Investigating possible gravity change rates expected from long-term deep crustal processes in Taiwan

M. Mouyen,[1] M. Simoes,[2] F. Mouthereau,[3] F. Masson,[4] C. Hwang[5] and C.-C. Cheng[5]

[1]*Institute of Earth Sciences, Academia Sinica, Taipei* 115, *Taiwan, R.O.C. E-mail: maxime@earth.sinica.edu.tw*
[2]*Institut de Physique du Globe de Paris, Sorbonne Paris Cité, Univ. Paris Diderot, UMR* 7154 *CNRS, F-75005 Paris, France*
[3]*UPMC Univ Paris* 06, *UMR* 7193, *Institut des Sciences de la Terre de Paris, F-75005, Paris, France*
[4]*Institut de Physique du Globe de Strasbourg, IPGS - UMR* 7516, *CNRS et Université de Strasbourg (EOST), France*
[5]*Department of Civil Engineering, National Chiao Tung University, Hsinchu* 300, *Taiwan, R.O.C.*

SUMMARY

We propose to test if gravimetry can prove useful in discriminating different models of long-term deep crustal processes in the case of the Taiwan mountain belt. We discuss two existing tectonic models that differ in the deep processes proposed to sustain the long-term growth of the orogen. One model assumes underplating of the uppermost Eurasian crust with subduction of the deeper part of the crust into the mantle. The other one suggests the accretion of the whole Eurasian crust above crustal-scale ramps, the lower crust being accreted into the collisional orogen. We compute the temporal gravity changes caused only by long-term rock mass transfers at depth for each of them. We show that the underplating model implies a rate of gravity change of -6×10^{-2} µGal yr^{-1}, a value that increases to 2×10^{-2} µGal yr^{-1} if crustal subduction is neglected. If the accretion of the whole Eurasian crust occurs, a rate of 7×10^{-2} µGal yr^{-1} is obtained. The two models tested differ both in signal amplitude and spatial distribution. The yearly gravity changes expected by long-term deep crustal mass processes in Taiwan are two orders of magnitude below the present-day uncertainty of land-based gravity measurements. Assuming that these annually averaged long-term gravity changes will linearly accumulate with ongoing mountain building, multidecadal time-series are needed to identify comparable rates of gravity change. However, as gravity is sensitive to any mass redistribution, effects of short-term processes such as seismicity and surface mass transfers (erosion, sedimentation, ground-water) may prevent from detecting any long-term deep signal. This study indicates that temporal gravity is not appropriate for deciphering the long-term deep crustal processes involved in the Taiwan mountain belt.

Key words: Numerical approximations and analysis; Time variable gravity; Dynamics: gravity and tectonics.

1 INTRODUCTION

Temporal gravimetry is an efficient tool to monitor mass transfers, such as those implied by volcanic reservoir charge (e.g. Battaglia *et al.* 2008), groundwater movements (e.g. Jacob *et al.* 2010), or crustal thickening (e.g. Sun *et al.* 2009). The physical principle is that changes in the gravity value are due to changes in mass distribution. When an absolute gravimetre is used, the absoluteness of the measurement of gravity does not need further calibration or adjustment to a particular reference frame or datum (Van Camp *et al.* 2005). As every geodetic quantity, gravity variations integrate different processes. The challenge therefore relies in separating the contribution of each process in the measured signal. Here, we aim at testing the potential of this geodetic technique to study deep crustal

mass transfers during mountain building. We choose the Taiwan mountain belt as a test site, because it is characterized by extremely rapid erosion and deformation rates (e.g. Yu *et al.* 1999; Hu *et al.* 2001; Dadson *et al.* 2003; Willett *et al.* 2003), and therefore by potentially extremely rapid mass transfers at depth. In addition, gravity surveys have been performed across southern Taiwan since 2006 (Mouyen *et al.* 2013), in the framework of the Absolute Gravity in the Taiwan Orogen (AGTO) project. The modelling presented here aims at improving our understanding and interpretation of these field measurements.

The Taiwan mountain belt results from the ongoing collision of the Luzon volcanic arc (on the Philippine Sea Plate—PSP, Fig. 1) with the Chinese continental margin (hereafter CCM, on the Eurasian Plate—EP). Despite the numerous studies since the
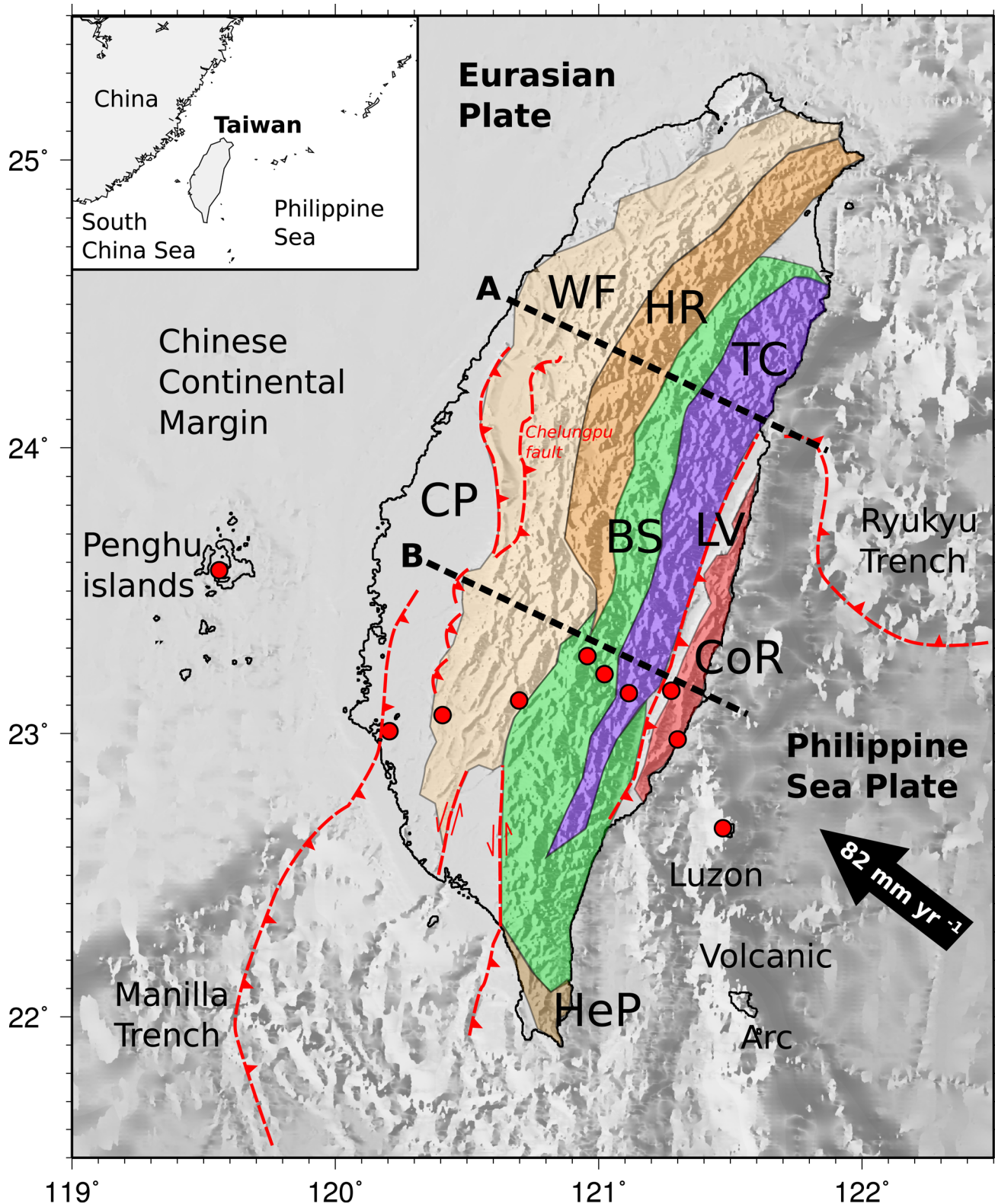
Figure 1. Map of the main tectonophysiographic ensembles in Taiwan (after Ho 1986; Simoes *et al.* 2012). CP, Coastal Plain; WF, Western Foothills; HR, Hsuehshan Range; BS, Backbone Slates; TC, Tananao Complex; HeP, Hengchun Peninsula; LV, Longitudinal Valley; CoR, Coastal Range. We have reported the Chelungpu fault (central Taiwan) that broke during the 1999 Chi-Chi earthquake. The red dashed lines are the major actives faults after Shyu *et al.* (2005). The red dots are the ten sites where absolute gravity measurements are performed since 2006 (Masson *et al.* 2008). The black dashed lines A and B are the locations of the cross-sections where the two models considered here (A and B, respectively) predict the kinematics and the density field used in our calculations.
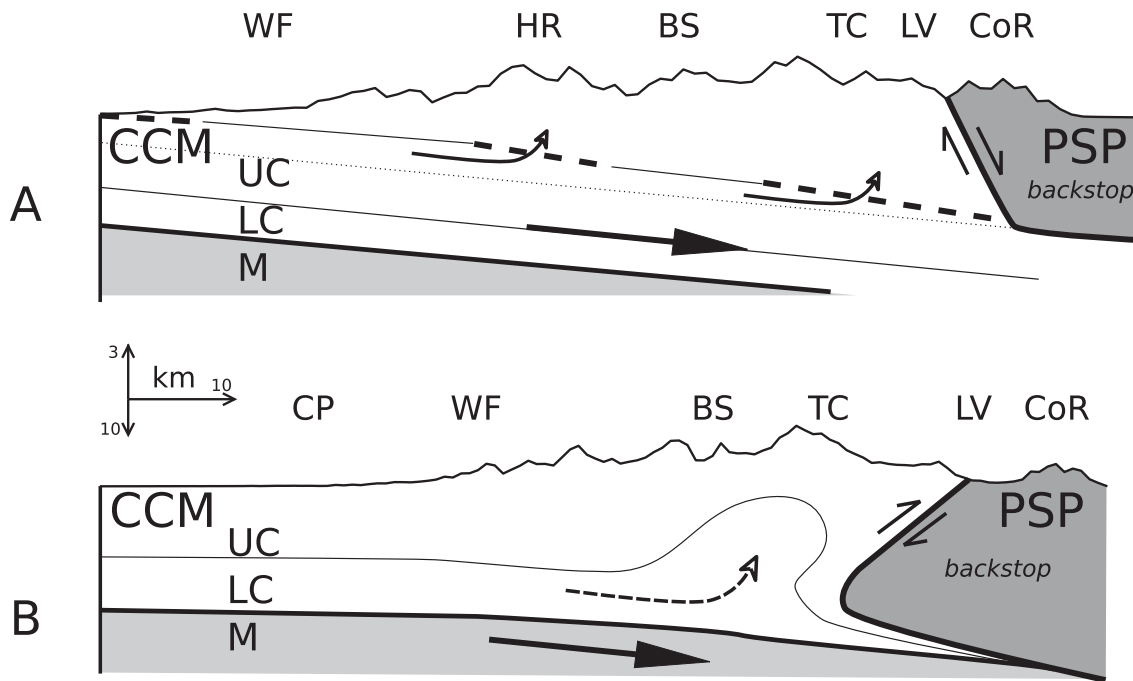
Figure 2. Sketches illustrating the processes proposed in models A (top, Simoes *et al.* 2007a) and B (bottom, Yamato *et al.* 2009) to account for the growth of the Taiwan orogenic wedge. In the case of model A, the underplating windows are indicated by dashed lines beneath the HR and the TC. Note that underplating under the WF in model A in fact represents frontal accretion of material into the wedge. As of model B, the flow of lower crust part beneath the most internal part of the wedge is represented by the dashed arrow below the Backbone Slates and the Tananao Complex. CP, Coastal Plain; WF, Western Foothills; HR, Hsuehshan Range; BS, Backbone Slates; TC, Tananao Complex; LV, Longitudinal Valley; CoR, Coastal Range; PSP, Philippine Sea Plate; CCM, Chinese Contiental Margin; UC, Upper Crust; LC, Lower Crust; M, Mantle.

1980's, the long-term deep crustal processes involved during mountain building remain debated and several models have been proposed in the literature (e.g. Simoes *et al.* 2012). Mountain building in Taiwan has first been approached by the critical taper wedge theory (Davis *et al.* 1983; Dahlen *et al.* 1984). Early models suggested that the orogenic wedge grows by frontal accretion of the crust of the CCM above a shallow décollement with 25 per cent of underplating at most (Barr & Dahlen 1989; Dahlen & Barr 1989; Barr *et al.* 1991). Frontal accretion is considered here as the discrete accretion of thin crustal slices at the front of a growing wedge. On the other hand, underplating represents the discrete accretion of crustal material from the lower to the upper plates (Fig. 2a) by duplexing below the inner portions of a mountain wedge (e.g. Dunlap *et al.* 1997; Konstantinovskaia & Malavieille 2005; Bollinger *et al.* 2006).

Recent data have questioned the details of these early models and suggested that underplating can contribute between 50 and 90 per cent of the incoming crustal flux into the wedge, in order to balance exhumation and erosion (Hwang & Wang 1993; Fuller *et al.* 2006; Simoes *et al.* 2007a). In general, whatever the contribution of underplating to mountain building, all these models suggest that only ∼7 km of the upper crust of the CCM are accreted to the orogen, a value that is essentially constrained by the accretion rates needed to compensate for exhumation and erosion at the surface (Suppe 1984; Simoes & Avouac 2006). Because the present-day crustal thickness of the undeformed CCM west of Taiwan is ∼30 km (Shih *et al.* 1998; Yeh *et al.* 1998; Yen *et al.* 1998; Kim *et al.* 2005; Lin 2005), this result implies that a significant portion of the downgoing crust could be subducted into the mantle, unless the underthrusted crustal thickness was significantly thinner, as suggested from recent seismic profiles southwest of Taiwan (Lester *et al.* 2013; McIntosh *et al.* 2013).

Considering the expected buoyancy of the CCM crust and the oceanic nature of the PSP, major crustal subduction is not expected to be significant. As expected for a young continental margin and as suggested by modelling of its long-term rheology, the CCM crust is probably weak (Mouthereau & Petit 2003; Mouthereau *et al.* 2013). For this reason, Yamato *et al.* (2009) proposed a crustal scale accretion model in which the whole 30 km of the CCM crust are accreted discontinuously into the orogen, through high-angle thrust ramps that affect deeper basement crustal levels (Mouthereau & Lacombe 2006), and in which exhumation is essentially sustained by a flow of the lower crust beneath the most internal part of the wedge (dashed arrow in Fig. 2b).

Simoes *et al.* (2012) noticed that, despite their differences, all these models fit low-temperature thermochronological data that correspond to the shallow exhumation history of rocks. Indeed, these models essentially differ in their proposed kinematics at depth and in the depth of the wedge. They pointed out the need for additional constraints on deep processes, including geophysical data, to improve the understanding of mountain building in Taiwan. In this paper, we investigate whether or not the study of present-day time variable gravity can prove useful in discriminating these different long-term models.

In a previous study, Mouyen *et al.* (2009) investigated the gravity change rate expected at short-term periods (years to decades) from two possible wedge geometries, either from a shallow (∼10 km) or deep (∼30 km) wedge, as proposed in different geometric models in the literature. Following Hsu *et al.* (2003), the short-term kinematics was driven by movements on the major faults of Taiwan in an elastic half space in order to model the possible interseismic elastic behaviour of active faults. The modelling was constrained by deformation rates observed by GPS measurements. Using material

densities proposed by Dahlen *et al.* (1984) and Lin & Watts (2002), the corresponding rate of gravity change with time was computed. Once free air effects were removed, it returned a maximum rate of 0.1 and 0.3 μGal yr^{-1} for the shallow and deep wedge models, respectively (1 μGal = 1 $\times$ 10^{-8} m s^{-2}). Only the elastic interseismic loading of active faults was modelled in this previous study, so that long-term deep crustal mass transfers were not taken into account.

Hereafter, we compute the gravity change rates expected from deep mass transfers considering two models of the long-term evolution of the Taiwan wedge. We find that these rates have opposite signs, suggesting that the two models tested could in theory be discriminated from gravity. However, these rates are two orders of magnitude lower than the usual uncertainty of gravimetres, thus requesting decades of measurements to retrieve any potential signal. We finally discuss the possibility of detecting such small signals since (i) short-term processes (earthquakes, landslides, hydrological variations) other than deep crustal mass transfers are likely to occur at much higher rates and (ii) deep crustal processes sustaining long-term mountain building are not expected to be linearly downscaled to short-term timescales, because they are discontinuous over time. These results critically lower the practical interest of using repeated land-based gravity measurements to unravel the long-term deep crustal processes in the Taiwan mountain belt.

2 GEOLOGICAL BACKGROUND

2.1 General settings

The collision between the CCM and the Luzon volcanic arc started 4–9 myr ago (Lin *et al.* 2003; Sibuet & Hsu 2004; Tensi *et al.* 2006; Beyssac *et al.* 2008). It has migrated southward at a rate of 31 to 80 mm yr^{-1} (Suppe 1984; Byrne & Liu 2002; Simoes & Avouac 2006), as a result of the oblique direction of convergence between the two plates. For this reason, different transects across the mountain belt can be taken as representative of different temporal stages of orogenic evolution, from an oceanic subduction to the south to mature collision in central Taiwan. The different tectonophysiographic units of the orogenic wedge have on average a nearly north-south azimuth. Following the nomenclature of Ho (1986), we find from west to east: 1) the Coastal Plain (CP), which corresponds to the foreland basin and is composed at the surface of quaternary alluvial deposits; 2) the Western Foothills (WF) fold and thrust belt, which form the deformation front of the orogen with Miocene to Quaternary sediments; 3) the Slate Belt (SB), which has a higher topography and is made mainly of Eocene to Miocene sediments that have been metamorphosed. In the northern part of Taiwan, the SB is divided into two units, the Hsuehshan Range slates (HR) to the west and the Backbone Slates (BS) to the east—the HR is absent in southern Taiwan—4) the Tananao complex (TC), which is composed of the exhumed and metamorphosed pre-Tertiary basement of the CCM; 5) the Hengchun Peninsula (HeP),which represents the accretionary prism of the CCM subducting below the EP in the southernmost part of Taiwan; 6) the Longitudinal Valley (LV), which corresponds to the suture between the two plates; 7) the Coastal Range (CoR), which is a remnant of the Luzon volcanic arc, composed of Neogene andesite rocks and turbiditic sediments. The Taiwan mountain belt is composed of accreted sequences from the CCM, and therefore corresponds to the different units located west of the LV. The present-day convergence rate across the whole plate boundary is of $\sim$82 mm yr^{-1} (Yu *et al.* 1997), and $\sim$40 mm yr^{-1}

of this total rate is accommodated over the long-term across the mountain belt (Simoes & Avouac 2006).

2.2 Tectonic models considered for modelling gravity change rates

As already discussed, several models have been proposed in the past to account for long-term mountain building processes in Taiwan. These models differ essentially in their long-term kinematics and in their wedge geometry (see Simoes *et al.* (2012) for a review). In the case of shallow wedges (Suppe 1984; Barr & Dahlen 1990; Fuller *et al.* 2006), with maximum depths of 10–15 km at the rear of the mountain belt, peak metamorphic temperatures are of $\sim$350 $^\circ$C at most and metamorphic grade does not go over the lower greenschist facies. Density contrasts at depth within the wedge, and hence any gravity signal related to these contrasts, are thus expected to be very limited. We therefore choose to consider here the models by Simoes *et al.* (2007a) and Yamato *et al.* (2009), with deeper wedge geometries ($\sim$30 km). They are expected to produce higher gravity signals and therefore to represent an extreme in the possible signals to be detected from long-term deep crustal mass transfers. These models have also the advantage of providing predicted rock densities at depth within the wedge. It is to note that in both models, only the Taiwan wedge west of the LV suture zone is considered, the LV and CoR being considered as the backstop of the wedge.

From a 2-D thermokinematic approach, Simoes *et al.* (2007a) suggest that underplating of the underthrust CCM beneath the orogen contributes significantly to the long-term growth of the Taiwan wedge. This model is adjusted to kinematic (Simoes & Avouac 2006; Simoes *et al.* 2007b,c), petrologic (Beyssac *et al.* 2007, 2008) and thermochronological (Lo & Onstott 1995; Tsao 1996; Liu *et al.* 2001; Willett *et al.* 2003; Fuller *et al.* 2006; Beyssac *et al.* 2007) data. In particular, the geometry at depth as well as the geometry of the LV fault is prescribed such as to fit peak metamorphic temperatures constrained from the Raman spectroscopy of carbonaceous material data. Because the $\sim$40 mm yr^{-1} shortening rate across the wedge is accommodated mostly (if not at all) by the frontal faults of the foothills, Simoes & Avouac (2006) propose that exhumation of the internal portions of the orogen is sustained by underplating. Based on thermochronological and metamorphic data, underplating is suggested to be localized beneath the HR and TC. The fit to all the data implies temporal changes in the kinematics, with an increasing contribution of underplating over time, in particular since 1.5 Myr ago. Only the uppermost 7 km of the downgoing crust contribute in this model to the growth of the wedge, suggesting that a significant portion of the CCM crust has been subducted into the mantle or that the underthrust crust was thinner than that presently observed west of Taiwan (Lester *et al.* 2013; McIntosh *et al.* 2013). The modelled pressure and temperature fields are used to compute the final density field within the whole wedge, from the methods described by Bousquet *et al.* (1997) and Henry *et al.* (1997). Hereafter, this model will be referred to as model A (Fig. 2a).

Using a fully-coupled 2-D thermomechanical model, prescribing only plate convergence rates, Yamato *et al.* (2009) suggest that exhumation and crustal thickening patterns within the Taiwan range are accounted for by the accretion of the whole CCM crust and by a lower crustal influx at the rear of the Taiwan wedge. Their model is adjusted to thermochronological data (Fuller *et al.* 2006; Beyssac *et al.* 2007), erosion and sedimentation rates (Dadson *et al.* 2003; Lin *et al.* 2003; Willett *et al.* 2003; Tensi *et al.* 2006), heat flow (Lee & Cheng 1986) and deformation patterns observed in southern

Taiwan, where the HR does not exist (Mouthereau & Lacombe 2006) or has not yet be exhumed (Simoes & Avouac 2006; Simoes *et al.* 2007a). To evaluate the changing densities of the mineralogical phases due to the changes of pressure and temperature conditions in the course of increasing plate convergence, a thermodynamical code (De Capitani 1994) is coupled to the thermomechanical model. This model implies that the crust of the CCM is essentially not subducted and is rather decoupled from the EP lithospheric mantle. Such a decoupling would result from the inherited rheology of the CCM (Mouthereau & Petit 2003; Mouthereau *et al.* 2013) and would account for the exhumation of the hot and ductile lower crust beneath the TC and BS. This model also implies temporal changes in kinematics with a thickening of the lower crust focused beneath the internal portions of the wedge, and recent propagation of the deformation towards the foothills over the last 1–2 Myr. Hereafter, this model will be referred to as model B (Fig. 2b).

It should be noted here that not only do models A and B differ in their kinematics and the processes that account for the long-term growth of the wedge, but they also do differ in the thermodynamical grids used to compute wedge densities. The kinematic and density grids predicted by these 2 models correspond to two different transects across Taiwan (Fig. 1). In particular, model A is located further north than the area where gravity surveys are conducted. This should not be critical to what concerns the TC because metamorphic and thermochronological data suggest that exhumation has either reached a steady state or has been synchronous along strike within the TC laterally from latitudes 23 °N to 24.5 °N (Simoes *et al.* 2012). The major differences between the two transects considered by models A and B rely on the presence or not of the HR. As further discussed below, this does not impact our conclusions.

3 GRAVITY MODELLING

3.1 Modelling approach

The primary objective of the following computations is to compare the gravity signal produced over time by tectonic models A and B. These models have comparable kinematics at shallow depths since they both fit the same low-temperature thermochronological data (Simoes *et al.* 2012). This suggests that the gravity signal produced in both models by this shallow kinematics should be comparable. The most significant difference between models A and B should therefore only rely in the gravity signal related to their deep crustal kinematics, either underplating or lower crustal flow (Fig. 2). To quantify the gravity signal produced only by these different deep crustal kinematics, we therefore choose not to compute the signal related to shallow processes. Practically, we do this by artificially keeping topography flat at sea-level and by not computing any gravity signal related to the kinematics above sea-level. This implies that the gravity signals computed here do not correspond to any actual signal that could be measured at the surface, but to the only contribution of deep crustal mass processes to any observable signal.

The velocity and density fields predicted by models A and B are presented in cross-sections (Figs 3 and 4 for models A and B, respectively). These A and B cross-sections were provided at resolutions of 10 × 5 km and 2 × 4 km (horizontal × vertical), respectively, and are converted into 2-D regular grids at a 2 km resolution. The velocity and density fields are interpolated to this grid. The topography is not taken into account and the upper surface of the wedge is kept flat and at sea level. We first compute the vertical component (as if measured by a gravimetre) of the initial gravity field g_i at

the surface for both tectonic models using the GRANOM software (Hetényi *et al.* 2007). The grids are then deformed according to the respective velocity fields. The displacement is computed for 1 yr and is distributed at the edges of the different cells so that they remain connected.

The resulting gravity field g_f is then calculated and used to compute $g_f - g_i$ for models A and B. This gives the rate of gravity change as a function of the horizontal distance along the section. As both models imply horizontal advection, $g_f - g_i$ is computed at the same surface site but not at the same horizontal distance along the section. Indeed we model gravity changes as if they were measured by a gravimetre, that is at a site that moves with the ground. The velocity fields produced by both models A and B will also move the surface vertically. Vertical displacements have a significant effect on the gravity signal, mostly because the measurement position is moved away or closer to the Earth's centre of mass (Hofmann-Wellenhof & Moritz 2006). As we focus on gravity changes induced only by deep crustal mass transfers, the effect on the computed gravity of these vertical displacements must be removed. Therefore, in our calculations, we do not take into account the contribution of this ground motion to the computed gravity signal, and topography remains artificially flat and at sea level. This is a key point in our study because one of our purposes is to verify whether or not the different deep kinematics implied by these two models can be distinguished from gravity data. Because our modelling only spans 1 yr, we have neglected mineralogical phase and associated density variations, that is, we consider a constant density grid. Indeed, phase equilibrium mainly occurs at Myr timescales (e.g. Sobolev & Babeyko 1989; Peacock 1991; Henry *et al.* 1997). The results are presented in Figs 5(a) and (b) (plain blue line). For both models, we also computed separately the effects of vertical (dashed red line) and horizontal (dashed green line) components of the velocity field on the gravity signal.

3.2 Temporal gravity changes with model A

Model A implies an overall decrease of gravity (negative gravity change rate), down to a value of -6×10^{-2} μGal yr^{-1} around the LV (Fig. 5a, plain blue line). This behaviour is due to the contrast of velocity and density between the orogenic wedge and the subducted lower CCM. If we first only consider the contribution of the horizontal advection of material to the gravity signal (Fig. 5a, dashed green line), the density boundary between the subducting CCM and the orogenic wedge moves eastward relative to the surface, where gravity is computed. West of the LV, the denser material of the lower CCM and mantle is thus closer to surface inducing a gravity increase (positive gravity change rate). East of the LV, as the densities reverse between the CCM and the lower PSP, gravity decreases. If we now only consider the contribution of the vertical component of the velocity field (Fig. 5a, dashed red line), gravity decreases mostly west of the LV, because denser material moves deeper. This decrease diminishes eastward, because of the reversal in the density values at the boundary between the PSP and the CCM. Model A also implies a very local increase of the gravity, with a rate of 2×10^{-2} μGal yr^{-1} at most near the limit of the TC and the LV. This local increase of the gravity change rate is due to the rise of denser materials below the TC because only the vertical component of the velocity field contributes to this gravity change signal. On the other hand, below the HR, both the upward velocity and the density contrasts of the materials are too weak to significantly change the gravity.
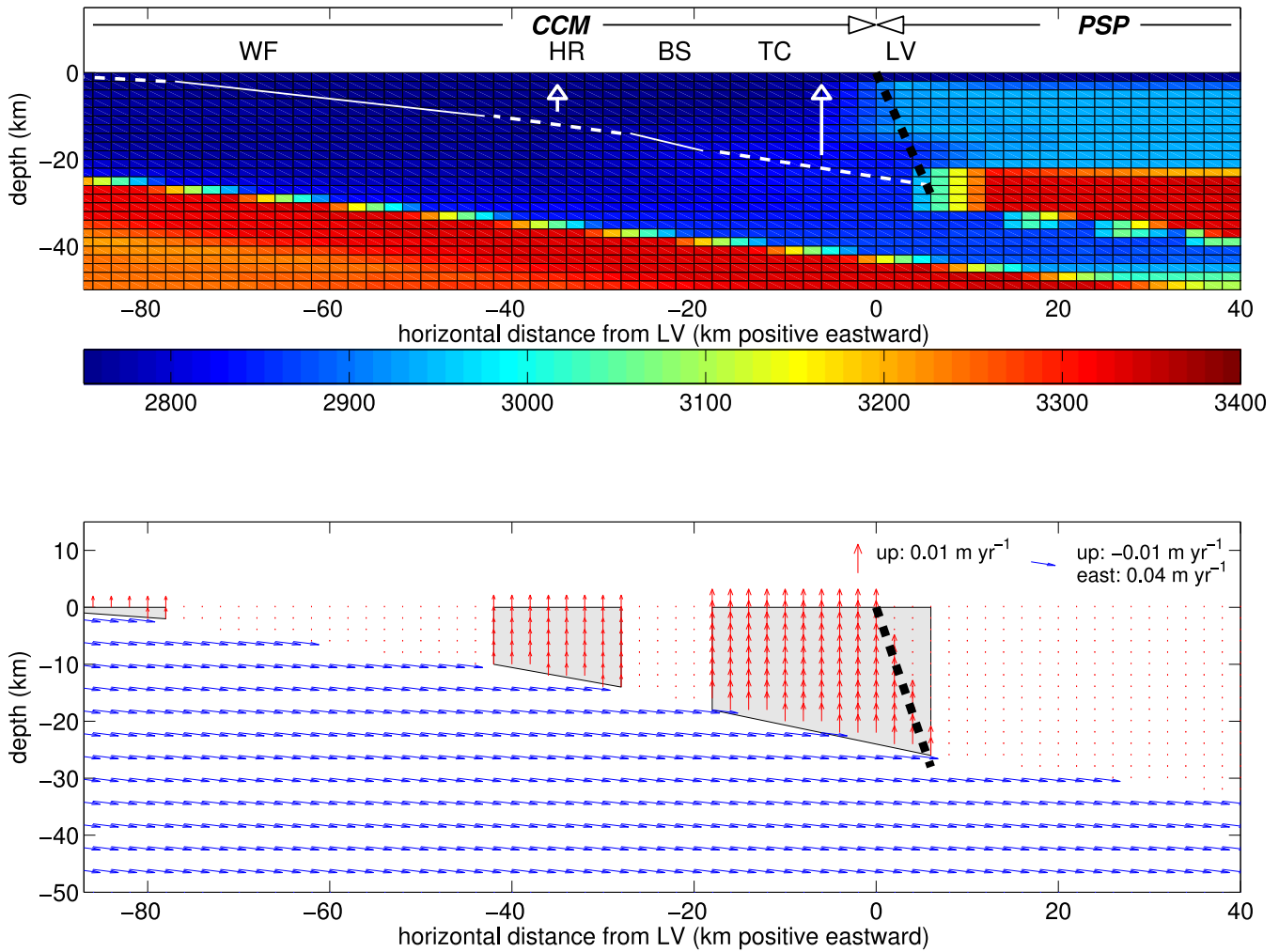
Figure 3. Top: Density field for model A as proposed by Simoes *et al.* (2007a). The colour bar gives density values in $kg\,m^{-3}$. The white lines delineate the wedge and, if dashed, the underplating window, as in Fig. 2(a). In the East, the denser material between 20 and 30 km depth is the lower crust of the PSP. Bottom: Kinematics of long-term displacements relative to the backstop (CoR and PSP). The black-dashed line indicates the Central Range fault along the Longitudinal Valley. The grey patches correspond to the regions where underplating is supposed to occur. For more legibility, the scales of the velocity arrows are different within the wedge (red) and within the downgoing crust (blue), the latter are horizontally three times shorter than the former. WF, Western Foothills; HR, Hsuehshan Range; BS, Backbone Slates; TC, Tananao Complex; LV, Longitudinal Valley; PSP, Philippines Sea Platered; CCM, Chinese Continental Margin.

These different results suggest that crustal subduction contributes the most to the predicted gravity change rate in model A. To further test this, we quantify the contribution of only underplating to the gravity signal in model A. This provides also an extreme estimation of the gravity change rate expected for such model of mountain-building in the case that crustal subduction is overestimated in model A, in particular if the underthrusted crust was much thinner than the one observed today within the passive margin. For that, we artificially remove the subduction movements and compute the gravity effect of mass transfers only above the underplating interface, within the orogenic wedge. The result is shown by the cyan line in Fig. 5(a). We still observe the local increase of gravity in the TC located above the largest window of underplating (Fig. 3). However, it is related to the shallow rise of material (at $z \simeq 10$ km) rather than the input of mass at the $\sim$25–30 km deep underplating window. Indeed, such deep rise of mass would create a wider gravity increase, not observed here. None of the underplating windows has a clear effect on gravity changes, because the density contrast at these interfaces is too low. Our results indicate that crustal subduction contributes the most to the gravity signal. The two estimates

(with and without subduction) provide a lower and upper bound for the gravity changes expected for model A.

3.3 Temporal gravity changes with model B

Model B implies an increase of the gravity change rate over a $\sim$60 km wide region, from the BS to the LV. It reaches $7 \times 10^{-2}\,\mu Gal\,yr^{-1}$ at most in the TC (Fig. 5b, plain blue line). The decomposition of the gravity signal shows that it is mainly related to the horizontal component of the displacements. Indeed, the velocity field in model B implies that rocks at the surface move eastward faster than rocks at depth. Moreover, densities are greater at depth to the east, especially east of the LV. Therefore, the cells located at the surface get closer to this dense material during deformation, while the deeper cells stay at the same place. Because gravity is computed at the surface, its value increases. Vertical rock displacements contribute less to the gravity signal. Their effect is confined to the BS and TC, where an increase of $2 \times 10^{-2}\,\mu Gal\,yr^{-1}$ is predicted, as a result of the rise of denser material below this region. Like
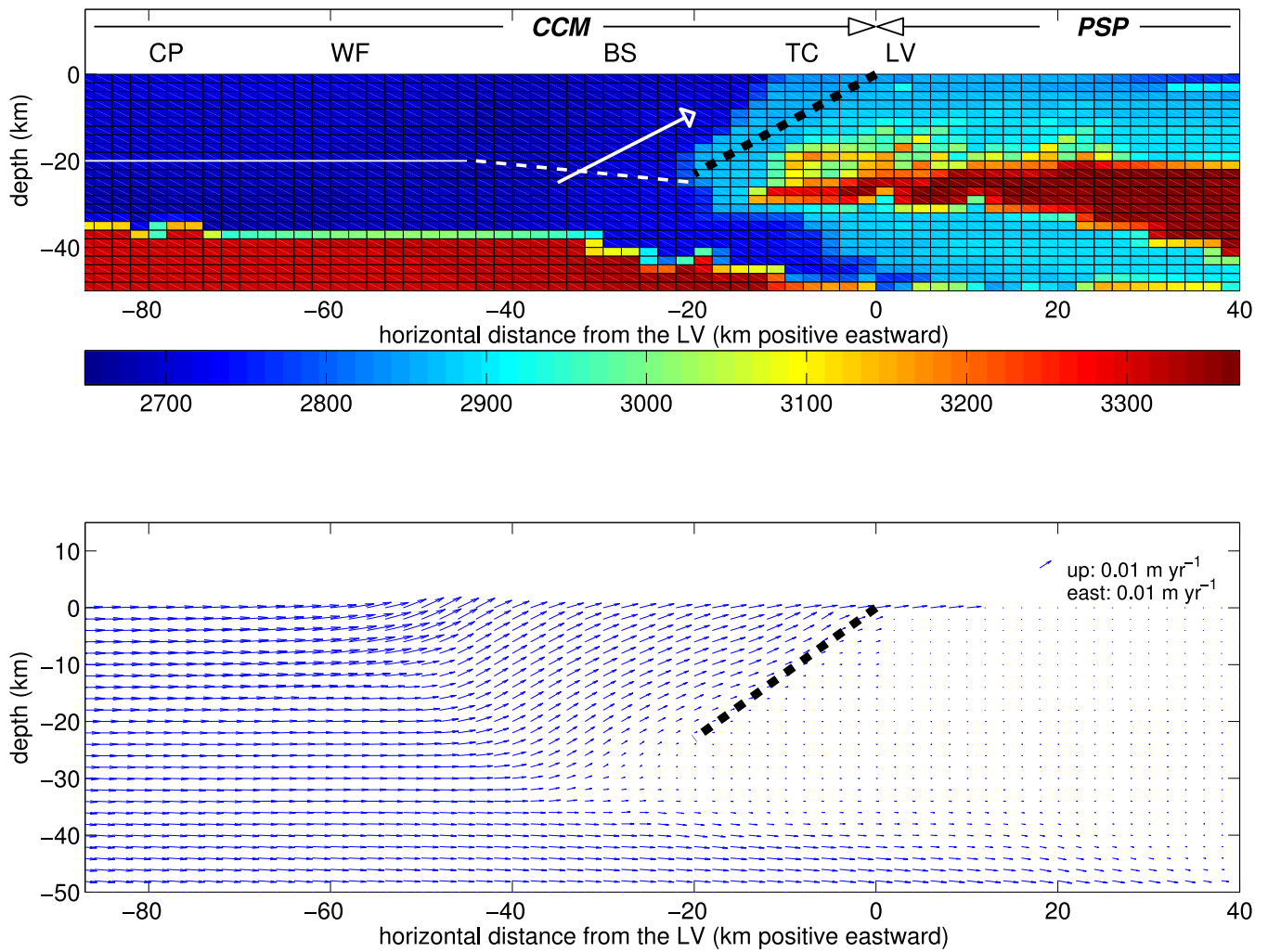
Figure 4. Top: Density field of model B as proposed by Yamato *et al.* (2009). The colour bar gives density values in kg m^{-3}. The white lines delineate the wedge and, if dashed, the window with an input of lower crust, as in Fig. 2(b). Bottom: Kinematics of model B, the black-dashed line shows the Central Range Fault along the Longitudinal Valley. CP, Coastal Plain; WF, Western Foothills; HR, Hsuehshan Range; BS, Backbone Slates; TC, Tananao Complex; LV, Longitudinal Valley; PSP, Philippines Sea Plate; CCM, Chinese Continental Margin.

model A, most of the density contrasts at depth beneath the BS and the TC are too low to generate a noticeable gravity signal at the surface. The deep crustal mass transfers implied by model B thus have a low influence on gravity changes rates.

4 DISCUSSION: LONG-TERM AND SHORT-TERM TECTONIC AND SURFACE CONTROL ON GRAVITY CHANGES

It should be reminded here first that the gravity signals computed for models A and B and discussed hereafter are only meant to represent the contribution of their corresponding deep crustal mass transfers to any possible gravity signal measured at the surface, and not the actual total gravity signal.

Models A and B differ not only in their deep kinematics but also in the procedures used to compute densities. Because the amplitude of the gravity change rates computed here is small and of the same order of magnitude, we propose that these differences have little influence on our main conclusions. Comparing the results obtained

from these two models may also not seem straightforward given the fact that these models aim at representing two different sections across the orogen (Fig. 1). However, Simoes *et al.* (2007a) found that the temperatures computed within the wedge increased by ∼40 °C at most from their southern to northern transects, essentially below the western underplating window sustaining the exhumation of the HR. Such small variations in temperature are not expected to induce significant variations in the density field within the wedge in western Taiwan, given the probable pressure (<15 km depth) and temperature conditions below the HR and WF. This suggests that the density profile of model A is still appropriate to compute the potential contribution of underplating to the gravity change rate in southern Taiwan, where gravity surveys are being conducted.

In the case of model A, the signal has a long wavelength, dominated by the subduction of the lower CCM. The most significant rate of gravity change is obtained in the TC: -6×10^{-2} μGal yr^{-1}. In the case that subduction is overestimated in the original model A, a value between -6×10^{-2} (with subduction) and 2×10^{-2} μGal yr^{-1} (subduction entirely neglected) seems more realistic. In the case of model B the gravity change rate is also maximum in the TC, with 7×10^{-2} μGal yr^{-1}. These two models therefore imply gravity
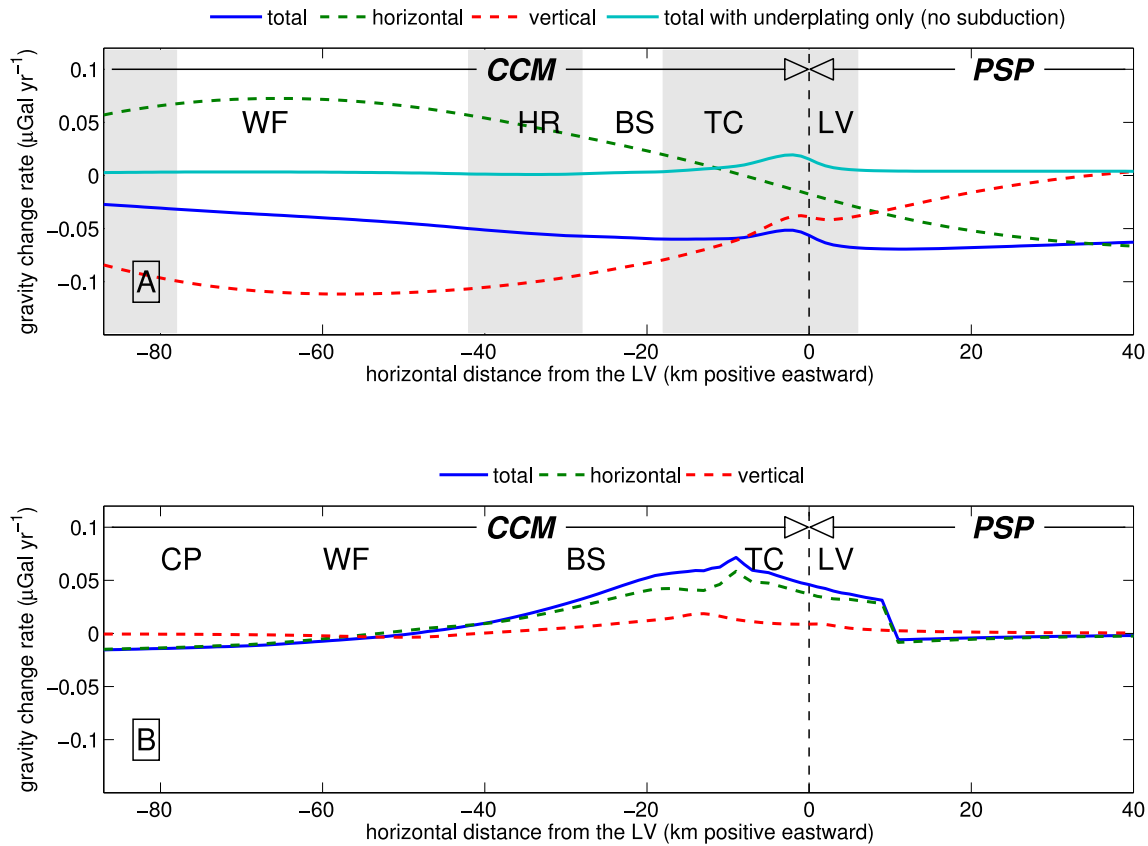
Figure 5. Gravity change rates predicted from the kinematics and density field from model A (top) and model B (bottom). The model inputs are those shown in Figs 3 and 4. The total gravity change rates computed are illustrated by the blue line. The effect of the horizontal and vertical displacements are shown by the dashed green and red lines, respectively. The vertical line corresponds to the surface position of the Central Range fault. In the case of model A the location of the underplating windows is indicated by the grey patches. We also computed a fourth signal (cyan line) that represents only the effect of underplating on the gravity signal, that is mass transfers above the basal décollement and therefore within the orogenic wedge; the crustal subduction of the CCM is neglected in this case.

change rates of the same order of magnitude, however with varying wavelength and with opposite signs in the case of significant crustal subduction in model A. No particular signal is associated with the growth and exhumation of the HR (model A). Models A and B both propose rather deep geometries (~30 km) for the Taiwan orogenic wedge. We did not test here the gravity change rates implied by shallow (~15 km deep) wedge models proposed in the literature (Suppe 1984; Barr & Dahlen 1990; Fuller *et al.* 2006). This is first because densities have not been computed in these models. In addition, the pressure (~15 km) and temperature (~350 °C) conditions at the rear of these shallow wedges are similar to those predicted at the base of the HR in model A, where our computations indicate no particular signal. By analogy we suggest that shallow wedge models might not produce any significant effect on gravity change rates. Taken all together, these models of mountain building in Taiwan will theoretically create distinguishable gravity signals related to their deep crustal kinematics (at least if we consider an important contribution of subduction in model A). However, the overall amplitude of the modelled gravity change rates is weak, at least 30 times lower than the present-day uncertainty of the best absolute gravimetres available (Niebauer *et al.* 1995; Van Camp *et al.* 2005). This indicates that such signals cannot be presently detected, preventing from favouring one or the other of these long-term models. The low amplitudes of the computed signals result from the low density contrasts involved in mass transfers predicted in these models, as well as from the depth of these mass transfers

(the direct gravitational attraction decreases with the square of the distance between the mass and the location where it is applied).

In comparison with our previous results of gravity change rates expected at a shorter (interseismic) timescale (Mouyen *et al.* 2009), the long-term mountain building effect is five times smaller. As a result, the gravity signal induced by short-term tectonic processes is expected to contaminate any possible contribution of long-term deep crustal mass transfers. Theoretically, the identification of such small rates (even those related to the interseismic loading of the faults) needs long time-series (several decades), to allow the gravity changes to accumulate and reach measurable values. However, even decades of measurements would still correspond to short geological timescales (yr to 100 yr), when compared to longer-term mountain-building processes (~Myr). It should be emphasized here that any signal related to a long-term process can be expected to be measured at a short human timescale provided that the process producing this signal is linear over time (over timescales of 10s yr to Myr). However, this should not be the case, neither for model A, nor for model B. For model A, underplating related to ramp migration is a discontinuous process over time. In such a model, only interseismic loading and the subsequent coseismic and postseismic displacement fields are expected to occur (and therefore be measured) at short human timescales. The long-term displacement field of model A (Fig. 3) results from the sum of these discontinuous short-term deformation patterns over several seismic cycles, coupled to a discrete migration of the ramp at depth. For model B, the

thickening at the base of the Taiwan wedge is also discontinuous in comparison with the timescale of the gravity measurements. The elasto-plastic-viscous rheology adopted in model B is intended to reproduce tectonic processes prevailing at both short-term and long-term timescales. The displacement field (Fig. 4) shows velocities averaged over the last 200 kyr and therefore represents an integration of discontinuous short-term deformational events occurring at the scale of the earthquake cycle. Then the signals computed here would not be measurable as such, even over several decades.

Gravity change rates related to processes occurring at the scale of the seismic cycle of active faults in Taiwan are more likely to be measured, in particular in the case of earthquakes that generate sudden large displacements. Indeed, gravity is very sensitive to vertical ground motions. Considering the theoretical complete Bouguer reduction (Hofmann-Wellenhof & Moritz 2006), 1 cm of vertical ground uplift decreases gravity by $\sim$2 μGal. For instance, the 1999 M_W7.6 Chi-Chi earthquake might theoretically have produced a gravity decrease of 900 μGal at most in the northern part of the Chelungpu fault, as a consequence of the 4.5 m of coseismic surface uplift (Ma *et al.* 2001). More recently, the 2010 M_W6.4 Jiashian earthquake might have decreased gravity by -5 μGal as a result of 2.5 cm of uplift near the epicentre (Hsu *et al.* 2011). One part of this signal is due to the addition of masses below the measurement site, inducing the uplift. Only gravimetry is sensitive to these masses. In the Bouguer reduction, this effect is computed assuming a density of 2670 kg m^{-3}. The other part of the signal accounts for the movement of the measurement site away from the Earth's centre of mass (free-air effect). This movement can be observed and measured more precisely by repeated leveling or GPS. Knowing the coseismic gravity change allows to retrieve the density value, giving insights into coseismic mass redistributions (Barnes 1966; Tanaka *et al.* 2001). This can also be applied to postseismic deformation provided its amplitude is large enough (Panet *et al.* 2007). In any case these signals are significantly larger than any of the possible long-term signals computed here.

It should be reminded that we have neglected in our modelling any mass transfers above sea level. This suggests that our modelling only provides the gravity signal related to deep crustal mass transfers, but not to the response of topography to this tectonic forcing. If the Taiwan mountain belt can be considered at a topographic steady state over large temporal and spatial scales (Suppe 1981; Willett & Brandon 2002), this may not be the case at the local and therefore short-term temporal scales (Stolar *et al.* 2007). The topography adjusts to the long-term tectonic forcing through discrete and discontinuous erosional events at shorter timescales, essentially related to earthquakes and typhoons (Dadson *et al.* 2004; Hovius *et al.* 2011). Because these topographic adjustments occur at the surface and hence close to the measurement sites, they are expected to produce the largest gravity signal. This implies that if gravity is to be used to track deep tectonic processes, measurements should be considered over a time period longer than the response time of the landscape to tectonic forcing. Such response time may range from 0.25 to 2.5 Ma (Whipple 2001) and gravimetry is not suitable for this timescale. As a result, for long-term mountain building processes, gravimetry faces an issue related both to the weakness of the signal and to the comparatively short timescale of the measurements.

Because gravimetry is sensitive to any mass transfer, hydrology will also contribute to gravity changes, especially at seasonal timescales as suggested by Hwang *et al.* (2009). This signal does not accumulate with time but reaches $\sim$10 μGal at annual frequencies. Van Camp *et al.* (2010) show that the hydrological signal will average out after a period of time that depends on its magnitude.

They found that 3.5 to 17 yr are required to separate a tectonic trend of 1 $\times$ 10^{-1} μGal yr^{-1} using continuous time-series measured by superconducting gravimetres. As we deal with a tectonic trend five times smaller and gravity time-series sampled at lower frequency (yearly), this separation time might significantly increase. In the case of Taiwan, the shallow redistribution of rocks by landslides and debris-flow deposits may also significantly contribute to gravity changes at short timescales, provided that these events occur nearby the site where gravity is measured. Taiwan mountainous areas have steep slopes that experience landslides during earthquakes (Dadson *et al.* 2004; Lin *et al.* 2004; Hovius *et al.* 2011) and typhoons (Dadson *et al.* 2003). As an example, in August 2009 typhoon Morakot brought heavy rains that triggered more than 9000 landslides in Taiwan (Tsai *et al.* 2010). The largest one redistributed 26 $\times$ 10^6 m^3 of rocks (Tsou *et al.* 2011). Such volumes accounted for 60 $\pm$ 2 to 285 $\pm$ 3 μGal of gravity changes in some locations (Mouyen *et al.* 2013) and will eventually alter the identification of gravity change rates due to tectonics. On the other hand, gravity changes can be used to constrain the volumes of landslides and deposits and to investigate the evolution of the landscape at these timescales.

5 CONCLUSION

We have investigated whether long-term mountain building deep crustal processes could be detected by temporal gravimetry in Taiwan in the case that such processes have a short-term signature.

Rates of gravity change, produced by deep crustal mass transfers only, were computed for two different models of Taiwan with similar wedge geometries, but implying different processes at depth to sustain the growth of the orogenic wedge. In the case of the underplating model (Simoes *et al.* 2007a), we computed a gravity change rate of -6×10^{-2} to 2 $\times$ 10^{-2} μGal yr^{-1} at most in the Tananao Complex, depending on whether crustal subduction is significant or not, respectively. The input of deeper mass by underplating itself, at the interface between the orogenic wedge and the upper crust of the CCM, has comparatively no effect. In the crustal scale accretion model of Yamato *et al.* (2009), the large-scale folding of the CCM and viscous thickening generates a gravity increase of 7 $\times$ 10^{-2} μGal yr^{-1} at most in the eastern part of Taiwan. By comparison with the predictions of model A in the Hsuehshan Range, which is less than 1 $\times$ 10^{-2} μGal yr^{-1}, we propose that shallow wedge models do not produce any significant gravity change.

The extreme kinematics of mountain building in Taiwan originally suggested that this site could produce significant (and therefore observable) gravity changes over time. This is the reason why this site was used for testing the potential of gravimetry in investigating mountain building processes. However, the amplitude of the gravity changes computed from long-term kinematics of mountain building at depth remains very small in comparison with the uncertainty of gravimetres, especially because gravity contrasts within the orogenic wedge are low. Moreover, assuming gravimetres could provide more precise measurements, the difficulty to perfectly correct the times-series from extreme surface processes (hydrological variations, landslides) would still prevent from identifying any tectonic signal over a few years of measurements. Finally, it also appears that, due to their discontinuous occurrence in time, long-term signals may not exist as such over short time periods. We therefore conclude that temporal gravity is not suited for investigating deep crustal mass transfers sustaining long-term mountain-building processes.

REFERENCES

Barnes, D.F., 1966. Gravity changes during the Alaska earthquake, *J. geophys. Res.,* **71**(2), 451–456.

Barr, T.D. & Dahlen, F.A., 1989. Brittle frictional mountain building: 2. Thermal structure and heat budget, *J. geophys. Res.,* **94**(B4), 3923–3947.

Barr, T.D. & Dahlen, F.A., 1990. Constraints on friction and stress in the Taiwan fold-and-thrust belt from heat flow and geochronology, *Geology,* **18**, 111–115.

Barr, T.D., Dahlen, F. & McPhail, D.C., 1991. Brittle frictional mountain building: 3. Low-grade metamorphism, *J. geophys. Res.,* **96**(B6), 10 319–10 338.

Battaglia, M., Gottsmann, J., Carbone, D. & Fernández, J., 2008. 4D volcano gravimetry, *Geophysics,* **73**(06), WA3–WA18.

Beyssac, O., Simoes, M., Avouac, J.-P., Farley, K.A., Chen, Y.-G., Chan, Y.-C. & Goffé, B., 2007. Late Cenozoic metamorphic evolution and exhumation of Taiwan, *Tectonics,* **26**(6), TC6001, doi:10.1029/2006TC002064.

Beyssac, O., Negro, F., Simoes, M., Chan, Y.C. & Chen, Y.G., 2008. High-pressure metamorphism in Taiwan: from oceanic subduction to arc-continent collision?, *Terra Nova,* **20**(2), 118–125.

Bollinger, L., Henry, P. & Avouac, J.-P., 2006. Mountain building in the Nepal Himalaya: thermal and kinematic model, *Earth planet. Sci. Lett.,* **244**(1), 58–71.

Bousquet, R., Goffé, B., Henry, P., Le Pichon, X. & Chopin, C., 1997. Kinematic, thermal and petrological model of the Central Alps: Lepontine metamorphism in the upper crust and eclogitisation of the lower crust, *Tectonophysics,* **273**(1–2), 105–127.

Byrne, T.B. & Liu, C.S., 2002. Introduction to the geology and geophysics of Taiwan, *Geol. Soc. Am.,* **358**, 5–8.

Dadson, S.J. *et al.*, 2003. Links between erosion, runoff variability and seismicity in the Taiwan orogen, *Nature,* **426**(6967), 648–651.

Dadson, S.J. *et al.*, 2004. Earthquake-triggered increase in sediment delivery from an active mountain belt, *Geology,* **32**(8), 733–736.

Dahlen, F.A. & Barr, T.D., 1989. Brittle frictional mountain building: 1. Deformation and mechanical energy budget, *J. geophys. Res.,* **94**(B4), 3906–3922.

Dahlen, F.A., Suppe, J. & Davis, D., 1984. Mechanics of fold-and-thrust belts and accretionary wedges: cohesive Coulomb theory, *J. geophys. Res.,* **89**(B12), 10 087–10 101.

Davis, D., Suppe, J. & Dahlen, F.A., 1983. Mechanics of fold-and-thrust belts and accretionary wedges, *J. geophys. Res.,* **88**(B2), 1153–1172.

De Capitani, C., 1994. Gleichgewichts-phasendiagramme: theorie und software, Jahrestagung der Deutsch. Mineral. Ges., *Beihefte zum Eur. J. Mineral.,* **6**, 1–48.

Dunlap, W.J., Hirth, G. & Teyssier, C., 1997. Thermomechanical evolution of a ductile duplex, *Tectonics,* **16**(6), 983–1000.

Fuller, C.W., Willett, S.D., Fisher, D. & Lu, C.Y., 2006. A thermomechanical wedge model of Taiwan constrained by fission-track thermochronometry, *Tectonophysics,* **425**(1–4), 1–24.

Henry, P., Le Pichon, X. & Goffé, B., 1997. Kinematic, thermal and petrological model of the Himalayas: constraints related to metamorphism within the underthrust Indian crust and topographic elevation, *Tectonophysics,* **273**(1–2), 31–56.

Hetényi, G., Cattin, R., Brunet, F., Bollinger, L., Vergne, J., Nábelek, J.L. & Diament, M., 2007. Density distribution of the India plate beneath the Tibetan plateau: geophysical and petrological constraints on the kinetics of lower-crustal eclogitization, *Earth planet. Sci. Lett.,* **264**(1–2), 226–244.

Ho, C.S., 1986. A synthesis of the geologic evolution of Taiwan, *Tectonophysics,* **125**(1–3), 1–16.

Hofmann-Wellenhof, B. & Moritz, H., 2006. *Physical Geodesy,* Springer-Verlag.

Hovius, N., Meunier, P., Lin, C.W., Chen, H., Chen, Y.G., Dadson, S., Horng, M.J. & Lines, M., 2011. Prolonged seismically induced erosion and the mass balance of a large earthquake, *Earth planet. Sci. Lett.,* **304**(3), 347–355.

Hsu, Y.J., Simons, M., Yu, S.B., Kuo, L.C. & Chen, H.Y., 2003. A two-dimensional dislocation model for interseismic deformation of the Taiwan mountain belt, *Earth planet. Sci. Lett.,* **211**(3–4), 287–294.

Hsu, Y.J., Yu, S.B., Kuo, L.C., Tsai, Y.C. & Chen, H.Y., 2011. Coseismic deformation of the 2010 Jiashian, Taiwan earthquake and implications for fault activities in southwestern Taiwan, *Tectonophysics,* **502**, 328–335.

Hu, J.C., Yu, S.B., Angelier, J. & Chu, H.T., 2001. Active deformation of Taiwan from GPS measurements and numerical simulations, *J. geophys. Res.,* **106**(B2), 2265–2280.

Hwang, W.T. & Wang, C., 1993. Sequential thrusting model for mountain building: constraints from geology and heat flow of Taiwan, *J. geophys. Res.,* **98**(B6), 9963–9973.

Hwang, C., Kao, R., Cheng, C.-C., Huang, J.F., Lee, C.W. & Sato, T., 2009. Results from parallel observations of superconducting and absolute gravimeters and GPS at the Hsinchu station of Global Geodynamics Project, Taiwan, *J. geophys. Res.,* **114**(B7), B07406, doi:10.1029/2008JB006195.

Jacob, T., Bayer, R., Chery, J. & Le Moigne, N., 2010. Time-lapse microgravity surveys reveal water storage heterogeneity of a karst aquifer, *J. geophys. Res.,* **115**(B6), B06402, doi:10.1029/2009JB006616.

Kim, K.H., Chiu, J.M., Pujol, J., Chen, K.C., Huang, B.S., Yeh, Y.H. & Shen, P., 2005. Three-dimensional VP and VS structural models associated with the active subduction and collision tectonics in the Taiwan region, *Geophys. J. Int.,* **162**(1), 204–220.

Konstantinovskaia, E. & Malavieille, J., 2005. Erosion and exhumation in accretionary orogens: experimental and geological approaches, *Geochem. Geophys. Geosyst.,* **6**(2), doi:10.1029/2004GC000794.

Lee, C.R. & Cheng, W.T., 1986. Preliminary heat flow measurements in Taiwan, in *Proceedings of the Fourth Circum-Pacific Energy and Mineral Resources Conference,* Singapore, pp. 1–9.

Lester, R., McIntosh, K., Van Avendonk, H. J.A., Lavier, L., Liu, C.S. & Wang, T.K., 2013. Crustal accretion in the Manila trench accretionary wedge at the transition from subduction to mountain-building in Taiwan, *Earth planet. Sci. Lett.,* **375**, 430–440.

Lin, A.T. & Watts, A.B., 2002. Origin of the West Taiwan basin by orogenic loading and flexure of a rifted continental margin, *J. geophys. Res.,* **107**, 2185, doi:10.1029/2001JB000669.

Lin, C.-H., 2005. Identification of mantle reflections from a dense linear seismic array: tectonic implications to the Taiwan orogeny, *Geophys. Res. Lett,* **32**, L06315, doi:10.1029/2004GL021814.

Lin, A.T., Watts, A.B. & Hesselbo, S.P., 2003. Cenozoic stratigraphy and subsidence history of the South China Sea margin in the Taiwan region, *Basin Res.,* **15**(4), 453–478.

Lin, C.W., Shieh, C.L., Yuan, B.D., Shieh, Y.C., Liu, S.H. & Lee, S.Y., 2004. Impact of Chi-Chi earthquake on the occurrence of landslides and debris flows: example from the Chenyulan River watershed, Nantou, Taiwan, *Eng. Geol.,* **71**(1), 49–61.

Liu, T.K., Hsieh, S., Chen, Y.G. & Chen, W.S., 2001. Thermo-kinematic evolution of the Taiwan oblique-collision mountain belt as revealed by zircon fission track dating, *Earth planet. Sci. Lett.,* **186**(1), 45–56.

Lo, C.H. & Onstott, T.C., 1995. Rejuvenation of KAr systems for minerals in the Taiwan Mountain Belt, *Earth planet. Sci. Lett.,* **131**(1–2), 71–98.

Ma, K.F., Mori, J., Lee, S.J. & Yu, S.B., 2001. Spatial and temporal distribution of slip for the 1999 Chi-Chi, Taiwan, earthquake, *Bull. Seism. Soc. Am.,* **91**(5), 1069–1087.

McIntosh, K., van Avendonk, H., Lavier, L., Lester, W.R., Eakin, D., Wu, F., Liu, C.-S. & Lee, C.-S., 2013. Inversion of a hyper-extended rifted margin in the southern Central Range of Taiwan, *Geology,* **41**(8), 871–874.

Masson, F. *et al.*, 2008. Study of the Taiwanese orogen from absolute gravity data, *AGU Fall Meeting,* abstract #T31C-2013.

Mouthereau, F. & Lacombe, O., 2006. Inversion of the Paleogene Chinese continental margin and thick-skinned deformation in the Western Foreland of Taiwan, *J. Struct. Geol.,* **28**(11), 1977–1993.

Mouthereau, F. & Petit, C., 2003. Rheology and strength of the Eurasian continental lithosphere in the foreland of the Taiwan collision belt: constraints from seismicity, flexure, and structural styles, *J. geophys. Res.,* **108**(B11), 2512, doi:10.1029/2002JB002098.

Mouthereau, F., Watts, A.B. & Burov, E., 2013. Structure of orogenic belts controlled by lithosphere age, *Nat. Geosci.,* **6**(9), 785–789.

Mouyen, M. *et al.*, 2009. Expected temporal absolute gravity change across the Taiwanese Orogen, a modeling approach, *J. Geodyn.,* **48,** 284–291.

Mouyen, M. *et al.*, 2013. Erosion effects assessed by repeated gravity measurements in southern Taiwan, *Geophys. J. Int.,* **192**(1), 113–136.

Niebauer, T.M., Sasagawa, G.S., Faller, J.E., Hilt, R. & Klopping, F., 1995. A new generation of absolute gravimeters, *Metrologia,* **32,** 159–180.

Panet, I. *et al.*, 2007. Coseismic and post-seismic signatures of the Sumatra 2004 December and 2005 March earthquakes in GRACE satellite gravity, *Geophys. J. Int.,* **171**(1), 177–190.

Peacock, S.M., 1991. Numerical simulation of subduction zone pressure-temperature-time paths: constraints on fluid production and arc magmatism, *Phil. Trans. R. Soc. Lond., A.: Phys. Eng. Sci.,* **335**(1638), 341–353.

Shih, R.C., Lin, C.H., Lai, H.L., Yeh, Y.H., Huang, B.S. & Yen, H.Y., 1998. Preliminary crustal structures across central Taiwan from modeling of the onshore-offshore wide-angle seismic data, *Terr. Atmos. Oceanic Sci.,* **9,** 317–328.

Shyu, J. B.H., Sieh, K., Chen, Y.G. & Liu, C.S., 2005. Neotectonic architecture of Taiwan and its implications for future large earthquakes, *J. geophys. Res.,* **110**(B8), B08402, doi:10.1029/2004JB003251.

Sibuet, J.C. & Hsu, S.K., 2004. How was Taiwan created?, *Tectonophysics,* **379**(1–4), 159–181.

Simoes, M. & Avouac, J.P., 2006. Investigating the kinematics of mountain building in Taiwan from the spatiotemporal evolution of the foreland basin and western foothills, *J. geophys. Res.,* **111**(B10), B10401, doi:10.1029/2005JB004209.

Simoes, M., Avouac, J.P., Beyssac, O., Goffe, B., Farley, K.A. & Chen, Y.G., 2007a. Mountain building in Taiwan: a thermokinematic model, *J. geophys. Res.,* **112,** B11405, doi:10.1029/2006JB004824.

Simoes, M., Avouac, J.P. & Chen, Y.G., 2007b. Slip rates on the Chelungpu and Chushiang thrust faults inferred from a deformed strath terrace along the Dungpuna river, west central Taiwan, *J. geophys. Res.,* **112,** B03S10, doi:10.1029/2005JB004200.

Simoes, M. *et al.*, 2007c. Kinematic analysis of the Pakuashan fault tip fold, west central Taiwan: shortening rate and age of folding inception, *J. geophys. Res.,* **112,** B03S14, doi:10.1029/2005JB004198.

Simoes, M., Beyssac, O. & Chen, Y.G., 2012. Late Cenozoic metamorphism and mountain building in Taiwan: a review, *J. Asian Earth Sci.,* **46,** 92–119.

Sobolev, S.V. & Babeyko, A.Y., 1989. *Phase Transformations in the Lower Continental Crust and Its Seismic Structure,* pp. 311–320, eds Mereu, R.F., Mueller, S. & Fountain, D.M., Geophysical Monograph 51, IUGG 6.

Stolar, D.B., Willett, S.D. & Montgomery, D.R., 2007. Characterization of topographic steady state in Taiwan, *Earth planet. Sci. Letters,* **261**(3), 421–431.

Sun, W., Wang, Q., Li, H., Wang, Y., Okubo, S., Shao, D., Liu, D. & Fu, G., 2009. Gravity and GPS measurements reveal mass loss beneath the Tibetan Plateau: Geodetic evidence of increasing crustal thickness, *Geophys. Res. Lett.,* **36(2),** L02303, doi:10.1029/2008GL036512.

Suppe, J., 1981. Mechanics of mountain building and metamorphism in Taiwan, *Mem. Geol. Soc. China,* **4,** 67–89.

Suppe, J., 1984. Kinematics of arc-continent collision, flipping of subduction, and back-arc spreading near Taiwan, *Mem. Geol. Soc. China,* **6,** 21–33.

Tanaka, Y., Okubo, S., Machida, M., Kimura, I. & Kosuge, T., 2001. First detection of absolute gravity change caused by earthquake, *Geophys. Res. Lett.,* **28**(15), 2979–2981.

Tensi, J., Mouthereau, F. & Lacombe, O., 2006. Lithospheric bulge in the west Taiwan basin, *Basin Res.,* **18**(3), 277–299.

Tsai, F., Hwang, J.H., Chen, L.C. & Lin, T.H., 2010. Post-disaster assessment of landslides in southern Taiwan after 2009 Typhoon Morakot using remote sensing and spatial analysis, *Nat. Hazards Earth Syst. Sci.,* **10,** 2179–2190.

Tsao, S.J., 1996. The geological significance of illite crystallinity, zircon fission-track ages and K-Ar ages of metasedimentary rocks of the Central Range, *PhD thesis,* Natl. Taiwan Univ., Taipei, Taiwan, p. 272.

Tsou, C.Y., Feng, Z.Y. & Chigira, M., 2011. Catastrophic landslide induced by Typhoon Morakot, Shiaolin, Taiwan, *Geomorphology,* **127,** 166–178.

Van Camp, M., Williams, S.D.P. & Francis, O., 2005. Uncertainty of absolute gravity measurements, *J. geophys. Res.,* **110,** B05406, doi:10.1029/2004JB003497.

Van Camp, M., Métivier, L., de Viron, O., Meurers, B. & Williams, S.D.P., 2010. Characterizing long-time scale hydrological effects on gravity for improved distinction of tectonic signals, *J. geophys. Res.,* **115**(B7), B07407, doi:10.1029/2009JB006615.

Wessel, P. & Smith, W.H.F., 1991. Free software helps map and display data, *EOS, Trans. Am. geophys. Un.,* **72,** 445–446.

Whipple, K.X., 2001. Fluvial landscape response time: how plausible is steady-state denudation?, *Am. J. Sci.,* **301**(4–5), 313–325.

Willett, S.D. & Brandon, M.T., 2002. On steady states in mountain belts, *Geology,* **30**(2), 175–178.

Willett, S.D., Fisher, D., Fuller, C., Yeh, E.-C. & Lu, C.Y., 2003. Erosion rates and orogenic-wedge kinematics in Taiwan inferred from fission-track thermochronometry, *Geology,* **31**(11), 945–948.

Yamato, P., Mouthereau, F. & Burov, E., 2009. Taiwan mountain building: insights from 2-D thermomechanical modelling of a rheologically stratified lithosphere, *Geophys. J. Int.,* **176**(1), 307–326.

Yeh, Y.H. *et al.*, 1998. Onshore/offshore wide-angle deep seismic profiling in Taiwan, *Terr. Atmos. Oceanic Sci.,* **9**(3), 301–316.

Yen, H.Y., Yeh, Y.H. & Wu, F.T., 1998. Two-dimensional crustal structures of Taiwan from gravity data, *Tectonics,* **17,** 104–111.

Yu, S.B., Chen, H.Y. & Kuo, L.C., 1997. Velocity field of GPS stations in the Taiwan area, *Tectonophysics,* **274**(1–3), 41–59.

Yu, S.B., Kuo, L.C., Punongbayan, R.S. & Ramos, E.G., 1999. GPS observation of crustal deformation in the Taiwan-Luzon, *Geophys. Res. Lett.,* **26**(7), 923–926.