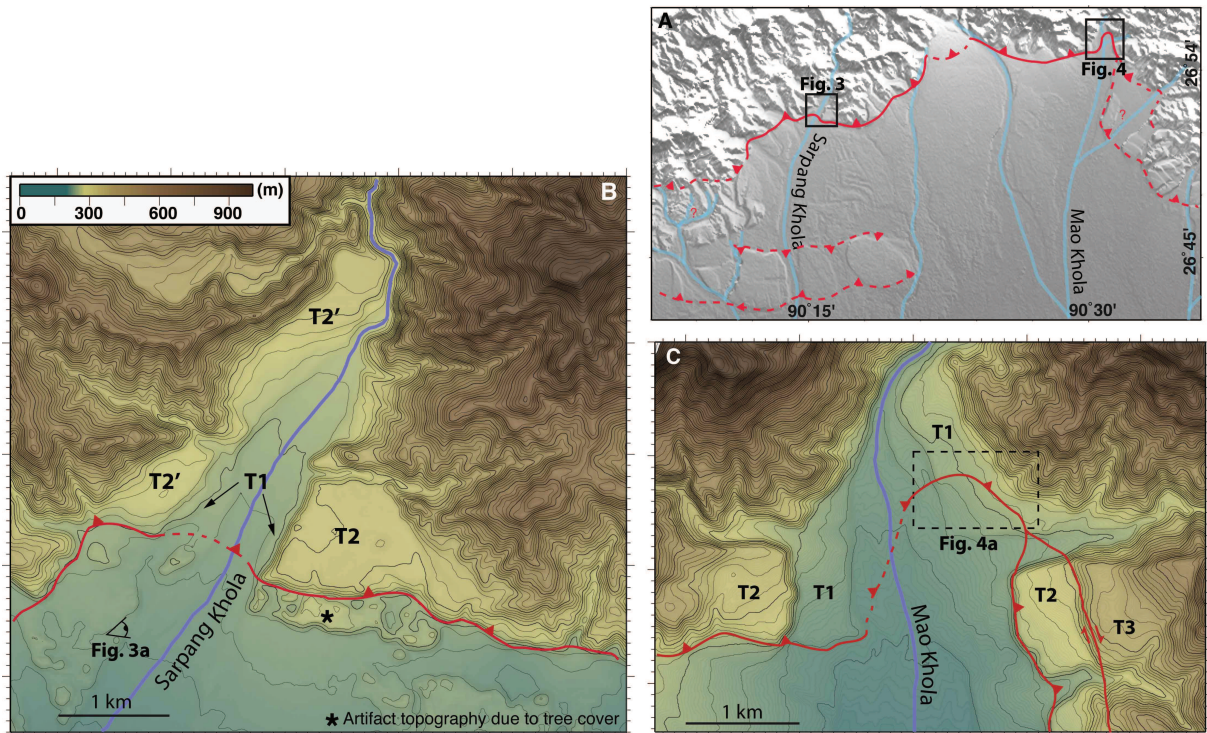


FIGURE 7.2 (A) Shaded SRTM3 topography of the Sarpang re-entrant in central Bhutan with drainage (blue) and active thrust faults (red). (B) and (C) Detailed topographic maps of the Sarpang and Gelephu sites showing uplifted river terraces.

strath horizon of the terrace T1 above a highly fractured bedrock unit (Baxa formation). These features suggest that the Baxa formation overthrusts the Quaternary deposits.

The terrace T2 is limited to the South by an abrupt east-west topographic front perpendicular to the flow of the Sarpang Khola, defining a cumulative fault scarp of several tens of meters that corresponds to the westwards extension of the E-W trending scarp affecting terrace T1. Topographic profiles of T2 across this scarp from a kinematic GPS survey and the 5 m resolution DEM allow estimating a cumulative offset of 53.4 ± 2.2 m (Fig. 7.3C). This corresponds to a minimum tectonic offset as the footwall can be partially buried by colluvium. At the eastern foot of the cumulative scarp we observe a smaller scarp of a height (4.4 ± 0.5 m) on the same order of magnitude than the one affecting terrace T1 (Fig 7.3D).

The abandonment age of the fluvial terrace T2 was determined by in situ-produced cosmogenic ^{10}Be from the measurement of ^{10}Be concentrations along a depth profile (eg *Ritz et al. 1995*). Three stratigraphic units are observed in the pit. The top unit corresponds to a 0.5 m thick sandy unit covering a second sandy unit with pebbles. Several charcoals were collected in this second unit and were dated with radiocarbon. The third

underlying unit corresponds to a thick debris-flow with decimetric cobbles included in a sandy matrix. The analysis of the ^{10}Be concentrations at depth combined with radiocarbon ages allows estimating an age of 6400 ± 1300 yr for the abandonment of the terrace T2 (see details in Sup. Material).

To date the deposition of terrace T1, we collected three charcoals within a 70 cm deep profile dug into the silty-sandy unit found on top of the terrace T1 within the hanging wall. Two sub-units can be distinguished. Two charcoals within the upper sub-unit came modern and were interpreted as belonging to a sandy unit having capped the terraces recently. The third charcoal collected within the lower unit yielded an age of $\text{AD } 1520 \pm 100$ that we interpreted as corresponding to the depositional age of terrace T1 (see details in Sup. Material).

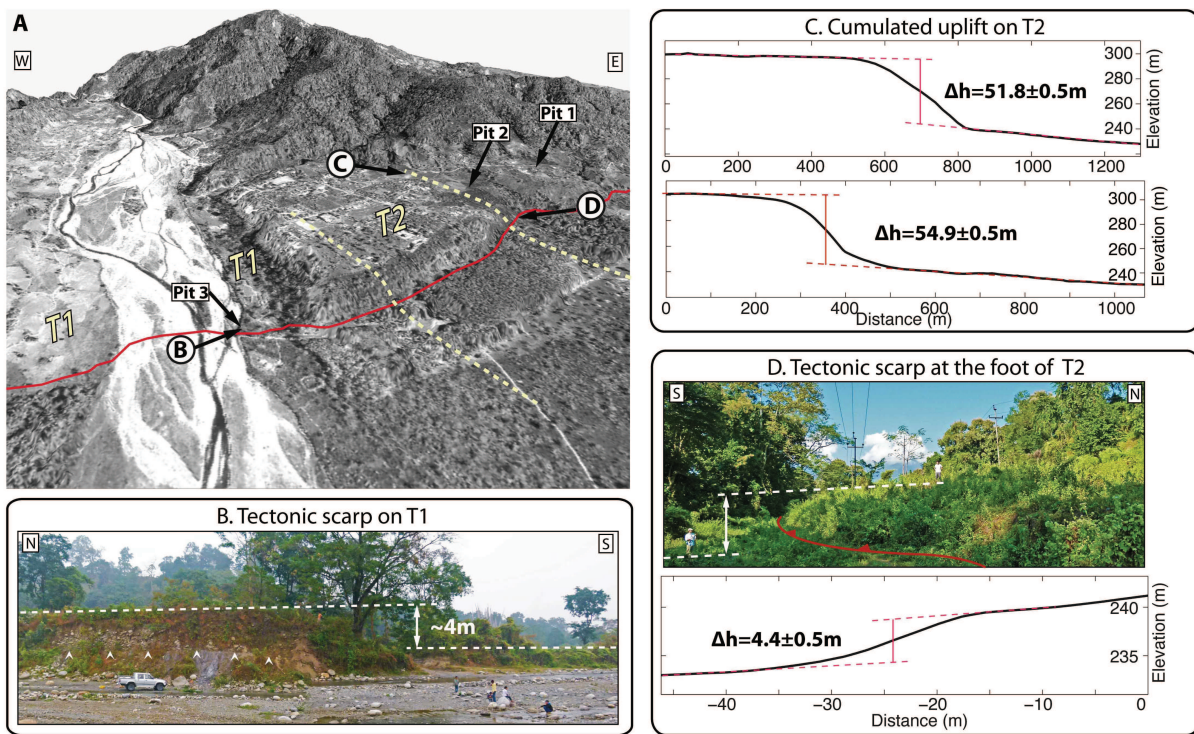


FIGURE 7.3 (A) Perspective view to the NE of the Sarpang Khola left bank (Pleiades image superimposed on a 5 m DEM) and locations of figures B, C, and D. The red and yellow dashed lines define the trace of the thrust fault and the profiles across T2, respectively. (B) Fault scarp affecting T1 terrace riser on Sarpang Khola left bank. (C) Topographic profiles across T2. (D) Photo and topographic profile of a fault scarp at the foot of T2.

7.2.2.2 Gelephu area

Figure 7.4 focuses on our study area on the left bank of the Mao Khola, near Gelephu. At this site, the terrace T1 can be sub-divided into two sub-terraces (T1a and T1b) that are affected by an ENE-WSW trending fault scarp. We survey this area using kinematic GPS to build a DEM of the area (Fig. 7.4A). This allowed us mapping the geometry of terraces and scarps and estimating the terraces offsets. The map shows that the fault scarp is buried under a debris fan where its strike changes from ENE-WSW to a NW-SE. The vertical offset of terrace T1a is 4.6 ± 0.5 m (Fig 7.4B). The vertical offset of terrace T1b is 7.9 ± 0.5 m (Fig 7.4C) across the ENE-WSW scarp, and 7.4 ± 0.5 m (Fig 7.4D) across the NW-SE scarp.

We used radiocarbon and in situ-produced ^{10}Be to date the terrace T1b. The 2 m deep soil pit dug into the terrace shows two main depositional units both composed of a clast-supported layer topped by a sandy-silty horizon. A brown organic-poor modern soil developed in the silty upper part of the first unit. Radiocarbon dating on humic acids from a bulk sample collected in the sands that topped the second unit yields the date of $\text{AD } 955 \pm 65$. This is consistent with the distribution of the ^{10}Be concentration with depth that yields an age of $1,500 \pm 700$ yrs (see details in Sup. Material).

7.2.2.3 Holocene slip rate along the Bhutanese frontal thrust

Combining the mean offset of 53.4 ± 2.2 m across terrace T2 in Sarpang and its ^{10}Be age of $6,400 \pm 1,300$ yrs yields a mean vertical slip rate of 8.7 ± 2.0 mm/yr along the Bhutanese frontal thrust for the Holocene period (Fig. 7.5). Offsets and ages obtained near Sarpang on terrace T1 and near Gelephu on terrace T1b are consistent with this estimate (Fig. 7.5). From the DEM constructed around the Gelephu site we estimate a dip angle of $\sim 20^\circ$ for the thrust. This yields a Holocene shortening rate of 25.4 ± 5.8 mm/yr on the Bhutanese frontal thrust.

7.2.2.4 Magnitude and age constraints on the two last surface rupturing events in Bhutan

Our observations in Sarpang and Gelephu areas show evidences for a ~ 4.5 m high fault scarp affecting the most recent alluvial terrace T1 (see Figs 7.3B, 7.3D and 7.4B).

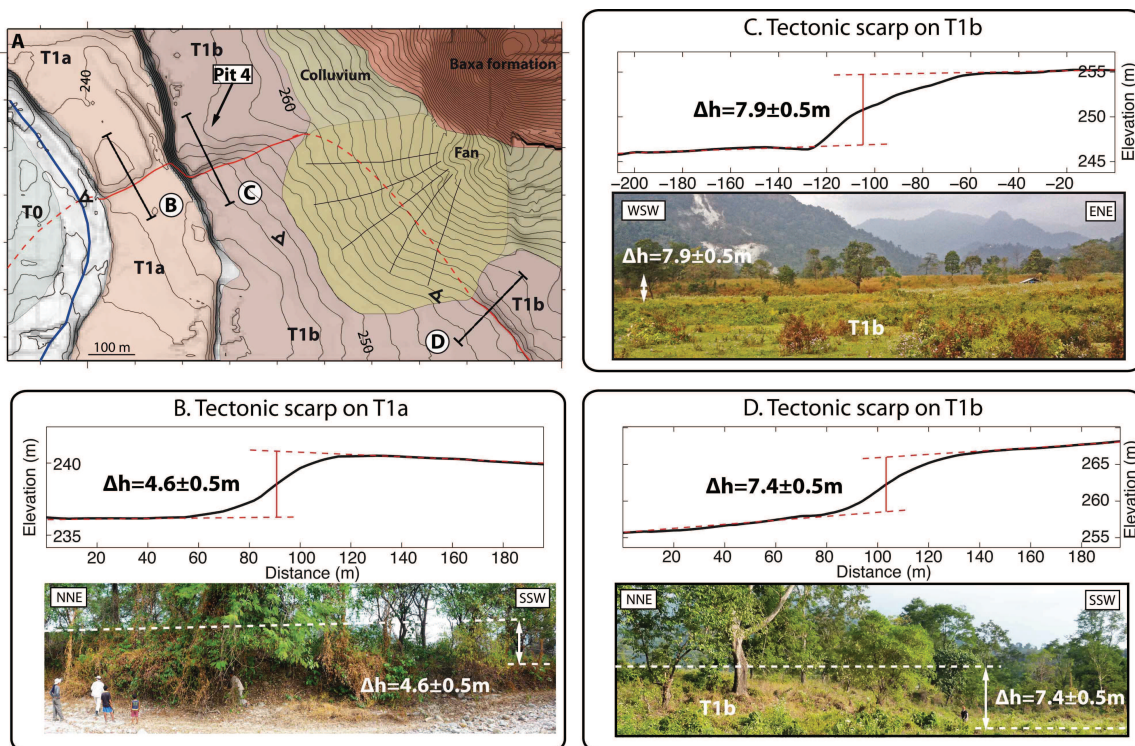


FIGURE 7.4 (A) Map of the T1 terraces sequence observed on the Mao Khola left bank (see location on Fig. 7.2C). The contour lines are every 2 m; the red line corresponds to the fault trace; black lines are topographic profiles. (B) and (C) Profiles and photos of the ENE-WSW trending fault scarp affecting T1a and T1b, respectively. (D) Profile and photo of the NW-SE trending scarp affecting T1b (note the debris fan burying the scarp).

We interpret this offset as the result of the latest surface rupturing event along the Bhutanese frontal thrust. Considering a 20° dip angle for this thrust at surface, this vertical displacement would result from a ~ 13 m slip. At the Gelephu site terrace, T1b is affected by a fault scarp of ~ 8 m that is about twice the height of the scarp affecting T1a (4.6 m). We consider that T1b has been displaced by two events with vertical components of ~ 4.5 m for the youngest and of ~ 3.5 m for the penultimate. This yields a ~ 10 m slip for the penultimate event considering a 20° dip for the fault. These results, comparable with co-seismic displacements associated with the 1934 and 1950 events in the Himalayas (*Jayangondaperumal et al. 2011; Sapkota et al. 2013*) suggest $M > 8$ for the two last surface rupturing events along the Bhutanese frontal thrust.

The radiocarbon age obtained for the deposition of terrace T1 in Sarpang indicates that the first event occurred after $AD\ 1520 \pm 100$. This event could correspond to the AD 1713 event, the only historical earthquake reported for the region (*Ambraseys and Jackson 2003*). Age constraints on T1b in Gelephu indicate that the penultimate event occurred after $AD\ 955 \pm 65$.

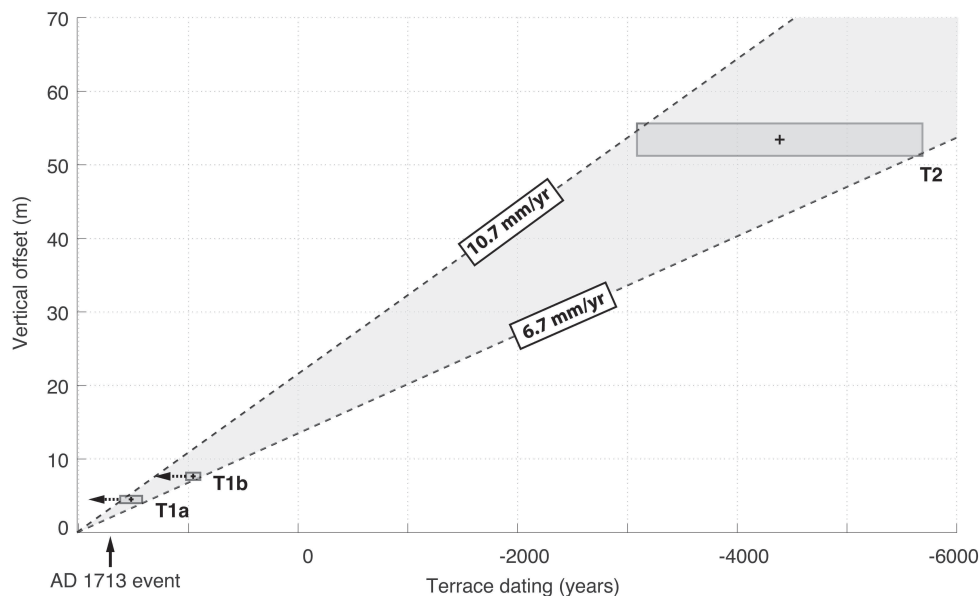


FIGURE 7.5 Uplift rate along the Bhutanese frontal thrust : lengths and widths of the rectangles correspond to the error on ages and elevations of terraces, respectively. Arrows within T1a and T1b rectangles indicate that the seismic event affecting each terrace is younger than their respective depositional age.

7.2.3 Discussion and conclusions

The morphotectonic study across the Bhutanese frontal thrust brings new constraints on the active tectonics in the eastern Himalayas. A Holocene vertical slip rate of 8.7 ± 2.0 mm/yr was estimated. This yields a horizontal shortening rate of 25.4 ± 5.8 mm/yr with an estimated fault dip of 20° at surface. This estimate is consistent with both the 21.5 ± 1.5 mm/yr and the 23.0 ± 6.2 mm/yr shortening rates estimated in Central Nepal (*Lavé et al. 2005*) and in India east of Bhutan (*Burgess et al. 2012*), respectively. This suggests that the Himalayan convergence is mainly accommodated on the MFT. New GPS observations allow a preliminary estimate of ~ 20 mm/yr for the present-day convergence across Bhutan (*Drukpa et al. 2012*), consistent with the slip rate estimated in this study as well.

This estimate of the horizontal slip rate along the Bhutanese frontal thrust suggests that the deformation accommodated within the thrust-related structures located south of Sarpang might be small. *Dasgupta et al. (2013)* estimated an uplift rate of 3mm/yr along this structure that they interpret as being controlled by a south-dipping thrust fault connected at depth to a southwards-propagating décollement located on top of the Indian crystalline basement. We consider that this vertical slip rate corresponds to a maximum estimate since the surface-dated clay horizon is likely not the oldest one trapped behind the structure. Moreover, we disagree with their structural interpretation and consider that the structure does correspond to a classic foreberg controlled by a northwards dipping thrust fault connected with the MHT at depth.

Our observations suggest that two major events with $M > 8$, cumulating ~ 8 m of vertical deformation, occurred during the past ~ 1100 years. The most recent event occurred after AD 1520 ± 100 AD and may correspond to the AD 1713 historical event. If so, the magnitude ($M_w \sim 7$) proposed by *Ambraseys and Jackson (2003)* for the 1713 event might be underestimated. This is not inconsistent with *Ambraseys and Jackson (2003)*'s report itself mentioning that this event 'destroyed all houses in all districts'. The penultimate event occurred between AD 955 and AD 1520. It could correspond to the medieval event of ~ 1100 described by *Lavé et al. (2005)* and *Kumar et al. (2010)*. However, the vertical offset associated with this event (~ 3.5 m) is much lower than the ~ 8 m estimated by *Lavé et al. (2005)*. The occurrence of two surface rupturing events during the past ~ 1100 years is similar to the observations made by *Sapkota et al. (2013)* in eastern Nepal where

two major earthquakes occurred in 1934 and 1255. These results challenge the interpretation of one single medieval mega-event that would have broken the MFT between central Nepal and Bhutan during the past millennium. These new insights suggest that a similar amount of Holocene deformation is accommodated along the MFT in Bhutan than in Nepal, and that the Bhutan Himalayas do not correspond to a seismic gap area. As a result, the deformation associated with the 1897 event in the Shillong plateau may not be related to the Himalayan convergence but rather to a reactivation process of ancient structures within the Indian plate.

Acknowledgments

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7.2.4 Supplementary Material - Dating of Morphotectonic features

In order to date the time of deposition or abandonment of the different terraces, we used radiocarbon as well as in-situ produced cosmogenic ^{10}Be measured along depth profiles. Four pits were excavated; two in T2 (pit P1 and P2), one in T1 (pit P3) at Sarpang and one in T1b (pit P4) at Gelephu. Charcoal samples for radiocarbon dating have been collected in P1 and P2 as well as in a bulk sample of sand from P4 where no visible charcoals could be easily identified in the field. ^{10}Be concentrations have been measured in samples from two depth profiles collected in P1b and P2.

7.2.4.1 Radiocarbon dating

Two samples from pit P2 (B12-P2-C02 and B12-P2-C08) as well as three from pit P3 (B12-P3-C04, B12-P3-C07 and B12-P3-C08) were sent to the Poznan Radiocarbon Laboratory in Poland. All samples were processed with a standard AAA extraction, graphitized and run through an Accelerated Mass Spectrometer (AMS). For more details and

quality checks, see laboratory website at <http://radiocarbon.pl/>. The bulk sand sample collected in P4 (B12-P4-C01) was submitted at a later stage to Beta Analytic. Due to the nature of the sample, humic acids were extracted and dated, thus addressing soil development rather than sediment deposition. All samples were checked for $\delta^{13}\text{C}$ consistency and calibrated using OxCal 3.1 (*Ramsey 1995*) with the IntCal09 calibration curve (*Reimer et al. 2011*) (Table 7.1).

7.2.4.2 In situ ^{10}Be cosmogenic depth profile in Bhutan : sampling and measurements

The concentration measurement of in situ produced cosmogenic ^{10}Be in surficial material allows either dating the initiation of exposure to cosmic ray if denudation is negligible or determining the long-term denudation rate if the studied surface has reached steady-state. In this paper, the depth profile approach has been used (*Braucher et al. 2009*). Cosmogenic nuclides distribution with depth not only allows dating and determining the mean denudation rate that affects the surface since its exposure, but also allows determining the deposition process and the potential inherited concentration prior to deposition. This depth profiling technique is based on the fact that cosmogenic nuclides are produced by at least two types of particles : neutrons in the first meters and muons at depth (*Brown et al. 1995; Braucher et al. 2011*). This approach had been successfully applied to date fluvial terraces (*Anderson et al. 1996; Repka et al. 1997; Hancock et al. 1999*). Because ^{10}Be produced by neutrons in the first meters of Earth reaches steady-state with respect to denudational loss much more rapidly than ^{10}Be produced at depth from reactions with the more penetrating muons particles, it can be used to estimate the denudation rate. ^{10}Be produced at several meters depth by muons can be used to estimate the exposure time. The shape of this attenuation is described by an exponential decrease and can be directly linked to denudation processes. In a context of young alluvial terraces in southern Bhutan this approach is challenging for several reasons. The low elevation of the sampled sites (250-350 m.a.s.l.) as well as their latitudinal situation (26.9° N) correspond to low production rates. All the more so as the supposed exposure ages for the sampled alluvial terraces are very young (a few ka). Another reason comes from the geodynamic and climatic settings characterized by large earthquakes and monsoon that induce high erosion rates and a difficult preservation of geomorphic features. In this study, we used in situ-produced

TABLE 7.1 Calculated dates from ^{14}C dating analysis of the organic material. Calibrated calendar age ranges were calculated using OxCal 3.1 (*Ramsey 1995*) with the IntCal09 calibration curve (*Reimer et al. 2011*).

Sample ID	Depth (m)	Unit	Mass (mg)	Age (yr BP)	$\Delta^{13}\text{C}$	Cal. date (2-sigma)
Pit 2						
P2-C02	0.65	P2, U20	3.333	1210	-26.1	AD 690 – AD 900
P2-C08	0.70	T2, U20	2.232	2535	-27.1	800 BC – 540 BC
Pit 3						
P3-C04	0.24	P3, U10	0.990	Modern	-20.5	
P3-C08	0.35	P3, U10	1.672	Modern	-27.9	
P3-C07	0.50	P3, U20	2.300	415	-29.3	AD 1420 – AD 1620
Pit 4						
P4-C01	0.45	P4, U21	203.2	1100	-21.4	AD 890 – AD 1020

^{10}Be in Si and O in quartz minerals. For each sample, we collected 40 lumps of $\sim 2\text{ cm}^3$ from granite, gneiss and quartzite pebbles. In order to take into account a possibly different erosion/transport history (inheritance) for the different lithologies, the ratio of each type of pebble is the same for all the samples of the same pit. After crushing and sieving, the 0.25-1.0 mm fraction is passed through magnetic separation. All mineral phases but quartz are then eliminated by a series of selective etching in hydrochloric and fluorosilicic acids. The sample is then cleaned in hydrofluoric acid to eliminate potential surface contamination by atmospheric ^{10}Be . After complete dissolution in hydrofluoric acid, the samples were spiked with $\sim 100\ \mu\text{l}$ of an in-house $3.066 \times 10^3\text{ g/g}$ of ^9Be carrier solution. Pure beryllium oxide samples were then obtained after chemical purification by successive anion exchange, cation exchange and alkaline precipitations. Finally, $^{10}\text{Be}/^9\text{Be}$ ratios were measured at the French national AMS (Accelerator Mass Spectrometer) ASTER facility at CEREGE (Aix-en-Provence, France). The AMS sensitivity was determined during each same analytical session by measuring the $^{10}\text{Be}/^9\text{Be}$ ratio of the SRM 4325 NIST reference material, which has an assigned $^{10}\text{Be}/^9\text{Be}$ value of $2.79 \pm 0.03 \times 10^{11}$ (*Nishiizumi et al. 2007*). We finally obtain 7 measurements in P2 and 4 in P4 (Table 7.2).

7.2.4.3 Dating of the T1 terrace at the Sarpang site

We excavated a 70-cm-deep pit in the hanging wall of T1 on the left bank of the Sarpang Khola (See Fig. 7.3D for location). It displayed three main units : a 10-cm-thick brown organic soil overlaying a 35-cm-thick greyish silty sand layer and basal yellowish silt and sand layer (Fig. 7.6). Radiocarbon dating (Fig. 7.6) of two charcoal samples embedded in the intermediate layer yielded modern ages whereas a single sample collected from the basal layer yielded a 2- σ -calibrated date of AD 1520 ± 100 .

7.2.4.4 Dating of the T1b terrace at the Gelephu site (26.93°N 90.52°E 292m)

We excavated a 2 m deep pit in the hanging wall of T1b on the left bank of Moa Khola at Gelephu site. As no visible charcoals could be retrieved from this pit, we collected a bulk sample of sand in U21 (Fig. 7.8). Radiocarbon dating of humic acids yield a 2σ -calibrated date of AD 955 ± 65 . Regarding the cosmogenic ^{10}Be depth profile we follow the approach described by (*Hidy et al. 2010*) where solutions to the concentration equation are generated using a constrained Monte Carlo procedure. In this approach, exposure age, erosion rate,

TABLE 7.2 Measured $^{10}\text{Be}/^9\text{Be}$ ratios and ^{10}Be concentrations used for the modeling of cosmogenic depth profiles in soil Pit 2 and 4. Uncertainties on ^{10}Be concentrations are calculated using the standard error propagation method using the quadratic sum of the relative errors associated to the counting statistics, AMS internal error (0.5% for ASTER), and error associated to procedural blanks (on the order of 3.10^{-15}). Sample thicknesses were $\sim 5\text{cm}$.

Sample ID	Depth (cm)	$^{10}\text{Be}/^9\text{Be}$	1 σ AMS error	^{10}Be concentration (atoms/g)
Pit 2				
P2-Be7	0.25	1.37692×10^{-14}	18.65	$25,792 \pm 4810$
P2-Be6	0.50	1.10049×10^{-14}	16.05	$27,568 \pm 4423$
P2-Be5	0.75	2.75228×10^{-14}	17.45	$21,333 \pm 3723$
P2-Be4	1.00	1.97171×10^{-14}	12.86	$15,679 \pm 2017$
P2-Be3	1.50	1.20700×10^{-14}	18.05	$9,726 \pm 1755$
P2-Be2	2.00	3.01754×10^{-14}	20.45	$23,449 \pm 4795$
P2-Be1	2.25	7.96204×10^{-15}	31.49	$4,745 \pm 1494$
Pit 4				
P4-Be7	0.00	2.10863×10^{-14}	20.76	$17,394 \pm 3611$
P4-Be6	0.25	1.70218×10^{-14}	24.29	$12,537 \pm 3045$
P4-Be4	0.75	1.70678×10^{-14}	29.63	$11,257 \pm 3335$
P4-Be3	1.00	1.68094×10^{-14}	24.62	$11,109 \pm 2735$

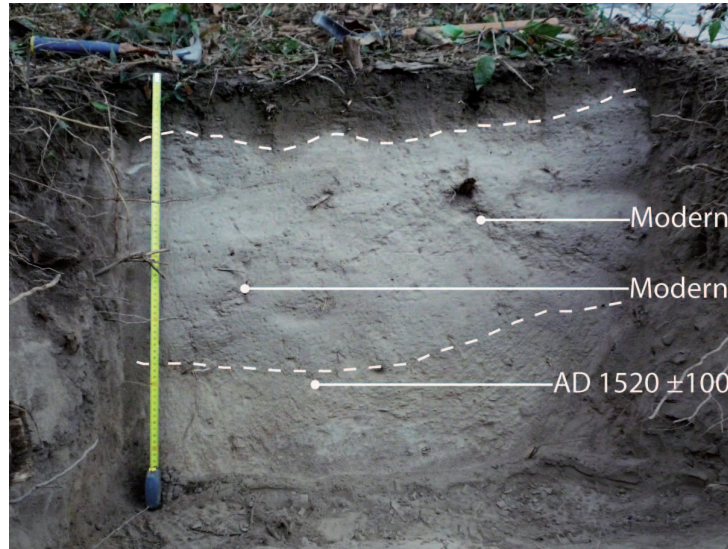


FIGURE 7.6 Photograph of the 70 cm deep soil pit excavated in T1 at the Sarpang site. White dashed lines show the limits of the 3 units. On top, the first unit is a 10-cm-thick brown organic soil. This soil is overlaid by a 35-cm-thick greyish silty sand layer. The deepest visible unit, below ~ 45 cm, is a yellowish silt and sand layer. White dots locates the charcoals with their associated datations.

inheritance and density are sampled from a user-assigned parameter range to produce a solution. The obtained solution and the data are then compared to compute a reduced χ_2 statistic. During the simulation, parameters are stored as a possible solution for a 1σ confidence window. We use a ^{10}Be spallation production rate of $4.49 \pm 0.39 \text{ atoms.g}^{-1}.\text{y}^{-1}$ (*Stone 2000; Nishiizumi et al. 2007*) which is scaled to the sample site using (*Stone 2000*) scaling scheme. The muon production is calculated after (*Balco et al. 2008*). Surrounding topography is accounted for but has a very weak impact on surface production rates. All the measured concentrations are taken into account and density is treated as a free parameter within the range of $1.8 - 2.3 \text{ g.cm}^3$. The shape of the upper few meters of a depth profile is very sensitive to exposure age and erosion rate. If erosion rate is treated as a completely free parameter, the erosion rate cutoff needed to resolve a finite age is 0.35 mm/y . In the Himalayas, this order of magnitude for erosion rates correspond to denudation rates integrated over drainage basins that average several erosional processes on a wider scale (*Godard et al. 2013*). The erosion rate determined from the depth profile is significantly smaller than this value as it corresponds to a local lowering of the terrace surface mainly controlled by deflation. To be conservative, we use an erosion rate within the range $0 - 0.035 \text{ mm/y}$. The maximum erosion rate used in our simulations corresponds to the highest erosion rate found by (*Siame et al. 2012*) for alluvial terraces in Taiwan, a

reasonable climate analog. We generate profile solutions within a 1σ confidence window up to a total of 100,000 solutions. Results are summarized in Figure 7.8-B where distribution of profile solutions as well as probability density functions for exposure age and density are shown. The exposure age yields a 2σ -value of $1,500 \pm 700$ yr. This ^{10}Be exposure age corresponds to date of $\text{AD } 563 \pm 650$. This is consistent with the radiocarbon dating of the bulk sample of sand in Unit 21 that yields $\text{AD } 955 \pm 65$. Considering the more favorable error bars produced by ^{14}C , we will retain $\text{AD } 955 \pm 65$ as the age constraint for T1b.

7.2.4.5 Dating of the T2 terrace at the Sarpang site (26.86°N 90.27°E 351m)

We excavated a 2.3-m-deep pit in the hanging wall of T2 at the Sarpang site. The two charcoal samples retrieved from Unit 20 (overlying the boulder and pebble unit) are dated at 670 ± 130 BC and $\text{AD } 795 \pm 105$. Considering the oldest dated charcoals as reworked, this yields a minimum deposition age of 1218 ± 105 yr for this unit. ^{10}Be concentration clearly decreases with depth except for the samples at 0.5 m and 2 m depth (Fig. 7.7A). We consider the 2 m sample as an outlier as such a high concentration of ^{10}Be cosmionuclides cannot be explained compared to the other samples. Similar concentrations from samples 0.25 m and 0.5 m from Unit 20 combined with the age constraint obtained from ^{14}C dating suggest reworking of the first 0.5 m of the profile. Hence Unit 20 would be composed of pebbles exposed at the surface since the emplacement of the terrace and reworked ~ 1200 years ago by a debris fan (in which Pit T1 was excavated and visible ~ 300 m north of the pit but erased by farming at the sampling site itself). We thus assume here that the top of the original terrace was 50 cm lower than today. And was covered 1200 years ago by the debris fan. To account for this, the ^{10}Be concentrations accumulated during 1200 years by the samples at their sampling depths have been subtracted to the measured concentrations and 50 cm have been subtracted to all samples depths but P2-Be7 (this means that the current 0.5 m deep sample was at the surface before being buried). Then the resulting ^{10}Be distribution with depth was modeled following the same approach as describe in the previous paragraph. Figure 7.7B show the solution space. The five samples simulation yields a mean 2σ -age of 5200 ± 1200 yr corresponding to a final exposure age of 6400 ± 1300 yr.

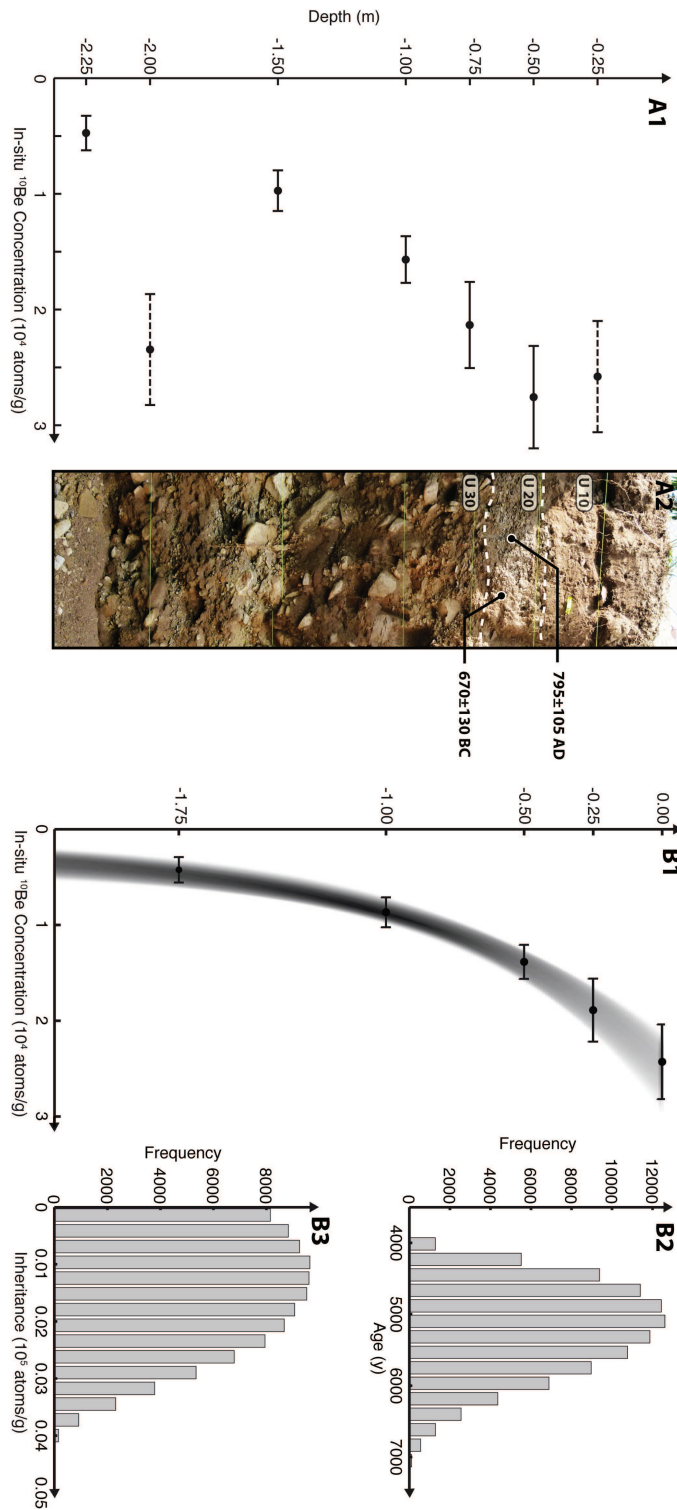


FIGURE 7.7 ^{10}Be Cosmogenic depth profile in P2 and exposure age of T2 at the Sarpang site. A1) In-situ ^{10}Be concentration of the 7 samples collected in P2 and their associated error bars. 0.25 and 2.00 m deep samples are dashed because those measurements are not used in the analysis. A2) Photograph of the 2.25 m deep soil pit and limits of the 3 main units. U10 : light brown silty layer covered by soil. U20 : Clast supported layer (angular, 0.5-2 cm in diameter), fine sand-silty matrix. U30 : debris-flow made of cobbles and pebbles (5-50 cm in diameter) included in a sandy matrix. Dated charcoals in unit 20 are located by black dots. B1) Depth profile showing used in-situ produced cosmogenic ^{10}Be concentrations together with the results of the modeling for the exposure duration of T2. (B2 and B3) Cumulative plots of the number of profiles generated by the Monte Carlo inversion for exposure age and inheritance, respectively.

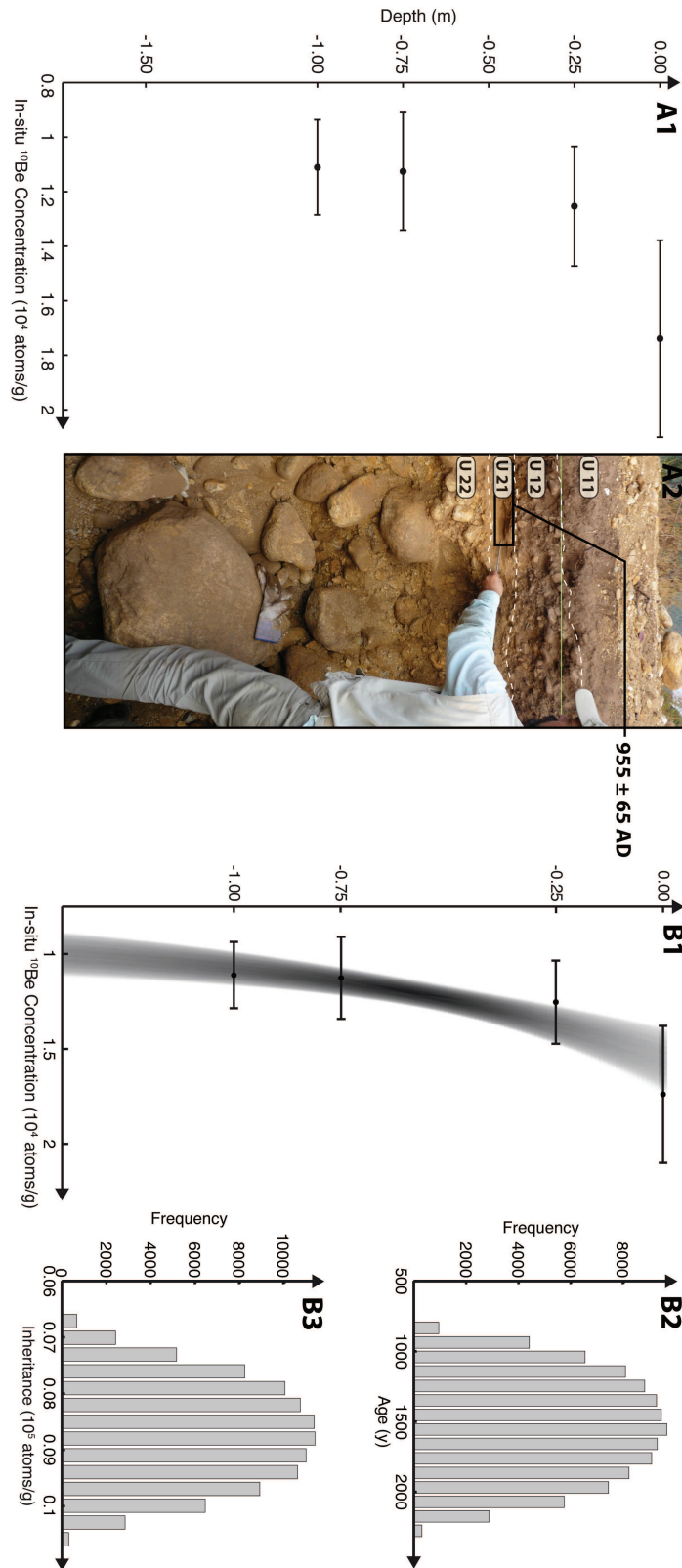


FIGURE 7.8 ^{10}Be Cosmogenic depth profile in P4 and exposure age of T1b at the Gelephu site. A1) In-situ ^{10}Be concentration of the 4 samples collected in P4 and their associated error bars. A2) Photograph of the ~2m deep soil pit and limits of the 4 main units U11 : light brown soil, 30 cm thick, silt. U12 light-brown pebbly unit (1-5 cm in diameter), sandy matrix. U21 : orangish sand. U22 : debris-flow made of cobbles (5-30 cm in diameter) included in a sandy matrix. B1) Depth profile showing used in-situ produced cosmogenic ^{10}Be concentrations together with the results of the modeling for the exposure duration of T1b. (B2 and B3) Cumulative plots of the number of profiles generated by the Monte Carlo inversion for exposure age and inheritance, respectively.

TABLE 7.3 Statistics for the simulation of the P2 and P4 profiles. Ages, inheritance, erosion rate and density are reported for the median, mean and lowest χ^2 for the 100,000 simulations.

	Age (yr)	Inheritance (atoms/g)	Erosion rate (mm/yr)	Density
Pit 2				
Median	5200	1500	0.0184	2.08
Mean	5200	1600	0.0181	2.07
Lowest χ^2	5600	500	0.0331	1.88
Pit 4				
Median	1500	8700	0.0178	2.06
Mean	1500	8700	0.0177	2.05
Lowest χ^2	1400	9300	0.0301	2.24

CHAPITRE 8

CONCLUSION ET PERSPECTIVES

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8.1 Apports de cette étude

Au cours de ce travail de thèse, j'ai effectué l'acquisition, le traitement et l'interprétation de nouvelles données gravimétriques en Himalaya. Ces données m'ont permis de contraindre des modèles thermo-mécaniques de flexure lithosphérique de la plaque Indienne le long de profils perpendiculaires à la chaîne couvrant une zone large de près de 1000 km. Des modèles numériques ont aussi été développés pour étudier les variations latérales de l'état de contrainte du MHT sur le dernier millénaire. De plus, j'ai étudié les déformations récentes au niveau du chevauchement frontal Bhoutanais. Pour ce faire, j'ai cartographié et échantillonné des terrasses alluviales affectées par ce chevauchement. Afin de dater la mise en place de ces marqueurs, j'ai effectué le traitement, l'analyse et la modélisation des données cosmogéniques (^{10}Be) issues de l'échantillonnage. Au cours de ces trois dernières années, j'ai donc combiné diverses disciplines afin d'apporter de nouvelles contraintes à l'étude des variations latérales des déformations crustales en Himalaya et notamment au Bhoutan, une zone peu étudiée jusqu'alors. Les principaux résultats sont synthétisés dans les paragraphes qui suivent.

Nouvelles données gravimétriques en Himalaya :

Au cours des 4 campagnes gravimétriques effectuées, nous avons établi 366 nouveaux points gravimétriques au Népal et au Bhoutan. Le réseau gravimétrique Népalais a été complété à l'Est de 85°E avec un profil Est/Ouest jusqu'à 80°E et deux profils Nord/Sud à 80,5°E et 83,5°E. Nous avons aussi établi le premier réseau gravimétrique Bhoutanais en effectuant des mesures le long de la plupart des routes accessibles du pays. Il est constitué d'un profil Est/Ouest à 27,5 °N et de trois profils Nord/Sud à 89,5°E, 90,5°E et 91,5°E. Reste à effectuer une (ou des) mesure(s) absolue(s) au Bhoutan afin de caler le réseau Bhoutanais et compléter le profil Nord/Sud entamé à 82°E au Népal.

Variation latérale de la structure lithosphérique :

Ces nouvelles données gravimétriques m'ont permis de contraindre des modèles thermo-mécaniques simulant la flexure de la plaque Indienne. Ces modèles suggèrent une homogénéité de la structure lithosphérique et de la rhéologie entre l'Est et le Centre Népal. Les données gravimétriques au Bhoutan montrent un signal différent qui témoigne d'une variation latérale du comportement flexural de la plaque Indienne. D'après nos modèles, cette différence ne peut pas être expliquée seulement par le chargement différentiel induit

par les variations topographiques observées entre le Népal et le Bhoutan. Les modèles thermo-mécaniques effectués montrent qu'une diminution de la résistance mécanique du manteau lithosphérique Indien permet de diminuer la rigidité flexurale de la plaque Indienne et d'expliquer les données gravimétriques au Bhoutan.

Si ces nouvelles données m'ont permis de mettre en évidence l'existence d'une variation latérale des structures lithosphériques entre le Népal et le Bhoutan, je n'ai pas caractérisé la nature de cette transition. Cette caractérisation est d'autant plus importante qu'une transition localisée aura un impact fort sur la segmentation de l'arc Himalayen et donc probablement sur l'extension latérale des ruptures associées aux séismes majeurs.

Variations latérales de l'état de contrainte du MHT :

Un modèle de dislocations dans un demi-espace élastique a été utilisé pour quantifier les variations spatiales et temporelles de l'état de contrainte du MHT sur le dernier millénaire. Nous avons montré que, même dans l'hypothèse où un méga-séisme ait rompu ~ 1000 km de l'arc Himalayen en ~ 1100 , la variation de contrainte associée à cette rupture a été compensée par le chargement séculaire accumulé depuis. Nous avons aussi montré que le séisme du Shillong de 1897 n'a que très peu d'effet sur l'état de contrainte actuel du MHT. Enfin, notre modèle suggère que la distribution de sismicité actuelle peut être expliquée à la distribution spatiale des séismes majeurs passés. Les variations de couplage intersismique ou de taux de convergence pouvant accroître encore d'avantage les hétérogénéités de contraintes sur le MHT.

Cependant, la modélisation effectuée est tributaire des incertitudes sur les surfaces de ruptures et les glissements moyens associés aux séismes. De plus la prise en compte d'une géométrie plus réaliste du MHT avec la présence de la rampe sous le Haut-Himalaya et d'éventuelles rampes latérales pourrait induire des variations latérales de l'état de contrainte sur le MHT non prises en compte dans nos modèles.

Déformations récentes au niveau du chevauchement frontal Bhoutanais :

Grâce à la cartographie et à la datation des terrasses alluviales surélevées au niveau chevauchement frontal Bhoutanais, nous avons montré que, comme au Népal, la majorité de la déformation Holocène semble être située au front de la chaîne. Nous avons aussi mis en évidence l'existence de deux ruptures de surface avec des décalages verticaux pluri-métriques associés à deux séismes majeurs sur le dernier millénaire. Si le manque

de données GPS au centre et à l'est Bhoutan ne permet pas, pour l'instant, d'estimer le couplage intersismique dans cette zone, ces évidences de séismes majeurs suggèrent que l'hypothèse d'une zone totalement découplée (utilisée dans le chapitre 6) est peu probable.

Les taux de surrection Holocène calculés sont robustes, cependant de meilleures contraintes sur le pendage de la faille sont nécessaires pour une meilleure estimation des taux de raccourcissement associés à ce chevauchement. L'obtention de contraintes d'âges plus précises pour ces deux événements est nécessaire dans le but de pouvoir effectuer des corrélations spatio-temporelles avec les autres ruptures de surfaces observées en Himalaya.

Outre l'apport intrinsèque des nouvelles contraintes que je viens de synthétiser pour l'étude des variations latérales de la déformation crustale en Himalaya, ce travail de thèse et notamment les premières données gravimétriques et sismo-tectoniques acquises au Bhoutan ont permis de poser les bases des projets *Bhutanepal* et GANSSER¹, respectivement financées par *l'Agence Nationale de la Recherche* et les *Fonds National Suisse de la Recherche Scientifique*. Dans la suite de ce chapitre, je présente deux perspectives qui font suite à cette thèse : le premier paragraphe concerne les différentes études à mener au Bhoutan pour une meilleure compréhension des mécanismes de déformation récente à l'échelle du prisme. Dans la seconde, je présente les travaux en cours sur les données gravimétriques satellitaires GOCE.

8.2 Déformations récentes dans l'Himalaya du Bhoutan

Suite à l'étude présentée dans le chapitre 7, il est nécessaire d'améliorer les contraintes d'âges et de quantité de glissement associées aux ruptures de surface observées. Pour ce faire, le nettoyage des affleurements en bord de rivière ou l'ouverture de tranchées permettrait une estimation plus robuste du glissement en surface ainsi qu'un échantillonnage ¹⁴C plus complet permettant de contraindre l'âge des séismes avec une meilleure précision. La diminution des barres d'erreurs sur les contraintes d'âges associées aux séismes majeurs est d'autant plus importante qu'elle permet de limiter les différentes interprétations sur l'extension spatiale des séismes passés en Himalaya. Le pendage de surface utilisé pour convertir les taux de surrection en taux de raccourcissement a été estimé à partir de la

1. <http://www.seismo.ethz.ch/research/groups/alrt/projects/gansser/index>

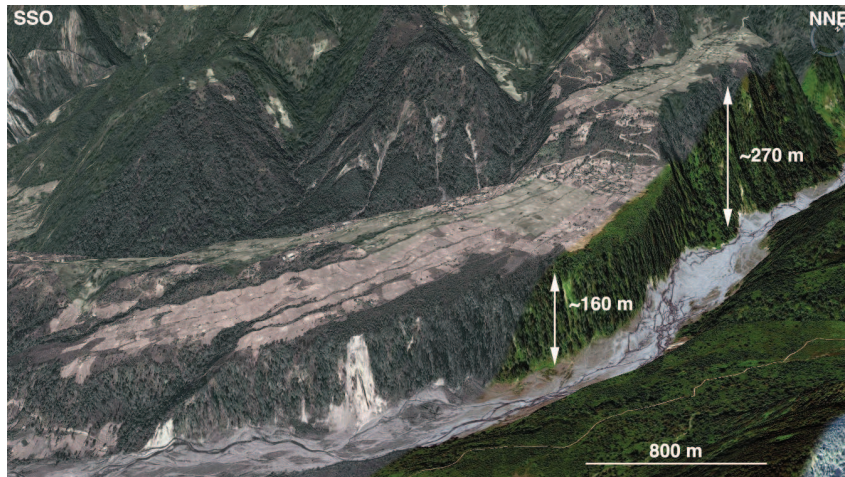


FIGURE 8.1 Terrasses surélevées situées à moins de 10 km au nord du chevauchement frontal Bhoutanais, au N/O de Samdrup Jongkhar dans l'est Bhoutan ($26,91^{\circ}\text{N}/91,75^{\circ}\text{E}$).

trace de la faille frontale. Des expériences de géophysiques de sub-surface (méthodes électriques et/ou sismiques) pourraient permettre une meilleure estimation de son pendage et donc un taux de raccourcissement plus robuste.

Afin d'étudier la variabilité latérale du taux de raccourcissement au niveau du front Himalayen au Bhoutan (et l'extension des ruptures des séismes majeurs), il est nécessaire d'étendre notre zone d'étude qui pour l'instant est concentrée au centre du Sud-Bhoutan. Nous avons déjà repéré des sites potentiels avec des terrasses alluviales surélevées notamment à l'Est Bhoutan dans la région de Samdrup-Jongkhar. Dans cette zone, deux terrasses situées à moins de 10 km du chevauchement frontal Bhoutanais sont surélevées d'environ 160 m et 270 m par rapport au niveau actuel de la lit de la rivière (Fig. 8.2).

Au cours des 9000 km parcourus sur les routes Bhoutanaises, nous avons observé de nombreuses évidences de terrasses alluviales surélevées à l'intérieur de la chaîne (Fig. 8.2). L'étude systématique de ces terrasses le long de profils perpendiculaires à la chaîne permettrait d'établir des profils de la déformation Holocène (*Lavé and Avouac 2001*). Dans le même temps, l'échantillonnage des sables de rivières et l'utilisation des cosmocléides permettrait de déterminer les taux d'érosion le long et au travers de l'Himalaya du Bhoutan (*Godard et al. 2013*).

Les re-mesures du réseau temporaire GPS mis en place en 2012-2013 dans le centre et l'est du Bhoutan permettront de déterminer le chargement séculaire et le couplage intersismique dans cette zone. Si la zone bloquée en intersismique au Népal est large d'environ 100 km, aucune contrainte ne permet pour l'instant de connaître la segmentation du



FIGURE 8.2 Terrasses surélevées situées dans la région de Trashingang à plus de 50 km au nord du chevauchement frontal Bhoutanais ($27,28^{\circ}\text{N}/91,45^{\circ}\text{E}$).

MHT au Bhoutan. Les données InSAR pourrait être utilisées pour étudier la déformation intersismique verticale qui peut servir de proxy à la localisation de la zone de transition entre la zone couplée et le glissement stable (*Grandin et al. 2012*).

Le réseau temporaire de stations sismologiques déployées cette année au Bhoutan dans le cadre du projet GANSSER permettra une meilleure localisation de la sismicité et peut-être la mise en évidence d'une concentration de micro-sismicité associée à la zone de transition du MHT (*Pandey and Agrawal 1999*). De plus ce réseau permettra l'imagerie des structures crustales et devrait nous donner de nouvelles contraintes sur le géométrie du MHT et du Moho au Bhoutan.

8.3 Vers l'utilisation des données gravimétriques satellitaires GOCE

Si les données gravimétriques terrestres nous ont permis de montrer l'existence d'une variation latérale du comportement flexurale de la plaque Indienne, la nature de cette transition n'a pas été déterminée. En terme de segmentation de l'arc et de longueur de rupture potentielle des séismes majeurs, une transition localisée aura un effet beaucoup plus important qu'une transition grande longueur d'onde liée à une flexure latérale. Répondre à cette question nécessite l'utilisation d'une approche en 3 dimensions. Afin de contraindre au mieux la géométrie de cette transition, nous avons débuté une collaboration avec *I. Panet* et *G. Pajot-Métivier* de l'*Institut National de l'Information Géographique et Forestière (IGN)*. L'approche consiste à utiliser, en plus des données terrestres, les données satellitaires GOCE. Les mesures de gradiométrie effectuées par GOCE (à ~ 265 km d'altitude) permettent d'accéder aux variations du champ de pesanteur selon les 3 composantes de l'espace. Ces gradients gravimétriques sont particulièrement sensibles à la géométrie des sources. Cependant l'utilisation du signal GOCE pour des problématiques régionales comme la notre reste peu développée.

La figure 8.3 présente les données de gradients issues de GOCE selon les composantes NN, EE et ZZ. Ces données constituent les éléments diagonaux du tenseur des dérivées secondes du potentiel perturbateur auxquels ont été soustrait les gradients de l'ellipsoïde de référence usuel. En d'autres termes, ces données correspondent à une analogie en gradient des anomalies à l'air libre mesurées à ~ 265 km d'altitude. Afin de voir si la variation latérale mise en évidence aux chapitres 3 et 4 est visible dans les gradients GOCE, nous avons calculé les variations latérales de ces gradients. Nos calculs suggèrent une variation latérale entre 88°E et 89°E (Fig. 8.4).

Travaux en cours

Afin de modéliser ce signal, nous sommes en train de mettre en place un modèle thermo-mécanique en 3 dimensions de la flexure de la plaque Indienne (Fig. 8.5). Pour calculer les gradients gravimétriques associés à ce modèle, nous utilisons les formulations de *Singh and Gupthasarma (2001)* qui permettent de calculer l'effet d'un polyèdre (correspondant à un élément dans le modèle thermo-mécanique) selon les 3 composantes de l'espace.

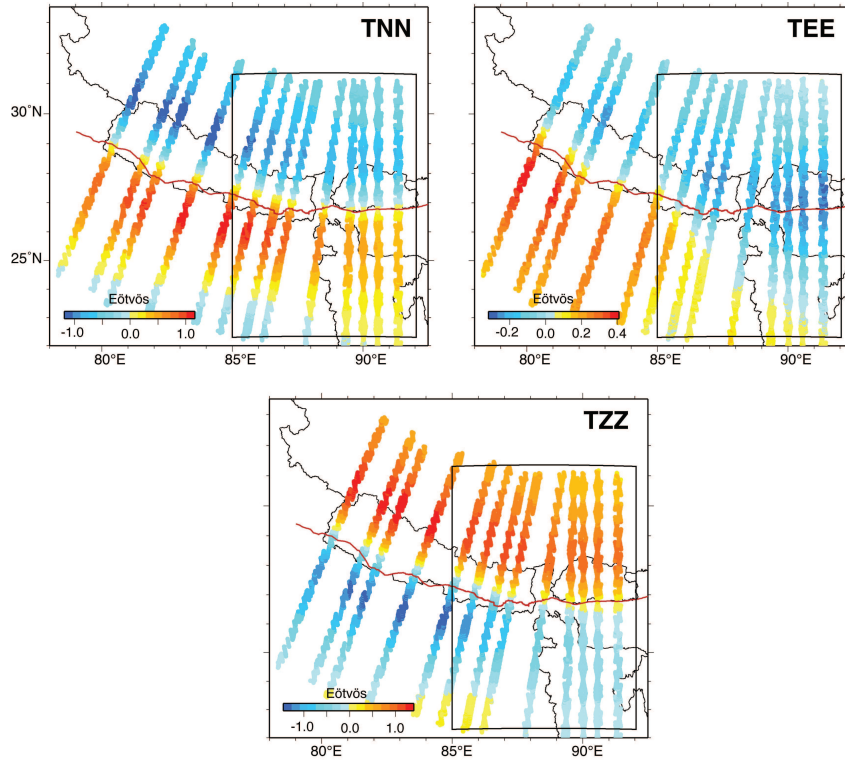


FIGURE 8.3 Gradients GOCE selon les composantes Nord/Nord (TNN), Est/Est (TEE) et profondeur/profondeur (TZZ).

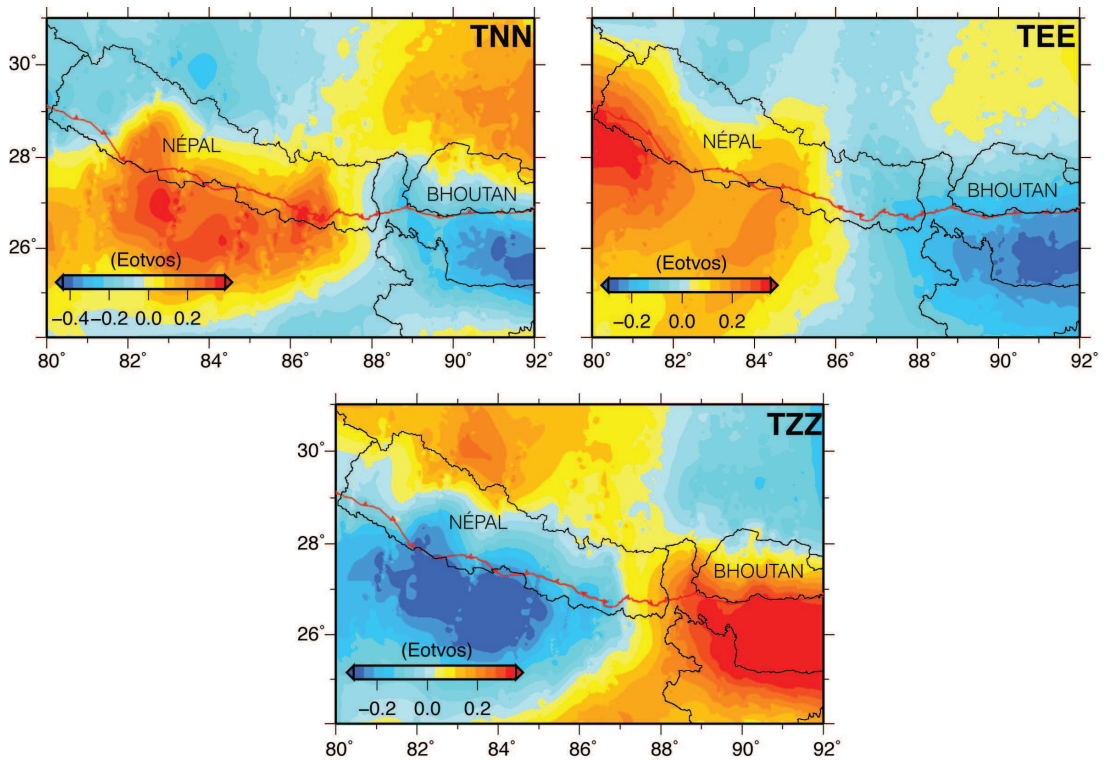


FIGURE 8.4 Cartes des variations latérales des gradients gravimétriques GOCE selon les composantes TNN, TEE et TZZ. Pour le détail du calcul, se reporter au paragraphe 5.2.2.

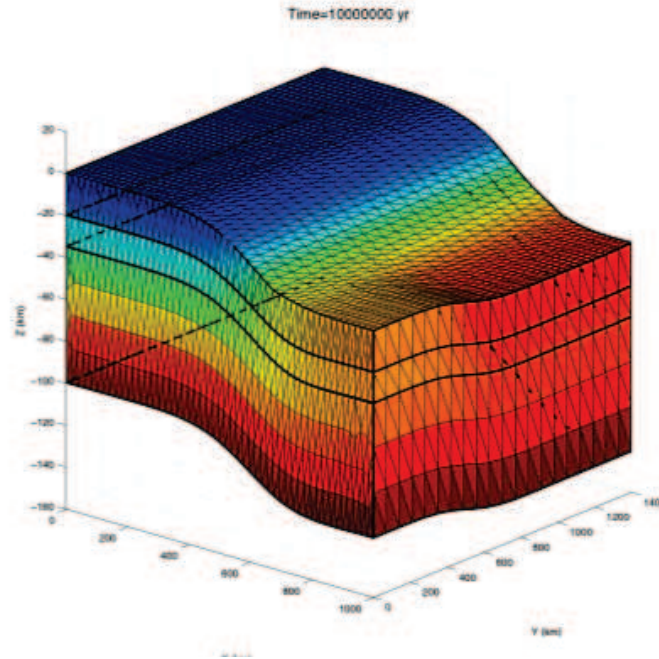


FIGURE 8.5 Géométrie issue d'un modèle thermo-mécanique 3D préliminaire de flexure de la plaque Indienne

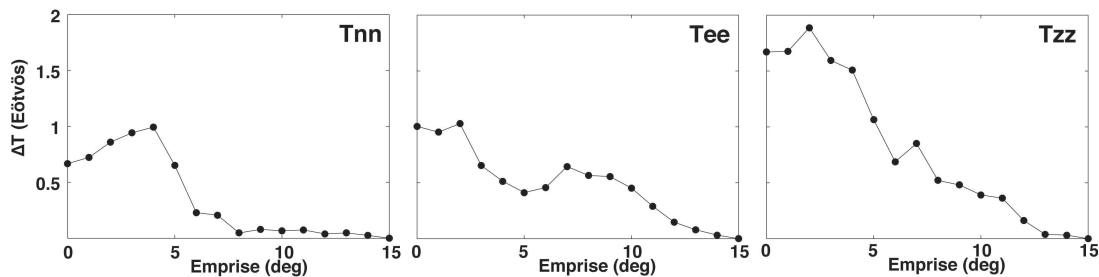


FIGURE 8.6 Effet de l'emprise sur les corrections de terrain calculées par la méthode des prismes selon les composantes NN, EE et ZZ.

Pour modéliser le signal complet de GOCE, la topographie doit également être prise en compte. Nous avons développé un code de calcul permettant d'estimer l'effet topographique à partir de prismes provenant du maillage d'un modèle numérique de terrain. Des tests sur l'emprise à prendre en compte ont été effectués (Fig. 8.6). Afin de valider ces calculs, nous sommes en train de comparer les résultats obtenus par cette méthode (Fig. 8.7A) et par des développements en harmoniques sphériques obtenus à l'IGN (Fig. 8.7B). La comparaison des résultats préliminaires montre une bonne cohérence de la forme des signaux cependant leurs valeurs diffèrent. Plusieurs pistes sont actuellement explorées :

- Vérification de la prise en compte correcte de l'épaisseur d'eau dans la méthode des prismes.

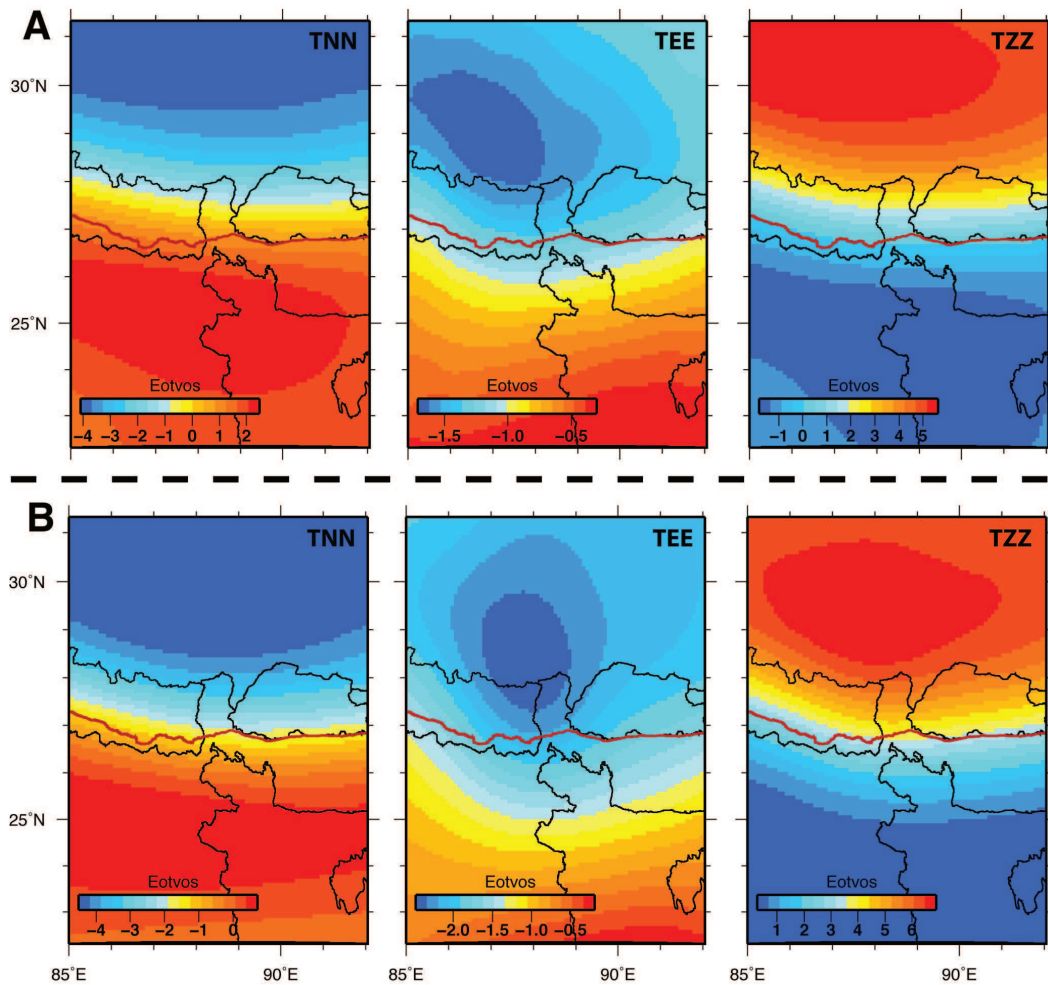


FIGURE 8.7 (A) Corrections de terrain selon les composantes NN, EE et ZZ avec la méthode des prismes pour une emprise de 14° . (B) Corrections de terrain selon les composantes NN, EE et ZZ issues de la méthode en harmoniques sphériques.

- Emprise et résolution dans la méthode des prismes (résolution de 25 km pour l'instant).
- Vérification des contrastes de densité utilisés dans le calcul en harmoniques sphériques.
- Vérification des projections sphère-ellipsoïde.

Travaux futurs

Ce travail est prometteur quant à l'utilisation des données GOCE pour étudier la nature de la transition observée entre le Népal et le Bhoutan et plus généralement la structure lithosphérique. Les trois points cités ci-dessous constituent les prochaines étapes de ce travail :

- Comprendre l'origine de la différence obtenue sur les calculs des corrections de ter-

rain par la méthode des prismes et celle en harmoniques sphériques (Fig. 8.7).

- Effectuer des tests de sensibilité à partir de modèles directs pour étudier leur impact sur le signal GOCE.
- Prendre en compte les variations du signal à plus grande échelle (sensibilité de GOCE aux structures mantelliques profondes).

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ANNEXE A

ANNEXES

A.1 Résumé des communications scientifiques

A.1.1 EGU 2012

Lateral variations of gravity anomalies in the Eastern Himalayas : preliminary results and interpretations (Poster)

THÉO BERTHET¹, GYÖRGY HETÉNYI², RODOLPHE CATTIN¹, CÉDRIC CHAMPOLLION¹, ERIK DOERFLINGER¹, SOMA NATH SAPKOTA³, DOWCHU DRUKPA⁴, JAMYANG CHOPHEL⁴

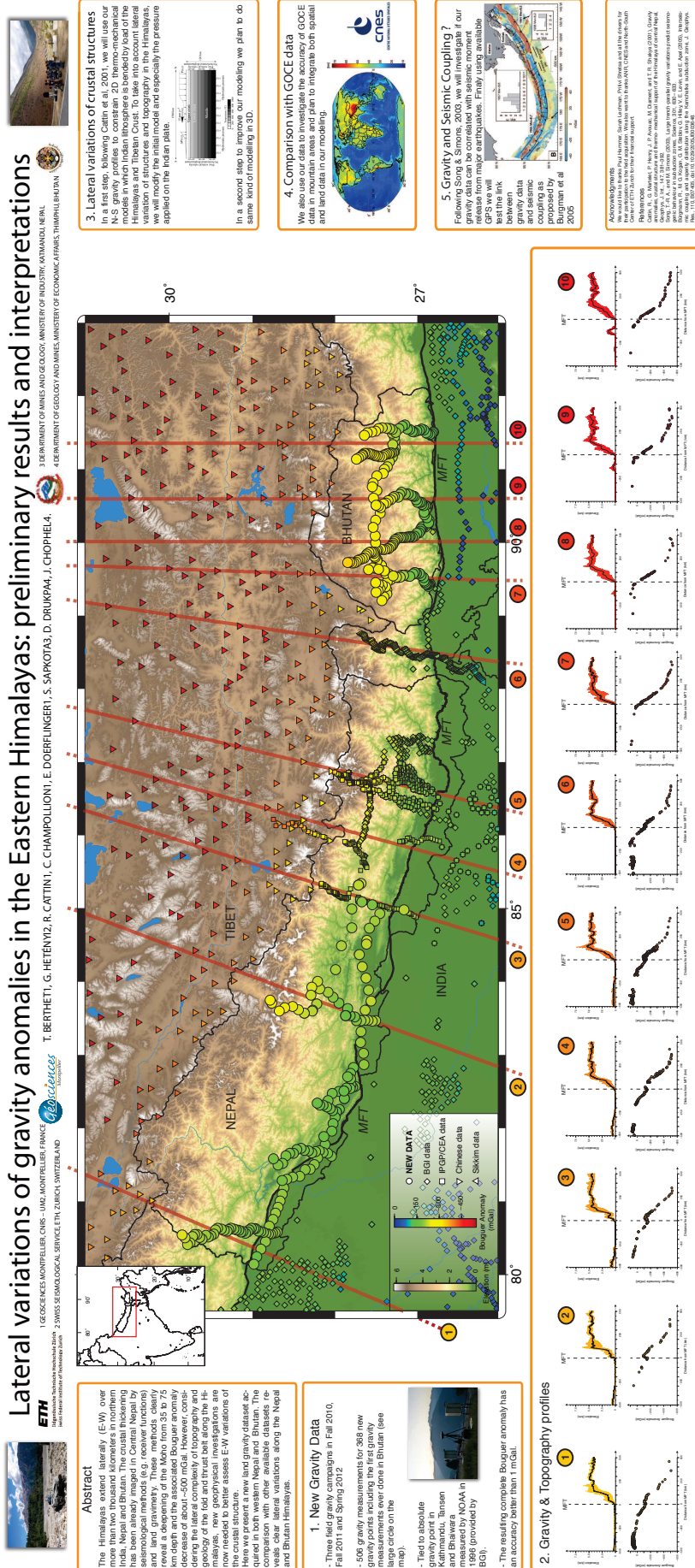
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The Himalayas are the most spectacular example of continental deformation in the world. They result from the evolution of the convergent margin between Indian and Eurasian plates and extend laterally (E-W) over more than two thousand kilometers in northern India, Nepal and Bhutan. In the same time, most of the crustal thickening is localized within only two hundreds kilometers (N-S) between India and Tibet. This pattern has been already imaged in Central Nepal by seismological methods (e.g. receiver functions) and land gravimetry. These methods clearly reveal a deepening of the Moho from 35 to 75 km depth and the associated Bouguer anomaly decrease of about 500 mGal. Hence, our current understanding of the Himalayas, especially for crustal structures and rheology, is based on interpretation and modeling of datasets acquired mostly across Central Nepal. However, considering the lateral complexity of topography and geology of the fold and thrust belt along the Himalayas, new geophysical investigations are now needed to better assess E-W variations of the crustal structure. Here we present a new land gravity dataset acquired during two field campaigns in Fall 2010 and 2011 in both western Nepal and Bhutan. Several N-S profiles have been measured including the first gravity measurements in Bhutan. Although new gravity points are in a wide range of elevation from hundred of meters to several thousands, the profiles show smooth decreasing gradients of Bouguer anomalies from South to North. The obtained new gravity profiles across the continental wedge confirm that gravity field is primarily controlled by the deepening of the Indian plate toward Tibet (longer, 100-km wavelength) and the geometry of the foreland basin (shorter, 10-km wavelength). The comparison with other available datasets reveals clear lateral variations along the Nepal and Bhutan Himalayas. To understand the observed variations, we perform 2D thermomechanical modelling of the flexure and compute synthetic Bouguer anomalies. We show that the E-W variations can be interpreted in terms of changes of the mechanical properties of the Indian plate.



A.1.2 WEGENER 2012

Along strike variations of crustal geometry in the Himalayas from new gravity anomalies data and numerical modelling (Oral)

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The crustal geometry of continental subduction zones is a key parameter to understand the geodynamics of collisional belts. In one hand, it enables to better assess the isostatic state of the range and the behaviour of the plates involved in the collision. On the other hand, the geometry of the crust beneath collisional mountain belt has an influence on the shape of the plates interface where largest earthquakes nucleate. The Himalayan arc is the most spectacular example of continental subduction. It results from the ongoing collision between Indian and Eurasian plates and extends laterally over 2500 km. Over the last three decades several geophysical studies including gravimetry and seismological methods have been undertaken to image this structure. These methods reveal a deepening of the Moho depth from 35 km to 75 km, which is associated to a Bouguer anomaly decrease of 500 mGal. But, as most these previous studies focused on Central Nepal, we can wonder whether our current understanding of the Himalayas also holds along the entire arc. To better constrain lateral variations of crustal structures along the Himalayas, 366 new gravity data points from Western Nepal to Eastern Bhutan have been acquired over five field campaigns. In addition to the available gravity anomalies dataset, these new measurements enable to compare several arc-perpendicular gravity anomaly profiles. Our results suggest east-west variation of the wavelength associated to the flexure of the Indian plate. Finally to understand these observed variations, we perform 2D finite-element thermomechanical modelling to assess the lateral variability of the effective elastic thickness of the Indian plate over more than 1200 km of the Himalayan arc.

Seismic cycle stress change along the Himalayas over the last 900 years (Poster)

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The Himalayan arc, which extends continuously along 2500 km, is one of the most spectacular manifestations of continental subduction. About 20 mm/yr of the present-day India-Eurasia convergence is accommodated across the Himalayas by interseismic loading and seismicity, which reveals some significant lateral variations, especially in the Bhutan Himalaya where present-day seismicity rate is low. This low seismicity is a matter of concern, as it may reflect two opposite fault behaviours : (1) a creeping zone or (2) a zone of stress accumulation for future great earthquakes. A host of information is now available regarding the major (M 8) earthquakes that occurred in this area over the last centuries, including Central-eastern Himalaya (1100), Central Himalaya (1505), Shillong plateau (1897), Kangra (1905), Bihar (1934) and Assam (1950). Taking advantage of previous paleoseismological studies, we combine major faults' geometry and estimated slip to compute stress interaction during earthquake cycles along the Himalayan arc over the last 900 years. Co-seismic stress due to major past earthquakes and secular stress loading of the major thrust fault system is computed using dislocation models in an elastic homogeneous half-space. Taken together, these two stress models result in a Coulomb failure stress model that predicts the state of stress change of the seismogenic frontal thrust fault in the Himalayas over the studied time-period. Several scenarios are tested to take into account the lack of constraints for the mean slip and the lateral extent of the past great earthquakes. Our results are compared to the present-day seismicity distribution along the Himalayan arc. We will discuss the relative effect of aseismic versus seismic deformation by studying the role played by interseismic loading. We will pay a specific attention to the Coulomb stress in Bhutan and to the robustness of our results with respect to the used parameters.

Seismic cycle stress change along the Himalayas over the last 900 years



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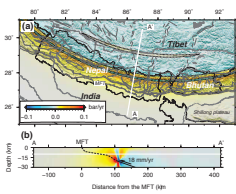


Eidgenössische Technische Hochschule Zürich
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Abstract

The Himalayan arc is a region of severe shaking and damage associated with frequent magnitude 7–8 earthquakes that can be related to major thrust faults and a high shortening rate. Over the last four decades many studies on historic and instrumental earthquakes as well as paleoseismological investigations have provided estimates of timing, magnitude and rupture parameters for these major events. Using the available information, we here focus on the central part of the Himalayas between 79–93°E and 25–31°N, i.e. from western Nepal to eastern Bhutan to calculate the Coulomb stress change in this area since the great historical rupture(s) reported in Nepal and in Assam region around A.D. 1100. We pay specific attention to the relationship between these major events and the interseismic loading, the stress interaction between these earthquakes as well as their effect on present-day seismicity rate.

Modelling approach



We calculate both coseismic and interseismic Coulomb stress change ΔCFF due to the major earthquakes and secular loading, respectively. We assume a homogeneous elastic half space, a constant interseismic loading of 18 mm/yr with no lateral variation (Fig. 2), and North-South "receiver" faults dipping 10° northward.

Figure 2: (a) Map showing the pattern of interseismic ACF rate at 10 km depth. White line AA' gives the location of the cross section displays on figure 2b. The location of the Main Frontal Thrust (MFT) is also given. (b) Vertical cross section showing ACF rate at depth along the profile AA'. White line gives the depth of map plotted on figure 2a. Dashed line gives the geometry of the locked portion of the Main Himalayan Thrust.

Spatial pattern of modern time Coulomb stress change

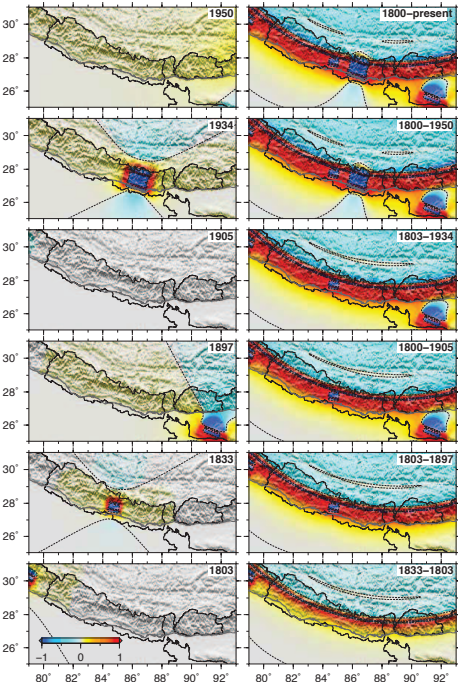


Figure 3: Coulomb stress change ACF along the Himalayan arc associated to both secular loading and largest earthquakes over the two last centuries. Date and time interval are indicated in the white box. Color scale gives ACF in bar. On the left, ACF due to each earthquake. On the right, cumulated ACF over the indicated time period.

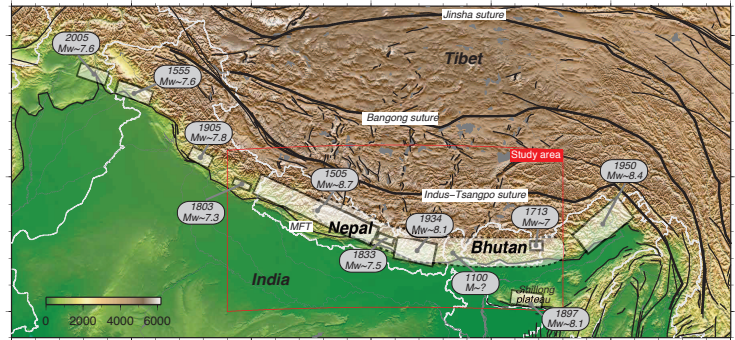
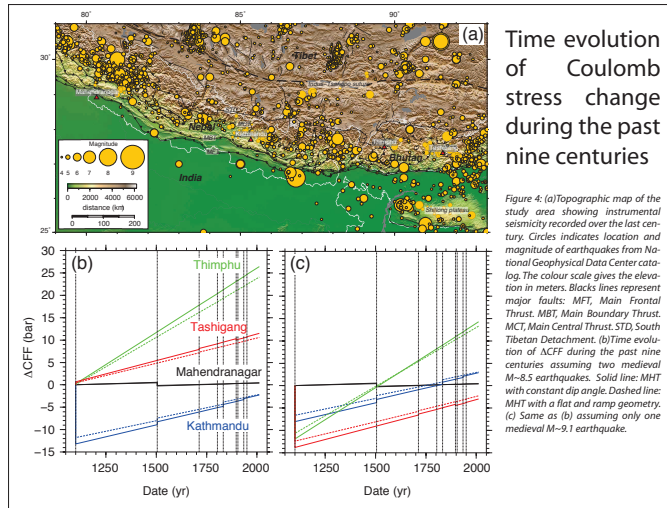


Figure 1: Simplified map of active faults in northern India, Nepal, and Tibet. The Main Frontal Thrust (MFT) and inferred rupture extent (shaded and labeled with year) of major historical earthquakes along the ~2500 km Himalayan arc are shown. The rupture extent for the 1505 central Himalayan earthquake (Mw=8.7), the 1555 Kashmir earthquake (Mw=7.6), the 1803 Kumaon-Garhwal earthquake (Mw=7.3), the 1833 Nepal earthquake (Mw=7.5), the 1905 Kanga earthquake (Mw=8.4), the 1934 Bihar-Nepal earthquake (Mw=8.1), the 1950 Assam earthquake (Mw=8.4), and the 2005 Kashmir earthquake (Mw=7.6) are adapted from published literature. Dotted rupture of the 1100 earthquake is speculative.



Time evolution of Coulomb stress change during the past nine centuries

Figure 4: (a) Topographic map of the study area showing instrumental seismicity recorded over the last century. Circles indicate location and magnitude of earthquakes from National Geophysical Data Center catalog. The colour scale gives the elevation in meters. Black lines represent major faults: MFT, Main Frontal Thrust; MBI, Main Boundary Thrust; MCT, Main Central Thrust; STD, South Tibetan Detachment. (b) Time evolution of ACF during the past nine centuries assuming two medieval M=8.5 earthquakes. Solid line: MHT with constant dip angle. Dashed line: MHT with a flat and ramp geometry. (c) Same as (b) assuming only one medieval M=9.1 earthquake.

Seismicity along the Bhutan Himalaya

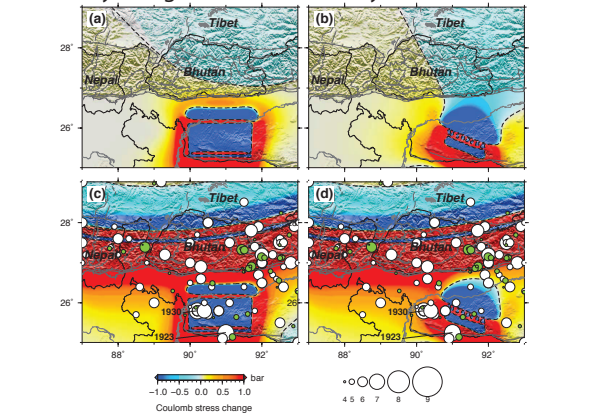


Figure 5: Coulomb stress change ACF along the Bhutan Himalaya. (a) ACF associated to the 1897 earthquake using the rupture model of Gahalaut & Chander (1992). (b) Same as (a) but using the rupture model of Biharam & England (2001). (c) ACF since the last major Bhutanesse earthquake in 1713 taking into account the cumulative effect of the 1897 Shillong from the Gahalaut & Chander (1992) model, the 1934 Bihar-Nepal and the 1950 Assam earthquakes and the interseismic loading. White circles give the magnitude and the location of earthquakes over the last century, for which the depth is unknown. Green circles are associated to shallow earthquakes (depth < 20 km). The location of the 1923 and 1930 earthquakes nearby the Shillong plateau is also given. (d) Same as (c) with the rupture model of Biharam & England (2001).

A.1.3 HKT 2012

Lateral variability of crustal geometry in the Himalayas from W-Nepal to Bhutan (Oral)

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The geometry of the crust beneath any collisional mountain range is a key information to understand the forces and dynamics that characterize the orogen. Compared to topography, crustal geometry determines the isostatic state of the orogen. The geometry of the underthrust plate is also relevant to assess its rheology and its effective elastic thickness. Furthermore, the overall crustal geometry exerts a control on the shape of the main plate interface or thrust fault system on which the largest earthquakes nucleate. Finally, the geometry of sub-surface structures is one of the most important input and control parameter of numerical models that aim to simulate the short- and long-term behavior of the orogen. In the Himalayas, studies on the above topics have all been undertaken. However, most studies focused on individual cross-sections across the Himalayas (mostly in Central Nepal), and proposed scenarios and models along 2D profiles. Whether the imaged structures and drawn conclusions also hold along the entire (ca. 2500 km long) Himalayan arc is seldom discussed due to difficult field conditions, and therefore also to the sparseness of available geophysical data. In the meantime, both topographical and geological maps suggest that significant variations exist along the arc, at least on surface. Investigating lateral variations at depth is therefore a prime target as the main geophysical characteristics (crustal thickness, flexural rigidity, maximum magnitude of potential earthquakes, etc.) may also very well vary along the entire Himalayan arc. In the past three years, we have focused our efforts on acquiring new geophysical data that can provide constraints on crustal geometry. In five field campaigns, we have carried out gravity measurements along several profiles perpendicular to the orogen from Far-West-Nepal to Eastern Bhutan, acquiring 366 new data points. This new dataset, complemented by existing data in Nepal and Sikkim, as well as further to the North and South, is now covering more than 1200 km of the Himalayan arc (Fig. A.1). This enables us to provide a first assessment of the lateral variability of crustal geometry by comparing numerous arc-perpendicular gravity anomaly profiles. Bouguer anomaly profiles crossing the orogen show a clear East-West variation in the shape of the underthrust India plate. Compared to the relatively low angle deepening of the crust in Central Nepal, it plunges at a steeper angle in Far-West-Nepal and in Bhutan. Thermo-mechanical numerical models that aim to explain this lateral variability invoke variations in both rheology and in the crustal structure of India entering the collision system. The lateral variability of structures, also seen on an interpolated map of existing Moho depth estimates, may well cause different seismogenic behavior in different parts of the arc. A first test to estimate regional earthquake hazard, following the method of (?), also will be presented.

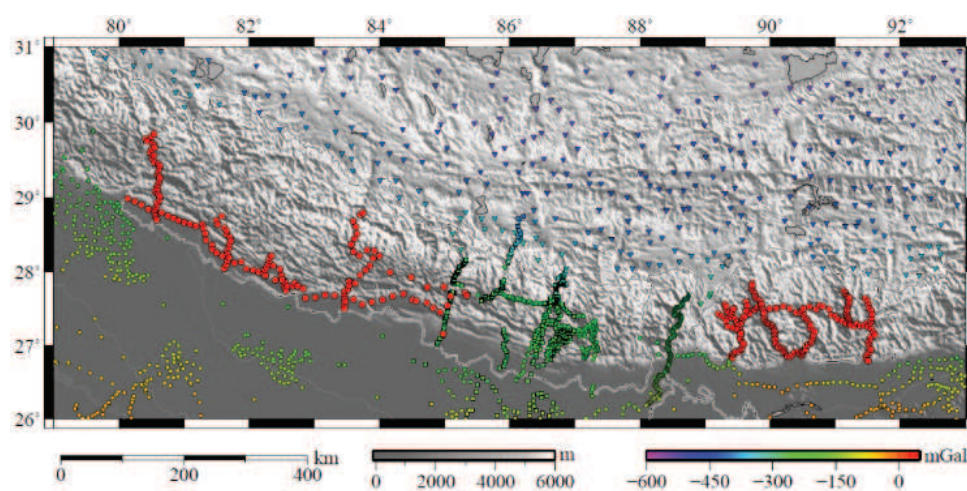


FIGURE A.1 Gravity data coverage in the Nepal, Sikkim and Bhutan Himalayas. Existing data is shown by colors according to Bouguer anomaly (in mGal). New data coverage is shown by red circles.

A.1.4 EGU 2013

Lateral variations of crustal structures in eastern Himalayas : a 3d modelling approach constrain by terrestrial and satellite gravimetry (Oral)

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Crustal geometry of collisional belts is a key information to constrain the evolution of orogens. The shape of the plates involved in the collision allows to determine the isostatic state of the range, to assess their rheologies and also to estimate their seismic potential. Over the last three decades, several geophysical studies, including gravimetry and seismology, have been undertaken in the Himalayas to decipher the deep structure of the orogen. Most of these studies focused on individual cross sections and 2D models, especially in Central Nepal. Due to the sparseness of available geophysical data the question of lateral variability of crustal structures remains unanswered. Topographical and geological map show significant variations along the arc and some studies proposed a segmentation of the Main Himalayan Thrust. Investigating lateral variations at depth is therefore crucial. In the past three years, we have carried out gravity measurements during four terrestrial gravimetry field campaigns. We have measured 366 new data points to fill the gap in Central and Western Nepal and in Bhutan. In addition with the existing data in Eastern Nepal and in Sikkim, as well as further to the North and South, terrestrial gravity is now covering more than 1200 km of the Himalayan arc. The Bouguer anomaly profiles across the arc show an East-West variation in the shape of the underthrust India plate especially between Nepal and Bhutan. The deepening of the crust beneath Bhutan is more abrupt than in Nepal. To clarify this variation in crustal structures, we jointly use the new terrestrial gravity dataset and satellite gravity gradients provided by GOCE to constrain a 3D thermo-mechanical flexural model.

Holocene slip rate and evidences of surface ruptures along the Main Frontal Thrust in Bhutan (Poster)

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On the long term, the Himalayas absorb about 20 mm/yr of the convergence between the India and Eurasia plates. Morphotectonics studies East of Bhutan yielded an Holocene slip rate of 23 ± 6.2 mm/yr consistent with the 21 ± 1.5 mm/yr estimated in Nepal. In the past 1000 years, most of this crustal shortening is accommodated along the Main Frontal Thrust (MFT) by major $M \geq 8$ earthquakes. Recent paleoseismic investigations between the meizoseismal areas of the 1934 Bihar-Nepal and 1950 Assam earthquakes west and east of the Bhutan suggest that a great earthquake may have ruptured a 700 to 800 km long section of the MFT during a single event around AD 1100. However, between 89°E and 92°E where the MFT runs along the southern border of the Kingdom of Bhutan, no major earthquakes has been recorded so far. In December 2012, we carried out a morphotectonic and paleosismologic study within two sites along the MFT in central Bhutan. We identified several sequences of abandoned fluvial terraces that have been uplifted along the fault. From a kinematic GPS survey, we estimated that the oldest terrace sequence (T2) displays a cumulative offset of 60 m. A younger sequence (T1) showed an offset of ~ 4 m on the lower terrace (T1a) and ~ 9 m on the upper one (T1b). We interpret this 4 m uplift as the vertical component of the last seismic event while the 9 m may be the cumulative offset of the two last earthquakes. These preliminary observations may suggest that the latest seismic event as well as the penultimate event correspond to two major $M \geq 8$ events. Samples for ¹⁴C, OSL and ¹⁰Be dating (charcoals, fine sands and quartz rich cobbles, respectively) have been collected in excavations within the different uplifted alluvial surfaces in order to determine the ages of these two last seismic events as well as the Holocene uplift and shortening rates. Compiled with the morphotectonic and paleoseismological data published in the literature, these new results will allow discussing the issue of seismic cycle along this portion of the Himalayan belt.

HOLOCENE SLIP RATE AND EVIDENCE OF SURFACE RUPTURE ALONG THE MAIN FRONTAL THRUST IN BHUTAN

J.-F. RITZ, T. BERTHELI, M. FERRY, D. DRUKPAZ, R. CATTINI, P. PELGANZ, J. CHOPHELZ, K. THINLEY, G. GEORGIENKO, MONTELLIER, CNRS - UNIV. MONTPELLIER, FRANCE

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INTRODUCTION
On the long term, the Himalaya absorb about 20 mm yr⁻¹ of the convergence between the India and Eurasia plates. Morphotectonic studies East of Bhutan yielded an Holocene slip rate of 2.3 ± 6.2 mm/yr consistent with the 21 ± 1.5 mm/yr estimated in Nepal. In the past 1000 years, most of this crustal shortening is accommodated along the Main Frontal Thrust (MFT) by major M8 earthquakes. Recent paleoseismic investigations between the mesoseismal areas of the 1954 Bihar-Nepal and 1950 Assam earthquakes west and east of the Bhutan suggest that a great earthquake may have ruptured a 700 to 800 km long section of the MFT during a single event around AD 1100. However, between 89°E and 92°E where the MFT runs along the southern border of the Kingdom of Bhutan, no major earthquakes has been recorded so far. In December 2012, we carried out a morphotectonic and paleoseismologic study within two sites along the MFT in central Bhutan.

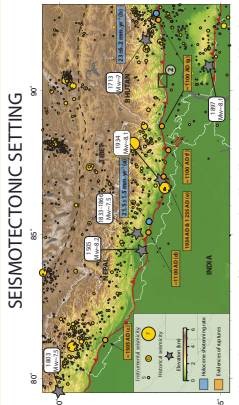


Figure 1. Seismotectonic map of the Eastern Himalayas. The red line represents the Main Frontal Thrust that accommodates most of the present-day shortening in the area. Yellow circles show the instrumental seismicity and black rectangles show the location of the study area. The black rectangle shows the location of the study area.

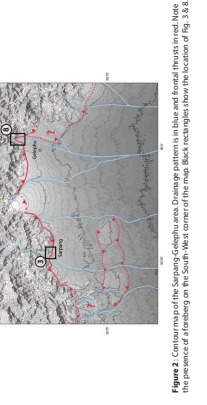


Figure 2. Contour map of the Sarpang Gephphu area. Drainage patterns in blue and fault thrusts in red. Note the presence of abandonment on the South-West corner of the map. Black rectangles show the location of Fig. 3 & 4.

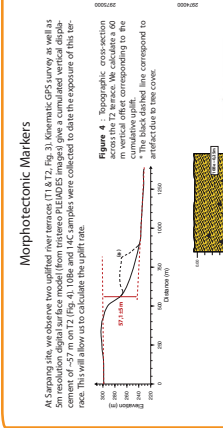


Figure 3. Topographic map of the Sarpang area. Resolution of the map is 10m. Red lines represent the T1 and T2 faults. Numbers in circles correspond to figure locations. Equidistance between contour lines is 10m.

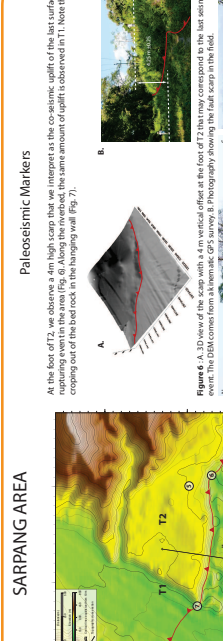


Figure 4. A. 3D view of the topographic map. The 3D view shows the topography of the area, with the T1 and T2 faults clearly visible. The map is oriented to show the fault scarps.



Figure 5. Stratigraphic cross-section across the T2 trace. The diagram shows soil profiles with different layers labeled. A black dashed line indicates the location of the soil profile shown in Figure 6.

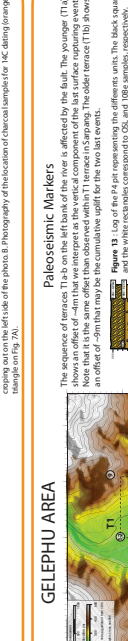


Figure 6. Stratigraphic cross-section across the T2 trace. The diagram shows soil profiles with different layers labeled. A black dashed line indicates the location of the soil profile shown in Figure 7.

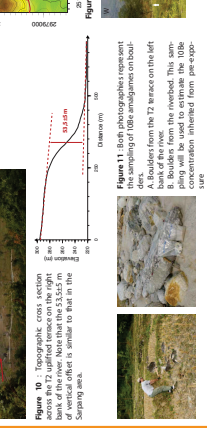


Figure 7. Stratigraphic cross-section across the T2 trace. The diagram shows soil profiles with different layers labeled. A black dashed line indicates the location of the soil profile shown in Figure 8.

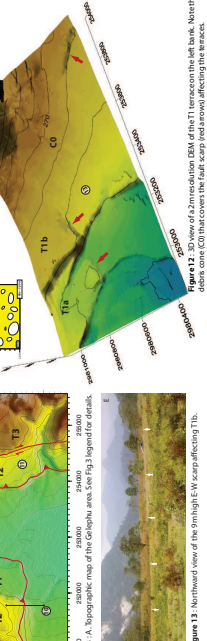


Figure 8. Stratigraphic cross-section across the T2 trace. The diagram shows soil profiles with different layers labeled. A black dashed line indicates the location of the soil profile shown in Figure 9.

Paleoseismic Markers
At the foot of T2, we observe a 4m high scarp that we interpret as the co-seismic uplift of the last surface rupturing event in the area (Fig. 6). Along the riverbed, the same amount of uplift is observed in T1. Note the cropping out of the black rock in the hanging wall (Fig. 7).

Morphotectonic Markers
At the Sarpang site, we observe two uplifted river terraces (T1 & T2, Fig. 3). Kinematic GPS survey as well as an instrumental seismicity (see also GEORGIENKO ET AL. 2012) suggest that the T1 and T2 faults are still active. This will allow us to calculate the uplift rate.

Morphotectonic Markers
At the Gephphu site, we observe the same sequence of uplifted river terraces (T1 & T2) with an additional higher terrace T3 present on the left bank. The cumulative uplift on T2 is the same as on the Sarpang site. This will allow us to calculate the uplift rate.

Paleoseismic Markers
The sequence of terraces T1 to T3 on the left bank of the river is affected by the fault. The younger (T3) terrace is more recent than the older (T1 & T2) terraces. Note that it is the same terrace on both banks of the river and T3 is offset by ~9m that may be the cumulative uplift of the last two events.

PERSPECTIVES
We identify sequences of abandoned fluvial terraces that have been uplifted along the fault. The younger (T3) terrace is more recent than the older (T1 & T2) terraces. Note that it is the same terrace on both banks of the river and T3 is offset by ~55 m. A younger sequence (T1) showed an offset of ~4 m on the lower terrace (T1a) and ~9 m on the upper one (T1b). We interpret this 4 m uplift as the vertical component of the last seismic event while the 9 m may be the cumulative offset of the two last earthquakes. These preliminary observations may suggest that the latest seismic event as well as the penultimate event correspond to two major M8+ events.

Samples for 14C, OSL and 10Be dating (charcoals, fine sands and quartz rich cobbles, respectively) have been collected in excavations within the different uplifted alluvial surfaces in order to determine the ages of these two last seismic events as well as the Holocene uplift rate and shortening rates are linked by the dip of the fault. Considering a 15° dip angle for the fault instead of a 30° one will increase the shortening rate by a factor of more than two.

Further missions are planned to do geophysical surveys of the fault to constrain its geometry. We also plan to open trenches at the fault scarp to confirm our interpretation in terms of surface rupturing events.

Uplift rate and shortening rates are linked by the dip of the fault. Considering a 15° dip angle for the fault instead of a 30° one will increase the shortening rate by a factor of more than two.

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ACKNOWLEDGEMENT
This work was supported by the French program: C.T.H.R.E.S. CNRS & TOC-COPE.

Preliminary study of E-W interseismic coupling variations on the Main Himalayan Thrust from western Nepal to central Bhutan (Poster)

PHILLIPE VERNANT¹, RODOLPHE CATTIN¹, KINZANG THINLEY², STÉPHANE MAZZOTTI¹, SOMA NATH SAPKOTA³, **THÉO BERTHET**¹, PHUNTSO PELGAY⁴, LAURENT BOLLINGER⁵, CÉDRIC CHAMPOLLION¹, DOWCHU DRUKPA⁴, FRANÇOIS JOUANNE⁶, JEAN-PHILLIPE AVOUAC⁷, ROGER BILHAM⁸, MIREILLE FLOUZAT⁵, JEAN CHÉRY¹, GYÖRGY HETÉNYI⁹,

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Compared to Nepal, seismogenic potential of Bhutan remains enigmatic due to its low present-day seismicity rate. Over the last three centuries, the only large reported event in this area is a Mw=8.1 earthquake in 1897 that occurred further south of the Main Himalayan Thrust (MHT) near the Shillong Plateau. New insights from a GPS network in Bhutan surveyed in 2003, 2006 and 2012 permit us to evaluate the convergence rate across the Bhutan Himalaya. Taking advantage of previously published GPS and levelling data in Nepal and Sikkim, we combine this new dataset with all the available velocity fields in a consistent India fixed reference frame. We also take into account the independent motion of the Shillong plateau relative to India and further evaluate the slip rate of the fault probably responsible for the 1897 earthquake. Assuming that interseismic deformations are related to dislocations in a homogeneous elastic half space, we invert surface displacements to assess the pattern of interseismic coupling on the MHT from western Nepal to central Bhutan. Our preliminary results indicate no major lateral variation in the coupling along the Himalayan arc in the study area. This suggests that elastic strain is building up along the front of Sikkim and Bhutan Himalaya similarly to what is observed along the Nepal Himalaya. Large earthquakes are probably required to release this strain. Despite the low historical seismicity of Sikkim and Bhutan, the seismic hazard might actually be as high there as along the rest of the Himalayan arc.

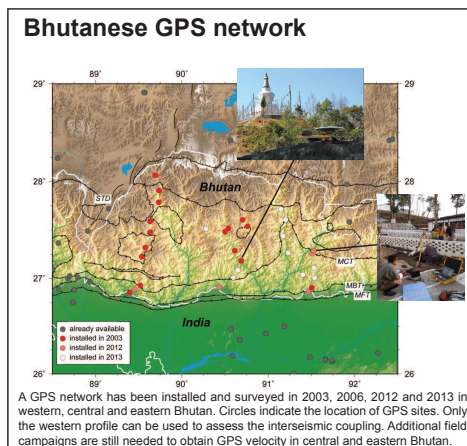
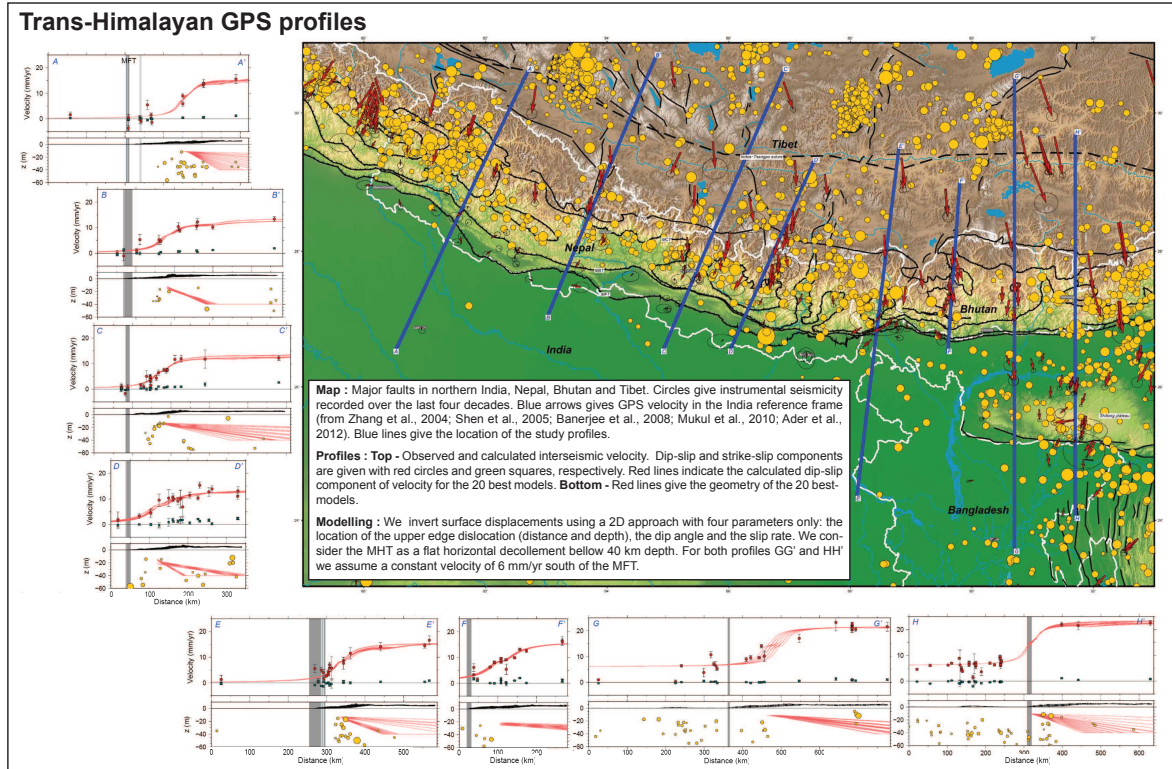
Along-strike interseismic coupling variations on the Main Himalayan Thrust from western Nepal to central Bhutan

R. Cattin¹, P. Vernant¹, K. Thinley², S. Mazzotti¹, S. N. Sapkota³, T. Berthet¹, P. Pelgay⁴, L. Bollinger⁵, C. Champollion¹, D. Drukpa⁴, F. Jouanne⁶, J.-P. Avouac⁷, R. Bilham⁸, M. Flouzat⁵, J. Chéry¹, and G. Hetényi⁹

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Abstract

Compared to Nepal, seismogenic potential of Bhutan remains enigmatic due to its low present-day seismicity rate. Over the last three centuries, the only large reported event in this area is a Mw=8.1 earthquake in 1897 that occurred further south of the Main Himalayan Thrust (MHT) near the Shillong Plateau. New insights from a GPS network in Bhutan surveyed in 2003, 2006 and 2012 permit us to evaluate the convergence rate across the Bhutan Himalaya. Taking advantage of previously published GPS and levelling data in Nepal and Sikkim, we combine this new dataset with all the available velocity fields in a consistent India fixed reference frame. We also take into account the independent motion of the Shillong plateau relative to India and further evaluate the slip rate of the fault probably responsible for the 1897 earthquake. Assuming that interseismic deformations are related to dislocations in a homogeneous elastic half space, we invert surface displacements to assess the pattern of interseismic coupling on the MHT from western Nepal to central Bhutan. Our preliminary results indicate no major lateral variation in the coupling along the Himalayan arc in the study area. This suggests that elastic strain is building up along the front of Sikkim and Bhutan Himalaya similarly to what is observed along the Nepal Himalaya. Large earthquakes are probably required to release this strain. Despite the low historical seismicity of Sikkim and Bhutan, the seismic hazard might actually be as high there as along the rest of the Himalayan arc.



Results

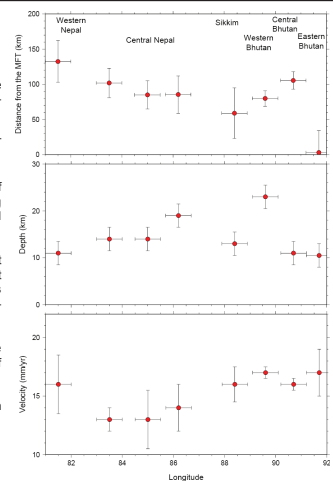
Our preliminary results indicate no major lateral variation in the coupling along the Himalayan arc from western Nepal to western Bhutan with

1. a locked ~100 km wide thrust fault (distance from the MFT between 60 and 130 km - depth from 8 and 24 km).
 2. a shortening rate of 11-19 mm/yr
- This suggests that elastic strain is building up along the front of western Bhutan Himalaya similarly to what is observed along the Nepal Himalaya. Large earthquakes are probably required to release this strain.

In central and eastern Bhutan the lack of a robust data-set leads to badly constrained conclusions. Our results suggest low coupling in eastern Bhutan whereas central Bhutan seems to behave as along the rest of the Himalayan arc. These results must be confirmed from additional data.

In addition ~6 mm/yr of convergence is observed between the MFT and ~150 km south of the Shillong plateau. The location of this convergence remains enigmatic.

Strike-slip component of the velocity field must be studied in future 3D numerical approach.



Acknowledgment : This work was supported by the French programs CT3-INSU CNRS, TOSCA-CNES and the GANSSER project from the Swiss National Foundation.

A.1.5 HKT 2013

Holocene vertical displacement rate along the Main Frontal Thrust in Bhutan from abandoned fluvial terraces study (Poster)

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HOLOCENE VERTICAL DISPLACEMENT RATE ALONG THE MAIN FRONTAL THRUST IN BHUTAN FROM ABANDONED FLUVIAL TERRACES STUDY

T. BERTHET¹, J.-F. RITZ¹, P. PELGAY², M. FERRY¹, D. DRUKPA²,
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INTRODUCTION

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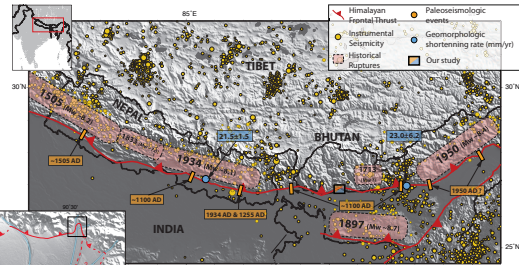


Figure 1: Seismotectonic map of the Eastern Himalayas. The red line represents the Main Frontal Thrust that accommodates most of the present day shortening in the area. Yellow circles show the instrumental seismicity and red rectangles the major historical events. Values of Holocene shortening rate are indicated in blue. Evidence of surface ruptures and the corresponding age are in orange. The black rectangle shows the location of Fig. 1a.

SARPANG SITE

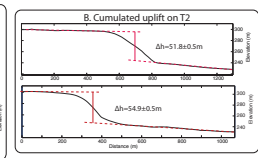
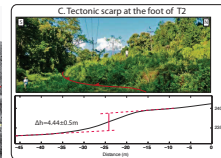
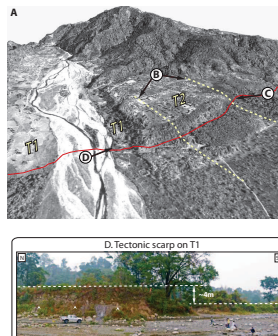
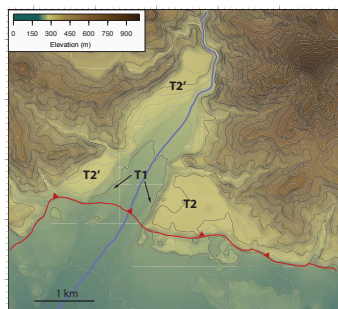


Figure 2: Topographic map of the Sarpang site. Thrust is shown in red and riverbed in blue. a) Perspective view of the left bank of Sarpang Khola (Pleiades image). Location of topographic cross sections and photos b) c) d) are shown.

At Sarpang, we observe a sequence of two uplifted river terraces T1 & T2 (Fig. 2). Both kinematic GPS survey and 5m resolution digital surface model (from tristereo PLEIADES images) give a cumulated vertical displacement of ~53.5 m on T2 (Fig. 2b). 10Be and 14C samples were collected to date the exposure of this terrace. At the foot of T2, we observe a ~4.4m high scarp (Fig. 2c). Along the riverbed, the same amount of uplift is observed in T1. Note the cropping out of the bedrock in the hanging wall argues in favour of a tectonic origin of this scarp (Fig. 2d).

GELEPHU SITE

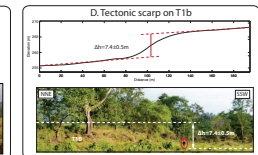
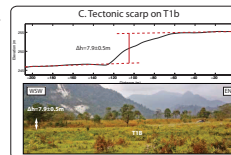
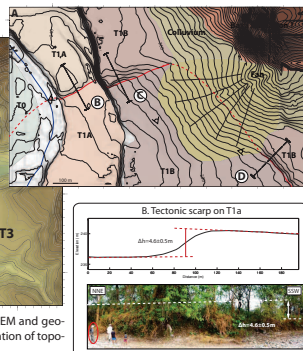
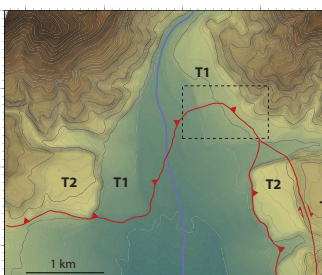


Figure 3: Topographic map of the Gelephu site. a) Detailed DEM and geomorphic map of T1 terrace on the left bank of Mao Khola. Location of topographic cross sections and photos b) c) d) are shown.

At Gelephu, we observe the same sequence of uplifted river terraces T1 & T2, with an additional higher terrace T3 preserved on the left bank (Fig. 3). The cumulated uplift on T2 is the same as on the Sarpang area. The sequence of terraces T1a-b on the left bank of the river is affected by the fault (Fig. 3a). The younger T1a shows an offset of ~4.6 m that we interpret as the vertical component of the last surface rupturing event (Fig. 3b). Note that it is the same offset than observed within T1 terrace in Sarpang. The older terrace T1b shows an offset of 7.9 m (Fig. 3c) that may be the cumulative uplift for the two last events.

DATATION

In order to date the terraces, we use radiocarbon dating as well as in-situ 10Be cosmogenic depth profiles. Three pits have been opened into T1a, T1b and T2. Ages constraints presented in Fig. 5 are constrained by radiocarbon dating for both T1a and T1b and by in-situ 10Be cosmogenic depth profile for T2. Details of the datation is not presented here but feel free to mail theo.berthet@gm.univ-montp2.fr for details. T1a and T1b are dated at 1520 ± 100 AD and 955 ± 100 AD, respectively. Exposure age of T2 from 10Be depth profile yield 6100 ± 1100 years.

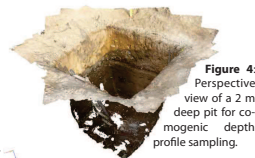


Figure 4: Perspective view of a 2 m deep pit for cosmogenic depth profile sampling.

RESULTS

We obtain a Holocene uplift rate of 9.35 ± 2.45 mm.yr⁻¹ that corresponds to a minimum rate of tectonic uplift as thrusting along the fault may lead to burial of the footwall. From the analysis of fault traces along T1a and T1b at Gelephu, we derive a dip of 20° at the surface. Further geophysical investigations are needed to determine the fault dip at depth. However, if we assume a fault dip of 20° and 9.35 ± 2.45 mm.yr⁻¹ of tectonic uplift rate, our results suggest a horizontal shortening of 25.7 ± 6.7 mm.yr⁻¹.

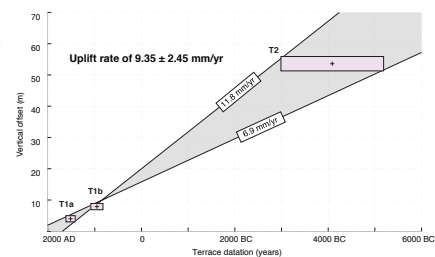


Figure 5: Uplift rate across the MFT in Bhutan from abandoned river terraces

For further details on this work (or post-doc opportunities), please feel free to contact theo.berthet@gm.univ-montp2.fr

Lateral variations of crustal deformations in the Himalayas

Abstract

Several major earthquakes have struck the 2500 km long Himalayan range during the past century. However, both the maximum size of such earthquakes and the probability of occurrence of a magnitude 9 megaquake are still a matter of debate. In this framework it is of key importance to investigate the lateral segmentation of the Himalayan arc in order to deepen our understanding of the mechanisms that control such events as well as their extents. Besides the remarkably uniform tectonic set-up and the fact that most studies on the Himalayas have focused on Central Nepal, several indications of along strike changes can be gathered. This thesis brings new constraints on lateral variations of crustal structure and of deformation between western Nepal ($\sim 80^\circ\text{E}$) and eastern Bhutan ($\sim 92^\circ\text{E}$).

The first part of this thesis focuses on the lateral variations of crustal structure. In the past three years we have carried out four land gravity campaigns. We have established 366 new gravity measurement points to fill the data gaps in central and western Nepal as well as in Bhutan. This new dataset, together with available seismological data, is used to constrain thermo-mechanical models of the bending of the India plate underneath the Himalayas. While the inferred crustal geometry does not exhibit major along-strike variations over central and eastern Nepal, the shorter wavelength flexure of the lithosphere in Bhutan is associated with a weaker upper mantle rheology.

The second part focuses on the lateral variations of recent deformation in the Himalayas. Fifteen years of paleoseismological investigations have extended the catalogue of major Himalayan events over the last millennium. Combining these information with secular loading, we assess the spatial and temporal stress changes on the Main Himalayan Thrust along the orogen over the last nine centuries. Our calculations indicate that inter-seismic loading has now nearly overcome the Coulomb stress decrease caused by the great ~ 1100 medieval earthquake. Our results also point out that the 1897 Shillong plateau earthquake does not have a major influence on the stress accumulated on the Main Himalayan Thrust since then. In order to better characterize active tectonics in the Bhutan Himalayas where no studies were done so far, we carried out a morphotectonic analysis in the south-central part of the kingdom. We show that the same amount of Holocene deformation is accommodated on the frontal thrust in Bhutan as on the neighbouring portions of the Himalayan arc. We also find evidences for two major ($M > 8$) earthquakes on this thrust in the Bhutan Himalayas during the last millennium. Our results therefore show that Bhutan cannot be considered as a seismic gap. They also challenge the interpretation of one single ~ 1100 medieval mega-event that would have ruptured the Main Frontal Thrust from central Nepal to eastern Bhutan.
