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Earthquake-induced gravitational potential energy change in the active Taiwan orogenic belt

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SUMMARY

The Philippine Sea Plate is converging against the Eurasian Plate near Taiwan at a velocity of 7–8 cm yr^{-1}; this has caused the Taiwan orogenesis and induced abundant earthquakes. In this study we examine the corresponding change of gravitational potential energy (ΔGPE) using 757 earthquakes from the earthquake catalogue of the Broadband Array in Taiwan for Seismology (BATS) from 1995 July to 2003 December. Our results show that the variation of the crustal ΔGPE strongly correlates with the different stages of the orogenesis. Except for the western Okinawa Trough and southern Taiwan, most of the Taiwan convergent region exhibits a gain of crustal ΔGPE. In contrast, the lithospheric ΔGPE in the Taiwan region exhibits a reverse pattern. For the whole Taiwan region, the earthquake-induced crustal ΔGPE and the lithospheric ΔGPE during the observation period are 1.03×10^{17} J and -1.15×10^{17} J, respectively. The average rate of the whole ΔGPE in the Taiwan region is very intense and equal to -2.07×10^{10} W, corresponding to about 1 per cent of the global GPE loss induced by earthquakes.

Key words: convergent plate boundary, earthquake, gravitational potential energy, Taiwan orogen.

INTRODUCTION

Earthquakes not only release seismic wave energy but also generate other types of energy such as rotational kinetic energy, gravitational potential energy (GPE) and elastic strain energy (Dahlen 1977). Of those, the kinetic energy of rotation is the smallest energy induced by earthquakes (Dahlen 1977). The energy released by seismic wave energy due to earthquake faulting must be dissipated into the Earth somewhere. The change in gravitational potential energy (ΔGPE) is because of the mass redistribution of the Earth due to earthquakes and is generally of a magnitude hundreds of times larger than the released seismic wave energy (Chao *et al.* 1995). Tanimoto & Okamoto (2000) and Tanimoto *et al.* (2002) have suggested that the crustal portion of the GPE change is a good tectonic indicator; compressional tectonic regions display a gain of crustal ΔGPE while extensional tectonic regions display a loss of crustal ΔGPE.

The Philippine Sea Plate is moving northwestwards with respect to the Eurasian Plate and is converging towards the Asian continental margin at a velocity of 7–8 cm yr^{-1} near the Taiwan region (Yu *et al.* 1997) (Fig. 1). The active collision between the two plates has caused the uplift of the Taiwanese mountain belt and generated numerous earthquakes. Off northeast Taiwan, the Philippine Sea Plate is subducting northwards beneath the Ryukyu Arc. Behind the Ryukyu Arc, the Okinawa Trough backarc basin is actively rifting and has propagated westwards into northeast Taiwan (Liu 1995). The intense rifting of the southernmost Okinawa Trough is considered to be a consequence of the post-collisional extension of the Taiwan orogenesis (Wang *et al.* 1999) (Fig. 1). On the other hand, in the south of Taiwan, the Luzon Arc is initially converging toward the Asian continental margin, considered to be an initial stage of the Taiwan orogenesis (Fig. 1). Thus, the region of the Taiwan Plate boundary provides a rare opportunity to observe the variation of the earthquake-induced ΔGPE for a complete set of orogenic processes.

EARTHQUAKE DATA

The earthquake data used in this study come from the Broadband Array in Taiwan for Seismology (BATS) (http://bats.earth.sinica.edu.tw/). The data set contains 757 major earthquakes and provides seismic moment solutions from 1995 July to 2003 December (Fig. 2). Fig. 3 shows the statistics of the earthquake source depths and magnitudes. The range of earthquake magnitudes is from $M_w = 3.12$ to $M_w = 7.6$ (Fig. 3a). The earthquake source depths range from 8 km to 180 km; however, most of the earthquakes are shallower than 30 km (Fig. 3b).

CALCULATION OF ΔGPE

ΔGPE can be calculated by

$$\Delta\text{GPE} = -\int_V \rho u g \, dV, \tag{1}$$

where ρ is the density distribution of the Earth, u is the radial static displacement caused by earthquakes, g is the gravity

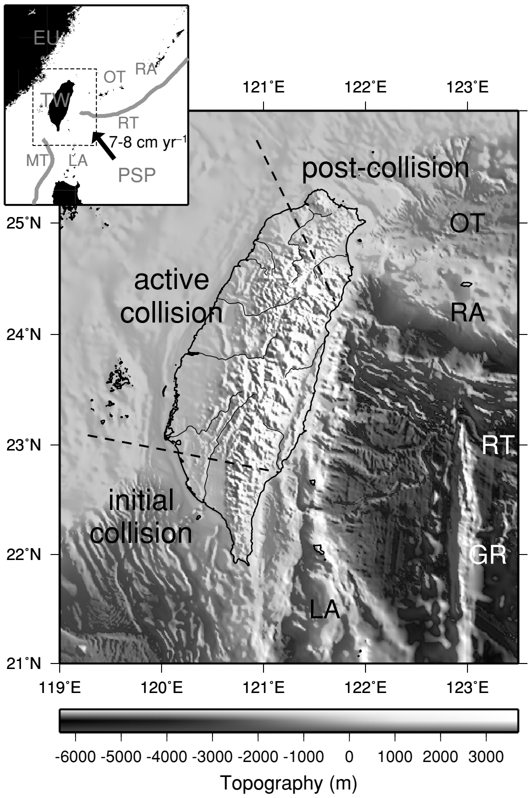

Figure 1. The tectonic setting of the Taiwan area. The Philippine Sea Plate has moved northwestwards relative to the Eurasian Plate with a velocity of 7–8 cm yr^{-1}: EU, Eurasian Plate; GR, Gagua Ridge; LA, Luzon Arc; MT, Manila Trench; OT, Okinawa Trough; PSP, Philippine Sea Plate; RA, Ryukyu Arc; RT, Ryukyu Trench; TW, Taiwan.

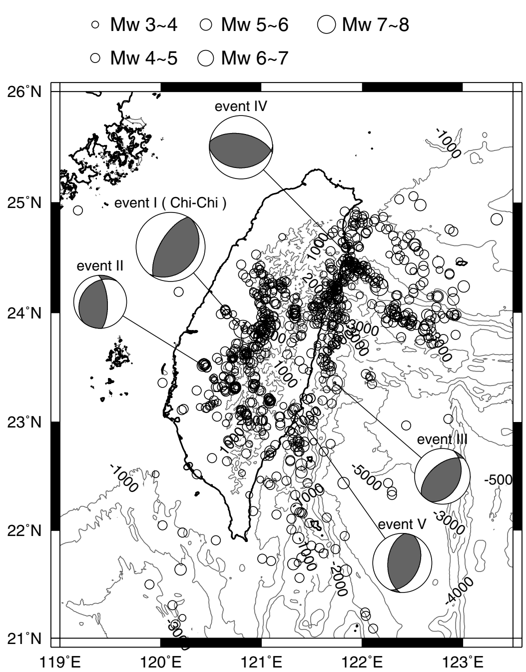

Figure 2. The distribution of the earthquakes used in the study area from 1995 July to 2003 December. Five major earthquake focal mechanisms contributing large ΔGPE are plotted (Table 1).

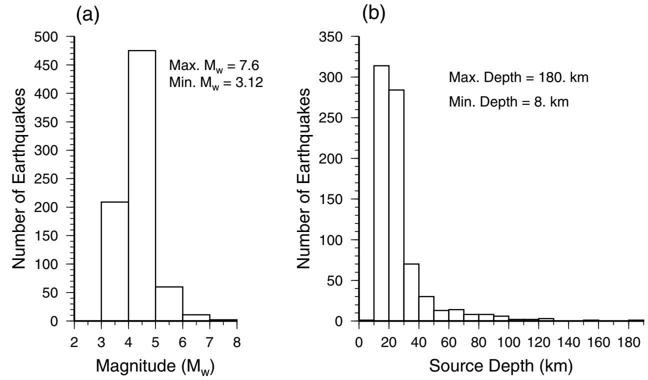

Figure 3. The statistical histograms of 757 earthquakes: (a) the distribution of the earthquake magnitudes; (b) the distribution of the earthquake depths.

acceleration and V is the calculated volume. For the spherical, symmetric, non-rotational, elastic and isotropic (SNREI) consideration of the Earth model and under the assumption of seismic moment tensor $M_{rr} + M_{\theta\theta} + M_{\phi\phi} = 0$, eq. (1) is reduced to

$$\Delta \mathrm{GPE} = M_{rr} \int_V K(r; r_s)\, dr, \tag{2}$$

where M_{rr} is the radial component of the seismic moment tensor and the $K(r; r_s)$ is so-called depth kernel (Okamoto & Tanimoto 2002). $K(r; r_s)$ is defined as

$$K(r; r_s) = 4\pi r^2 \rho(r) u_r(r; r_s) g(r), \tag{3}$$

r is the distance from the centre of the Earth while an earthquake occurs at r_s. The advantage of using this approach of direct numerical solution of ΔGPE is to avoid the truncated error of a harmonic function by using a normal mode solution (Okamoto & Tanimoto 2002).

As a first-order approximation, we adopt the parameters in the Preliminary Reference Earth Model (PREM) (Dziewonski & Anderson 1981) to use in eq. (1). However, the surface water layer is replaced by upper crustal material. The depth kernel of eq. (3) is demonstrated in Fig. 4 for shallow and deep earthquakes. The

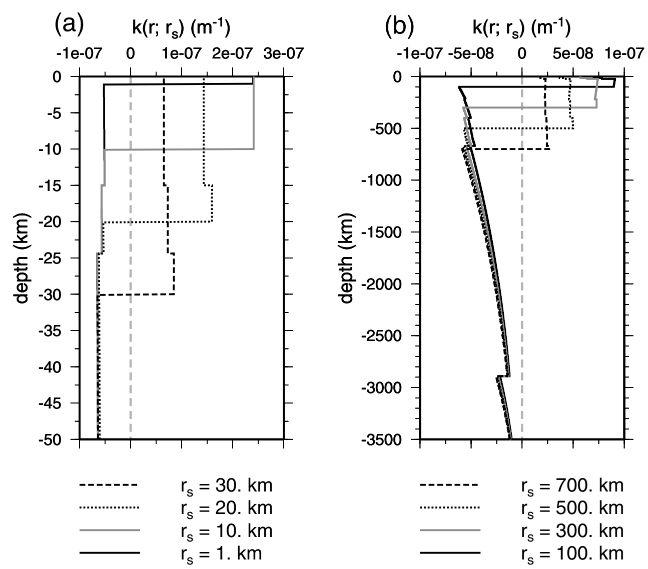

Figure 4. (a) Example of the depth kernel ($M_{rr} = 1$) values at different shallow earthquake source depths. (b) Example of the depth kernel values at different deep earthquake source depths.

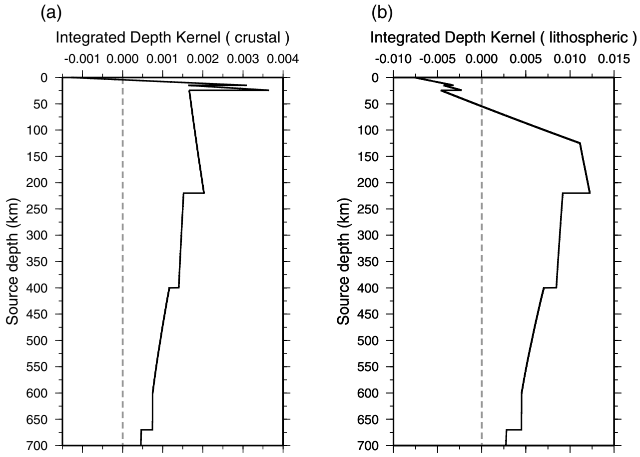

Figure 5. The distributions of the integrated depth kernel for the crustal portion (left) and the lithospheric portion (right).

integrated kernels are shown in Figs 5(a) and (b) for the crustal (24.4 km thick) and lithospheric (125 km thick) portions, respectively. It is noted that the neutral depths in the integrated depth kernel for the crustal and lithospheric portions are 4.39 and 54.82 km, respectively (Fig. 5).

SPATIAL AND TEMPORAL DISTRIBUTION OF THE ΔGPE

To have a general understanding of the ΔGPE distribution we divide the study region into $0.2^\circ \times 0.2^\circ$ grid cells. All the crustal ΔGPE inside a cell is summed. The distribution of the crustal ΔGPE in different years from mid-1995 to 2003 is shown in Fig. 6. It can be seen that prior to the 1999 Chi-Chi earthquake (Ma *et al.* 1999; Kao & Chen 2000) the crustal ΔGPE in central Taiwan was rather quiet (Fig. 6). In the Taiwan region areas of both crustal GPE gain and crustal GPE loss exist (Fig. 6). This situation is also revealed by the temporal variation of the positive and negative components of the radial seismic moments (Fig. 7a). However, temporally the overall crustal GPE has been increasing (Fig. 7b). In contrast, the lithospheric GPE in the Taiwan region has been decreasing (Fig. 7b). It is worth noting that the important jumps of ΔGPE were generated by some major earthquakes (Fig. 7b) (Table 1). Eventually, the crustal GPE gained about 1.03×10^{17} J and the lithospheric GPE lost about 1.15×10^{17} J (Table 2) (Fig. 7b).

REGIONAL TECTONICS AND ΔGPE

Because the complete process of orogenesis can be observed in the Taiwan region, we can examine the ΔGPE relationship during different orogenic stages. We separate the Taiwan orogen into eight regions, including the initial collision (plate convergence) in south Taiwan, the active collision in central Taiwan and the post-collisional extension in the Okinawa Trough (Figs 8–11).

Region A (Luzon Arc convergent zone I)

In the southeast of Taiwan, the Luzon Arc initially converges with Taiwan Island; the Eurasian Plate subducts eastwards beneath the Luzon Arc. This initial collision has caused frequent seismicity along the Luzon Arc (Fig. 2). The associated crustal ΔGPE has been increasing but is rather small except for the change created by the 2003 December 10 earthquake (event V in Fig. 2 and Table 1) (Fig. 9a). It is interesting to notice that prior to the earthquake event V, the accumulation of the crustal ΔGPE in region A was dramatically smooth for at least 4 yr (Fig. 9a).

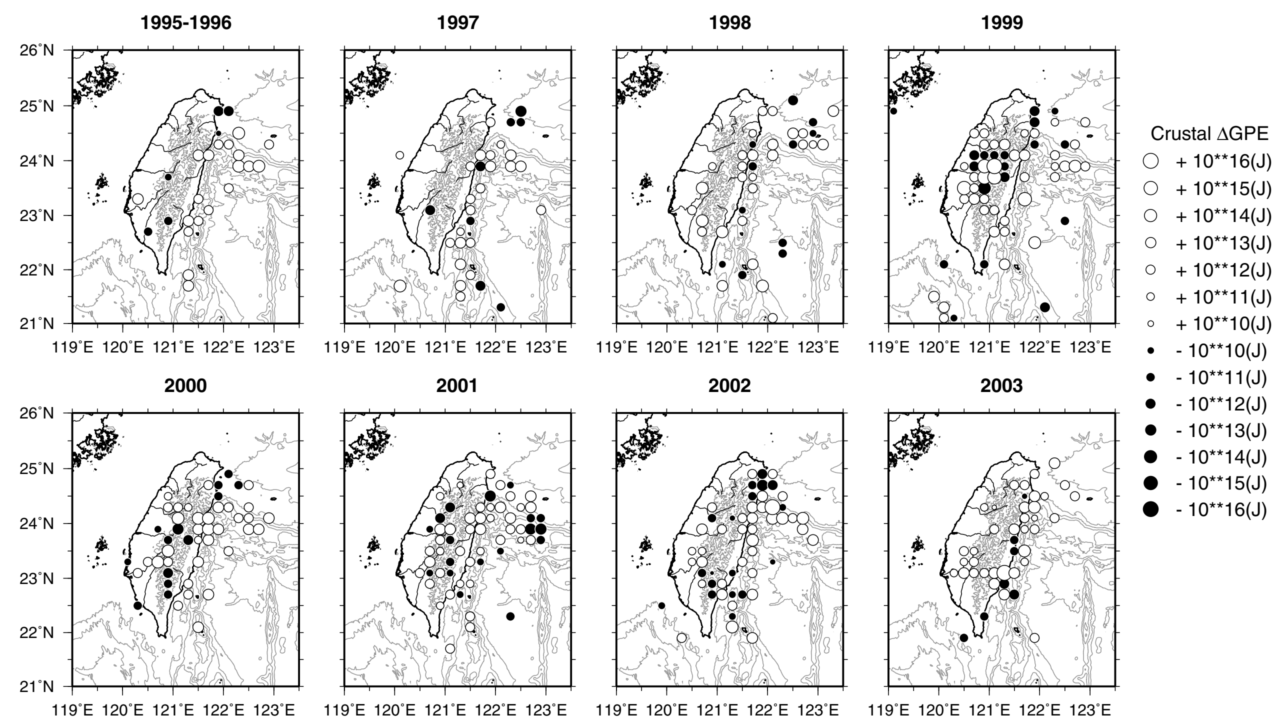

Figure 6. The distribution of the crustal ΔGPE in the Taiwan region from mid-1995 to 2003. The topography with a contour interval of 1 km is plotted for reference.

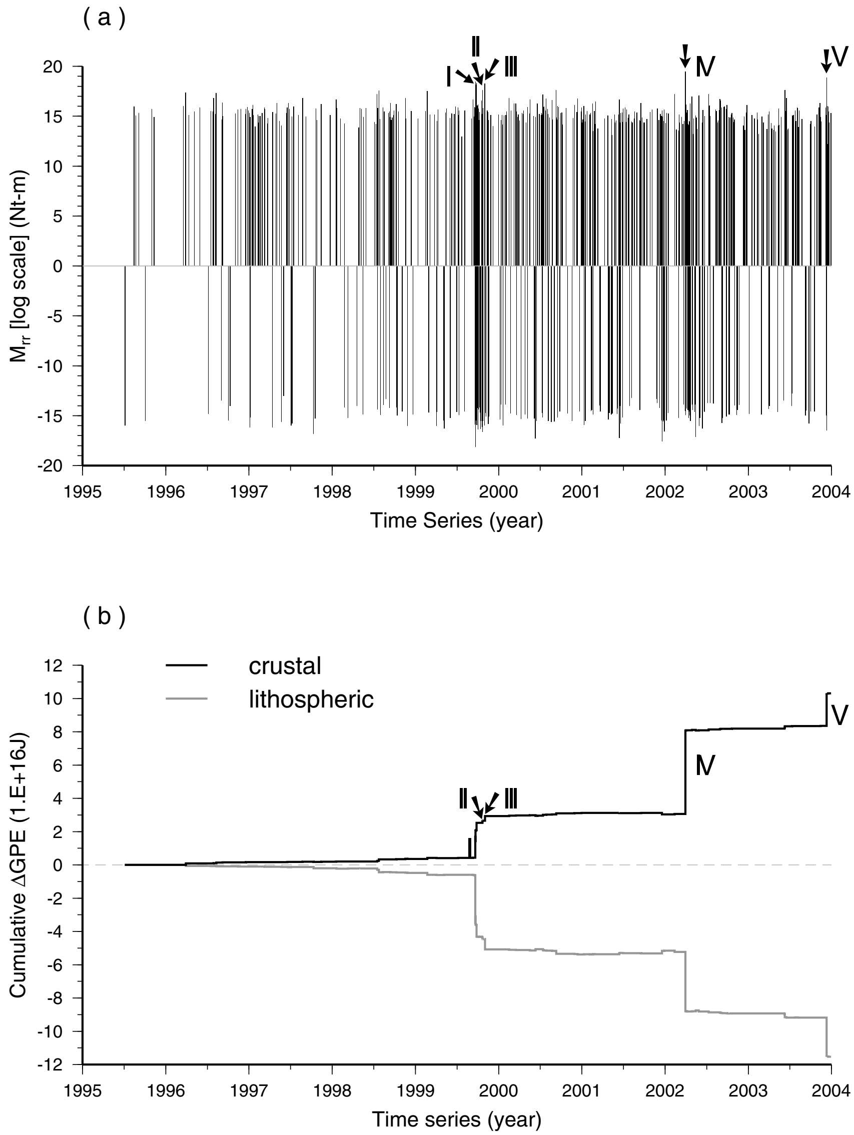

Figure 7. (a) The temporal distribution of the radial component (M_{rr}) of the seismic moment solutions used. (b) The cumulative crustal and lithospheric ΔGPE as a function of time. The marked numbers indicate large contributions from the earthquake events listed in Table 1.

Table 1. Earthquakes having large crustal ΔGPE contributions in the Taiwan region.

Events	Time	Location	Source depth (km)	M_w	M_{rr} $(N-m)$	Crustal ΔGPE (J)
I	1999 Sep. 20	120.82°E, 23.85°N	8	7.6	0.267E+19	0.276E+16
II	1999 Oct. 22	120.42°E, 23.52°N	21	5.9	0.440E+18	0.129E+16
III	1999 Nov. 01	121.73°E, 23.36°N	31	6.1	0.172E+19	0.288E+16
IV	2002 Mar. 31	122.17°E, 24.24°N	47	7.0	0.295E+20	0.502E+17
V	2003 Dec. 10	121.34°E, 23.10°N	20	6.5	0.724E+19	0.196E+17

Region B (onshore/offshore south Taiwan)

With respect to region A, region B is converging eastwards and downwards beneath the Luzon Arc and could be considered as the forearc region of region A (Fig. 8). As with region A, region B is in the initial stage of orogenesis but has displayed a loss of the crustal GPE (Fig. 9b). The crustal GPE loss in this region is probably due to the normal faulting earthquakes and is ascribed to the bending of the subducting lithosphere. In contrast to region A, the lithospheric GPE has been increasing in region B. However, it can be seen that on either the crustal or lithospheric scale ΔGPE is rather small (Fig. 9b).

Region C (Luzon Arc convergent zone II)

Located in the north of region A, region C is the northernmost Luzon Arc that is actively entering into collision in the Taiwan orogen (Fig. 8) (Hsu & Sibuet 1995; Shyu *et al.* 1996). Compared with the initially convergent zone of the Luzon Arc (region A), the increase

Table 2. Average ΔGPE for different regions and different layer thicknesses (unit J).

	$R_b = 24.4$ km (crust)	$R_b = 125.0$ km (lithosphere)	$R_b = 6371.0$ km (Earth radius)
Region A	2.13E+16	−2.60E+16	−8.23E+17
Region B	−1.93E+13	1.71E+13	1.14E+15
Region C	5.57E+15	−1.18E+16	−3.42E+17
Region D	2.24E+15	−2.67E+15	−8.50E+16
Region E	2.16E+16	−3.79E+16	−1.04E+18
Region F	1.36E+14	−1.51E+14	−6.49E+15
Region G	5.35E+16	−3.79E+16	−3.30E+18
Region H	−5.92E+14	1.28E+13	2.49E+16
Whole region	1.03E+17	−1.15E+17	−5.54E+18

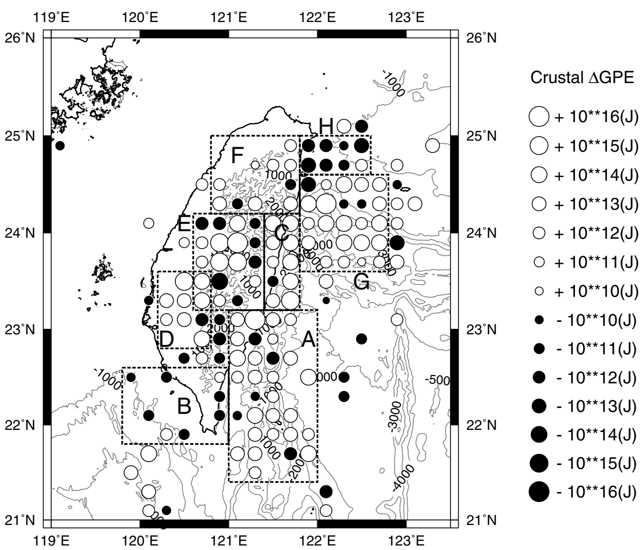

Figure 8. The summation of all the crustal ΔGPE in each 0.2° × 0.2° cell from mid-1995 to 2003.

of the crustal GPE or the decrease of the lithospheric GPE happens bit by bit (*cf.* Figs 9a and c). This may imply that the crust in region C is not as strong as in region A.

Region D (southwest Taiwan)

Regions D, E and F represent the Taiwan mountain belt (Fig. 8). From region B to region D, the tectonic stress has gradually changed from tension to compression (Yeh *et al.* 1991). In contrast to region B, the crustal ΔGPE in region D has become positive and the lithospheric ΔGPE is negative (*cf.* Figs 9b and d). It is worth noting that in this region the crustal ΔGPE curve as a function of time displays two tendencies; before about year 2000 (or the earthquake event II) the gain of crustal GPE is gradual but afterwards ΔGPE is almost unchanging. This suggests that the accumulation of the regional stress not released by earthquakes is increasing. This deserves attention as indicating an earthquake hazard in southwest Taiwan.

Region E (central Taiwan)

This region is directly located to the west of the actively convergent Luzon Arc (region C) (Fig. 8). Under strong horizontal compression, this region has the highest topography in Taiwan. As shown in Fig. 9(e), prior to the 1999 Chi-Chi earthquake (event I) the crustal ΔGPE was rather small. The gain of crustal GPE in this region is mainly due to the Chi-Chi earthquake sequence (Hsu & Lo 2004) (Fig. 6). For the observation period, the earthquake-induced ΔGPE in this region exhibits a crustal GPE gain in central Taiwan and a crustal GPE loss around the GPE gain area (Hsu & Lo 2004) (Fig. 8).

Region F (northern Taiwan)

This region represents a transition zone between the active collision zone in central Taiwan (region E) and the post-collisional extension zone of the Okinawa Trough (region H) (Fig. 8) (Hsu *et al.* 1996; Hu *et al.* 2002). The seismicity in this region is relatively quiet (Fig. 2). The crustal ΔGPE is still positive and the lithospheric ΔGPE is negative; however, their magnitudes are very small (Fig. 9f).

Region G (westernmost Ryukyu Arc)

The Philippine Sea Plate subducts northwards beneath the Ryukyu Arc, which has generated intensive subduction earthquakes (Kao *et al.* 1998). Because the thrust faulting earthquakes are predominant in this region, this region generally shows positive crustal ΔGPE (Fig. 8). However, in 2001 some negative crustal ΔGPE appeared in the northward prolongation of the Gagua Ridge (Fig. 6) and could be related to the subduction effect of the Gagua Ridge (Fig. 1).

Region H (westernmost Okinawa Trough)

Region H is a backarc basin of the Ryukyu subduction zone (Fig. 1). This region had probably suffered the collision of the Luzon Arc and is now under post-collisional extension (Hsu *et al.* 1996; Sibuet *et al.* 1998; Wang *et al.* 1999; Sibuet & Hsu 2004). Although the magnitude of ΔGPE is very small, this region generally demonstrates decreasing crustal GPE (Figs 8 and 9h). The total lithospheric ΔGPE is slightly positive but the temporal lithospheric ΔGPE is fluctuating (Fig. 9h). The strong fluctuation phenomenon is probably due to the coexistence of two types of earthquake at different depths in this region: shallow, rifting earthquakes and deep subduction earthquakes.

RELATIONSHIPS OF THE ΔGPE IN SUBDUCTION ZONES

Because our study area is situated between the Ryukyu subduction zone in the northeast of Taiwan and the Manila subduction zone in the south of Taiwan our results can provide information about the variation of the crustal GPE and the lithospheric GPE across the subduction zones. In the south, regions A and B combine to make a pair of ΔGPE observations (Fig. 8). The Luzon Arc region (region A) displays an increasing trend of crustal ΔGPE while the forearc region (region B) displays a decreasing trend (Figs 9a and b). The average crustal ΔGPE in region A is positive but in region B it is negative (Table 2). It is noted that the average magnitude of the crustal ΔGPE in region A is about three orders larger than in region B. The positive crustal ΔGPE of the Luzon Arc area is probably due to the subduction and thrust faulting earthquakes which make the overriding plate uplift, while the negative crustal ΔGPE of the forearc region can be ascribed to the normal faulting earthquakes due to the bending of the subducting plate (Fig. 8).

In contrast to the Manila subduction zone, the Ryukyu subduction zone has a well-developed backarc basin system (Sibuet *et al.* 1987, 1998). The northwestward motion of the northern tip of the

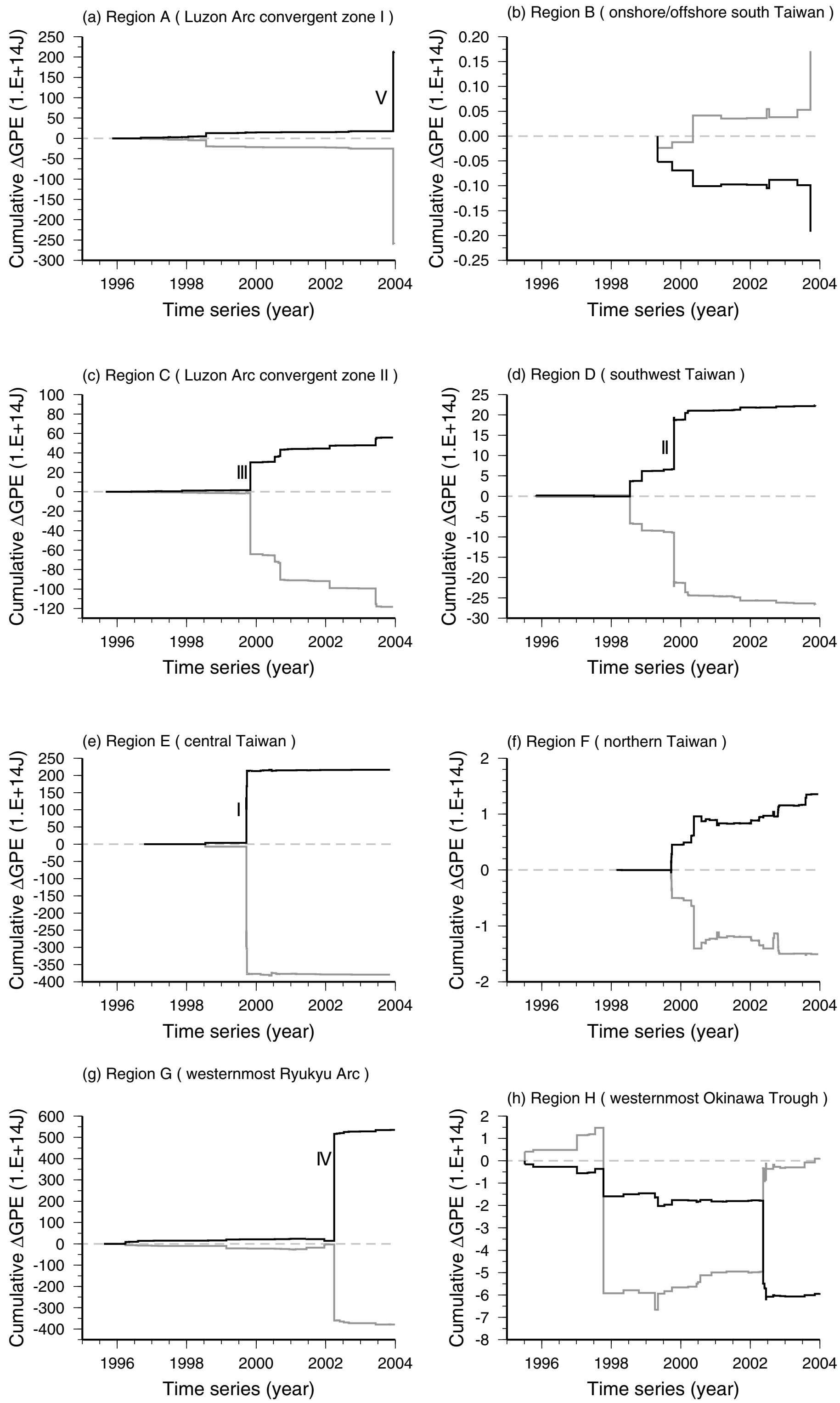

Figure 9. The temporal distributions of the crustal ΔGPE (heavy lines) and the lithospheric ΔGPE (grey lines) in different regions. The location of each region is shown in Fig. 8. The numbers indicate large contributions from the earthquake events listed in Table 1.

Luzon Arc collided with the westernmost Ryukyu Arc (Sibuet & Hsu 2004). After the collision, the Ryukyu Arc experienced a fast southward movement which caused the intense backarc rifting of the westernmost Okinawa Trough (Hsu 2001). Similar to region A, the average crustal ΔGPE in region G is positive (Fig. 8, Table 2). As shown in Fig. 8, the westernmost Okinawa Trough (region H) generally displays negative crustal ΔGPE which is generally due to the backarc rifting in terms of shallow normal faulting earthquakes. However, the magnitude of the crustal ΔGPE is about two orders less than in region G.

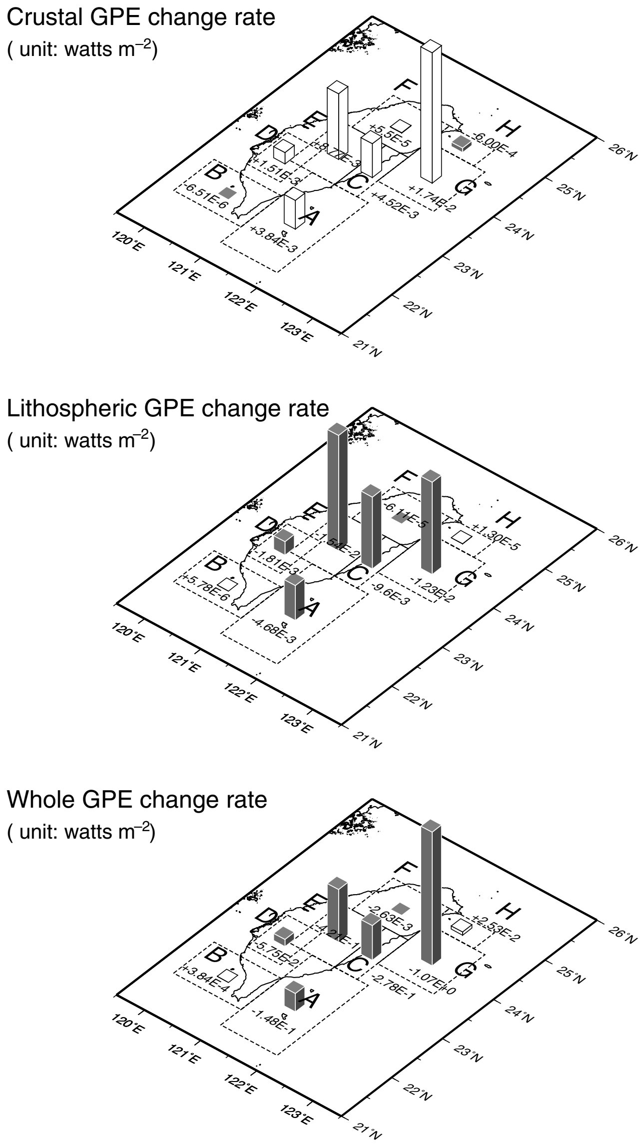

Figure 10. The average rate of ΔGPE per unit area. The black columns represent loss of GPE while the white columns represent gain of GPE. The value for each region is listed in Table 3.

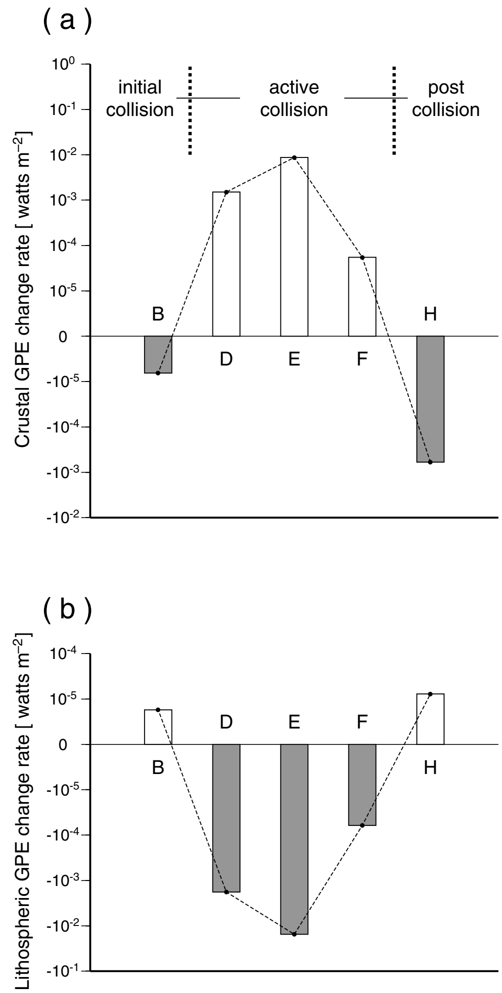

Figure 11. Diagram showing the relationship of (a) the crustal and (b) the lithospheric ΔGPE at different stage of the Taiwan orogenesis. The black columns represent loss of GPE while the white columns represent gain of GPE. The value for each region is listed in Table 3.

In the Manila subduction zone and the Ryukyu subduction zone the lithospheric ΔGPE shows a reverse pattern to the crustal ΔGPE (Table 2, Fig. 10). As mentioned previously, the lithospheric ΔGPE in the Okinawa Trough has fluctuated, which is probably because of the coexistence of shallow backarc rifting earthquakes and deep subduction earthquakes.

RELATIONSHIP BETWEEN ΔGPE AND THE OROGENESIS

To understand the relationship between the variation of the earthquake-induced GPE and the orogenic process, we have calculated the average rate of ΔGPE in our analysed regions (Fig. 8, Table 3). The result is illustrated in Fig. 10. It is noted that in front of the Taiwan mountain belt both the Ryukyu Arc and the Luzon Arc areas (regions A, C and G) show a positive crustal ΔGPE rate and a negative lithospheric ΔGPE rate. If we examine the ΔGPE rate of the Taiwan orogenic belt from the initial collision to the active collision and post-collision process (regions B, D, E, F and H), we find that increase or decrease of the crustal GPE follows the uplift or subsidence of the mountain belt (Figs 10 and 11). This phenomenon suggests that the crustal ΔGPE can be used as an indicator of mountain building status.

For the whole Taiwan orogenic region plate convergence generally induces compressive earthquakes and causes a loss of GPE by the Earth (Table 3, Fig. 10). In total, our study region exhibits a GPE loss of -5.54×10^{18} J or -2.07×10^{10} W, which corresponds to about 1 per cent of the global earthquake-induced GPE loss (Chao *et al.* 1995). The dramatic loss of GPE mainly happens in the actively converging region, i.e. central Taiwan (Fig. 10).

CONCLUSION

We have examined the earthquake-induced ΔGPE in the Taiwan orogenic region. The loss of GPE in the whole Taiwan region is very intense and is about 1 per cent of the global GPE loss. The main loss of GPE happens in the Luzon Arc, the Ryukyu Arc and central Taiwan. In contrast, these regions have gained the most crustal

Table 3. Average of the ΔGPE rate per unit area in different regions and different layer thicknesses (unit W m^{-2}).

	$R_b = 24.4$ km (crust)	$R_b = 125.0$ km (lithosphere)	$R_b = 6371.0$ km (Earth radius)
Region A (2.07E+10 m^2)	3.84E−03	−4.68E−03	−1.48E−01
Region B (1.10E+10 m^2)	−6.51E−06	5.78E−06	3.84E−04
Region C (0.46E+10 m^2)	4.52E−03	−9.60E−03	−2.78E−01
Region D (0.55E+10 m^2)	1.51E−03	−1.81E−03	−5.75E−02
Region E (0.92E+10 m^2)	8.77E−03	−1.54E−02	−4.21E+01
Region F (0.92E+10 m^2)	5.50E−05	−6.11E−05	−2.63E−03
Region G (1.15E+10 m^2)	1.74E−02	−1.23E−02	−1.07E+00
Region H (0.37E+10 m^2)	−6.00E−04	1.30E−05	2.53E−02
Whole region (2.59E+11 m^2)	1.49E−03	−1.66E−03	−7.99E−02

GPE. The variation of the crustal ΔGPE along the mountain belt expresses well the mountain building stages, from initial collision to post-collisional extension. However, the major changes in GPE were ascribed to several major earthquakes. This suggests the possibility of using crustal ΔGPE as an earthquake precursor. The gradual accumulation of GPE in a region could be regarded as a less harmless release of crustal stress. Thus, the lack of change in ΔGPE in southwest Taiwan after the year 2000 suggests an earthquake hazard.

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REFERENCES

Chao, B.F., Gross, R.S. & Dong, D.-N., 1995. Changes in global gravitational energy induced by earthquakes, *Geophys. J. Int.,* **122,** 784–789.

Dahlen, F.A., 1977. The balance of energy in earthquake faulting, *Geophys. J. R. astr. Soc.,* **48,** 239–261.

Dziewonski, A.M. & Anderson, D.L., 1981. Preliminary reference Earth model, *Phys. Earth planet. Inter.,* **25,** 297–356.

Hsu, S.-K., 2001. Lithospheric structure, buoyancy and coupling across the southernmost Ryukyu subduction zone: an example of decreasing plate coupling, *Earth planet. Sci. Lett.,* **186,** 471–478.

Hsu, S.-K. & Lo, C.-L., 2004. Change of crustal gravitational potential energy in the Taiwan orogen by the Chi-Chi earthquake sequence, *Earth planet. Sci. Lett.,* **222,** 573–581.

Hsu, S.-K. & Sibuet, J.-C., 1995. Is Taiwan the result of arc–continent or arc–arc collision?, *Earth planet. Sci. Lett.,* **136,** 315–324.

Hsu, S.-K., Sibuet, J.-C., Monti, S., Shyu, C.-T. & Liu, C.-S., 1996. Transition between the Okinawa Trough backarc extension and the Taiwan collision: new insights on the southernmost Ryukyu subduction zone, *Mar. Geophys. Res.,* **18,** 163–187.

Hu, J.-C., Yu, S.-B., Chu, H.-T. & Angelier, J., 2002. Transition tectonics of northern Taiwan induced by convergence and trench retreat, in *Geology and Geophysics of an Arc–Continent Collision,* Geological Society of America Special Paper 358, pp. 147–160, eds Byrne, T.B. & Liu, C.-S., Geological Society of America, Boulder, CO.

Kao, H. & Chen, W.-P., 2000. The Chi-Chi earthquake sequence: active, out-of-sequence thrust faulting in Taiwan, *Science,* **288,** 2346–2349.

Kao, H., Shen, S.-S.J. & Ma, K.-F., 1998. Transition from oblique subduction to collision: earthquakes in the southernmost Ryukyu arc-Taiwan region, *J. geophys. Res.,* **103,** 7211–7229.

Liu, C.-C., 1995. The Ilan plain and the southwestward extending Okinawa Trough, *J. Geol. Soc. China,* **38,** 229–242.

Ma, K.-F., Lee, C.-T., Tsai, Y.-B., Shin, T.C. & Mori, J., 1999. The Chi-Chi, Taiwan earthquake: large surface displacements on an inland thrust fault, *EOS, Trans. Am. geophys. Un.,* **80,** 605–611.

Okamoto, T. & Tanimoto, T., 2002. Crustal gravitational energy change caused by earthquakes in the western United States and Japan, *Earth planet. Sci. Lett.,* **195,** 17–27.

Shyu, C.-T., Chih, M.-C., Hsu, S.-K., Wang, C. & Karp, B., 1996. Northern Luzon Arc: location and tectonic features from magnetic data off eastern Taiwan, *TAO,* **7,** 535–548.

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

Sibuet, J.-C. et al, 1987. Back arc extension in the Okinawa Trough, *J. geophys. Res.,* **92,** 14 041–14 063.

Sibuet, J.-C., Deffontaines, B., Hsu, S.-K., Thareau, N., Le Formal, J.-P., Liu, C.-S. & the ACT party, 1998. Okinawa Trough backarc basin: early tectonic and magnetic evolution, *J. geophys. Res.,* **103,** 30 245–30 267.

Tanimoto, T. & Okamoto, T., 2000. Change of crustal potential energy by earthquakes: an indicator for extensional and compressional tectonics, *Geophys. Res. Lett.,* **27,** 2313–2316.

Tanimoto, T., Okamoto, T. & Terra, F., 2002. Tectonics signatures in coseismic gravitational energy change, *Geophys. J. Int.,* **149,** 490–498.

Wang, K.-L., Chung, S.-L., Chen, C.-H., Shinjo, R., Yang, T.F. & Chen, C.-H., 1999. Post-collisional magmatism around northern Taiwan and its relation with opening of the Okinawa Trough, *Tectonophysics,* **308,** 363–376.

Wessel, P. & Smith, W.H.F., 1998. New improved version of Generic Mapping Tools released, *EOS, Trans. Am. geophys. Un.,* **79,** 579.

Yeh, Y.-H., Barrier, E., Lin, C.-H. & Angelier, J., 1991. Stress tensor analysis in the Taiwan area from focal mechanism of earthquakes, *Tectonophysics,* **200,** 267–280.

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