

Joint Application of Altimeter Data and EGM96 to Submarine Tectonic and Geodynamic Study in West Pacific

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Abstract. With altimeter data from Geosat, ERS-1/2 and Topex/Poseidon, and the EGM96 geopotential model, we calculated various Earth gravity field models for the west Pacific, covering the area over 0~45°N and 100~150°E. Within the Philippine Sea and the South China Sea, lineations of the residual geoid relative to EGM96 are found to be parallel to or across spreading ridges. Some trails of spreading ridges and trench-arcs are evident in the gravity anomaly signature. The intensities of the subduction activity with the accompanying resistant compression are different along different segments of trench-arcs. The Moho depth is inversed from the Glenni geoid undulation, and the tectonic implications are discussed. The stress fields from long- and medium-wavelength mantle convections are calculated from EGM96. The stress field from short-wavelength mantle convection is calculated from the isostatic geoid undulation to a reasonable degree of precision. The models are discussed jointly to elucidate the geodynamic processes of evolution of the marginal sea basins and troughs. All these models display unique features that are representative of the tectonics and geodynamics of the marginal seas.

Keywords. Altimeter Data, EGM96, Tectonics, Geodynamics, Marginal Seas, West Pacific

1 Introduction

Satellite altimeter data can be used to study submarine tectonics by calculating the geoid undulation, refining the geopotential model and recovering the gravity anomaly. The relevant theory and methods have reached maturity to some extent (Rapp et al., 1991; Sandwell & Smith, 1997; Hwang, 1998). Studies of this kind have found many typical submarine structures, such as volcanic mountain chains, mid-ocean ridges, faults and trenches. Seafloor topographic relief and submarine lithospheric structures have been numerically

modeled in detail (Wang et al., 1995). The elastic thickness and the age of oceanic lithosphere can be determined from the isostatic analysis of seamount load (Watts & ten Brink, 1989). The west Pacific is characterized by large amounts of trenches, island-arcs and marginal seas (Fig. 1), and is an important region for tectonic and geodynamic studies. Although some models of geoid undulation, gravity anomaly and seafloor topography have been obtained for the west Pacific marginal seas (Hwang, 1998; Hwang, 1999; Hsu, 1999), they have not been closely tied to seafloor tectonics. In this study, we investigate the tectonics of this region with the altimeter data.

As an important study, Bowin (1981) used Geos-3 altimeter data directly to investigate the west Pacific tectonics. There are other studies that used the altimeter data indirectly. For example, Runcorn (1967) and Huang & Fu (1982) used geopotential model GEM10's series to study the west Pacific geodynamics, based on the Runcorn's theory of mantle convection.

With the launch of Geosat (1985~1989), ERS-1 (1991~1996), ERS-2 (1996~) and Topex/Poseidon (1992~) satellites, the precision and application of altimeter data have reached an unprecedented level. The altimeter data from Geosat, ERS-1/2 and Topex/Poseidon and the EGM96 geopotential model contain much more reliable and abundant information about the submarine tectonics and geodynamics than ever before. This study makes use of this rich data set for the west Pacific (Fig. 1). First, we discuss some improved techniques that we use to retain the precision and resolution of the altimeter data and to calculate the geoid undulation and gravity anomaly in combination with EGM96. We then describe methods that we use to extract submarine tectonic information, to eliminate topographic and isostatic effects, to inverse the Moho depth and to calculate the stress field from short-wavelength mantle convection. Finally, we discuss the tectonic and geodynamic implications of these derived models for the west Pacific region.

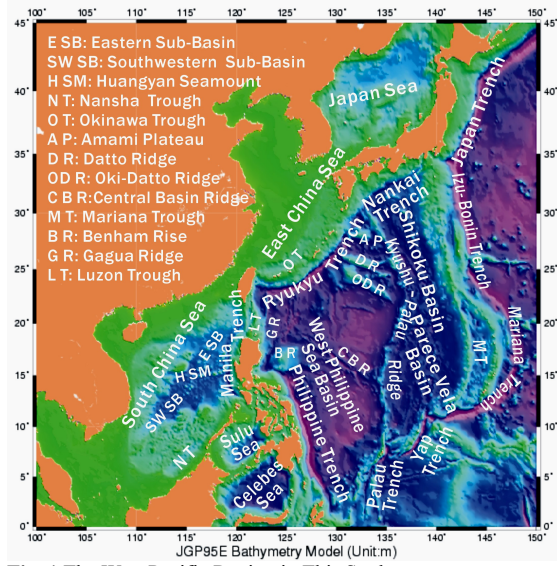


Fig. 1 The West Pacific Region in This Study

2 Calculation of Geoid Undulation and Gravity Anomaly from Altimeter Data and EGM96

2.1 Calculation of Geoid Undulation

We used the single-satellite crossover adjustment method (Tai, 1988) to reduce the radial orbit error of the Geosat T2/ERM altimeter data and the so-called dual-satellite crossover adjustment method (Le Traon et al., 1995) to reduce the radial orbit error of the ERS-1/2 data with Topex/Poseidon data. The high density of ERS-1/2 (Oct.,1992~Oct.,1999) combined with the high accuracy of Topex/Poseidon (Oct.,1992~ Oct.,1999) improved the quality of the altimeter data. After editing, correcting and making crossover adjustment of the altimeter data, all corrected SSH data are weighted and distributed onto their neighbouring grid nodes ($4' \times 4'$) in order to obtain the average sea level. In order to retain the high resolution of the ERS-1/2 data, we did not use collinear stacking average (Douglas & Cheney, 1981) to model the average sea level. We calculated the SST from oceanographic measurements and estimations (Zhao and Gao, 1999) and removed the SST from the average sea level to obtain the altimetry-derived geoid undulation (Fig.2). Geoid undulation over land is directly calculated from the EGM96 geopotential model.

In order to check the accuracy of the geoid undulation model, we subtracted a reference geoid model based on EGM96 to degree/order 360 from

our model. The residual geoid is displayed in Fig. 3. The residuals are distributed in the range $-2.6 \sim 3.2$ m with a mean of 0.35 m and a standard deviation of ± 16.8 cm. Fig. 3 shows that the residuals are somewhat larger near land. It may result partly from the incompleteness of tide and SST corrections, which are heavily affected by complicated oceanic dynamics off coast. Nevertheless, the geoid undulation model is very accurate in general. The residual geoid essentially reflects fine structures of the seafloor topography. Basins, trenches, ridges and seamounts (chains) are depicted more clearly by the residual geoid (Fig. 3) than by the geoid undulation model itself (Fig. 2). The high accuracy of the geoid undulation model is due to the long time span and closely spaced orbits of the altimetry satellite deployment.

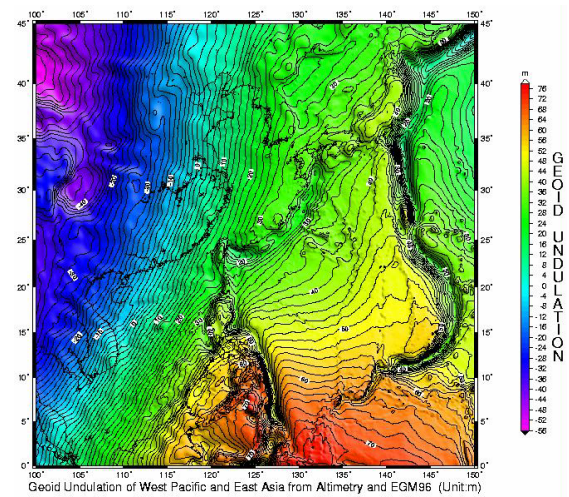


Fig. 2 Altimetry-derived Geoid Undulation (Unit: m)

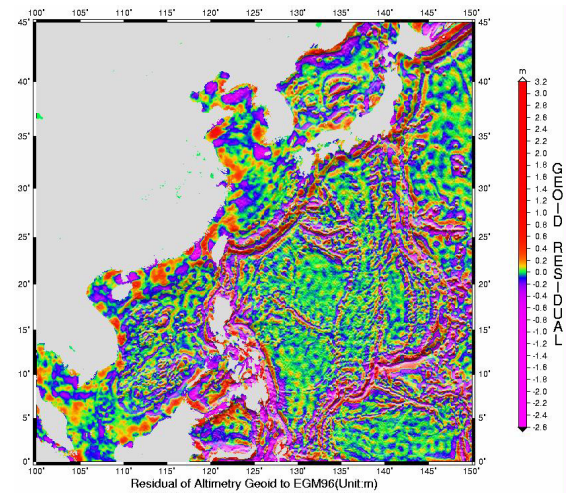


Fig. 3 Residual Geoid to EGM96 (Unit: m)

2.2 Recovery of Gravity Anomaly

We used an improved algorithm based on the method of vertical deflection (Sandwell & Smith, 1997; Hwang, 1998) to recover the gravity anomaly. We calculated along-track slope and direction for every cycle. Values for repeated cycles were averaged. Here we used collinear stacking average to reduce high-frequency errors contained in the vertical deflection. The remove-restore procedure of the reference model EGM96 recovers the gravity anomaly only for higher-frequency components. The better linearity of these components reduces the integrated area for each grid node and FFT is available in a small window (32×32). Since the quality of data obtained by the inverse FFT may be degraded at the window edges, we only retain results for the four center nodes. When the window was shifted all over the study region, we obtained the grid model of altimetry gravity anomaly shown in Fig. 4. Compared with shipboard gravity data for the South China Sea, it has a standard deviation of ±7.5 mGal.

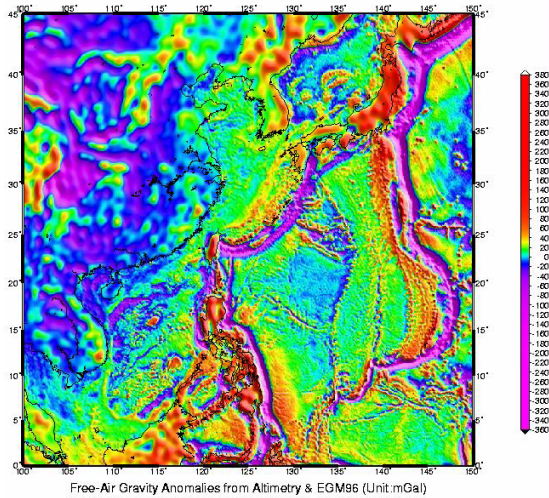


Fig.4 Altimetry Gravity Anomaly (Unit: mGal)

3 Extraction of Submarine Tectonic and Geodynamic Information

3.1 Topographic, Isostatic Reduction for Geoid Undulation and Gravity Anomaly

We expanded the algorithm of generalized topographic and isostatic reduction for gravity anomaly (Lei, 1984) to include corrections for geopotential field and geoid undulation. Fig. 5

shows a schematic diagram of a fan-shaped block in the spherical crust. A is the computing point with a geocentric radius $OA = R_A$. B(α, θ, R) is an arbitrary point in the fan-shaped block in the spherical crust. The distance between A and B is $\rho = (R_A^2 + R^2 - 2 R_A R \cos \theta)^{1/2}$.

Making unitless, let $r = R/R_A$, $r_1 = R_1/R_A$, $r_2 = R_2/R_A$, $L = \rho/R_A = (1 + r^2 - 2 r \cos \theta)^{1/2}$, and

$$E_{\delta I}(\theta, r) = L (2 + 2 r^2 - r \cos \theta - 3 \cos^2 \theta) - 3 \cos \theta \sin^2 \theta \ln(r - \cos \theta + L) \quad (1)$$

$$E_{\delta \Delta g}(\theta, r) = L (2 - r^2 - r \cos \theta - 3 \cos^2 \theta) + 3 \cos \theta \sin^2 \theta \ln(r - \cos \theta + L) \quad (2)$$

The geoid undulation and gravity anomaly at point A produced by the fan-shaped block are expressed as

$$\delta N = \frac{1}{6g} G \sigma R_A^2 (\alpha_2 - \alpha_1) \|E_{\delta I}(\theta, r)\|_{r_1}^{r_2} \Big|_{\theta_1}^{\theta_2} \quad (3)$$

$$\delta \Delta g = -\frac{1}{3} G \sigma R_A (\alpha_2 - \alpha_1) \|E_{\delta \Delta g}(\theta, r)\|_{r_1}^{r_2} \Big|_{\theta_1}^{\theta_2} \quad (4)$$

Here G is the gravitational constant, g is the normal gravity and σ is density difference to the normal crust. The fan-shaped block can be used to approximate any crust element on an ellipsoidal surface. The term $\sin^2 \theta \ln(r - \cos \theta + L)$ in Formulae (1) and (2) approaches zero when r approaches zero. The expression of reduction from this generalized fan-shaped block remains unchanged both in the inner and outer zones. The method has much higher precision and flexibility than other approximation methods do (Heiskanen & Moritz, 1967).

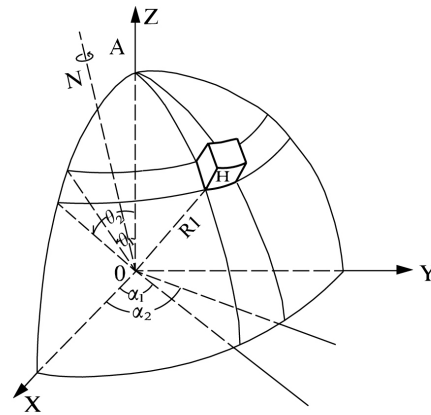


Fig.5 Fan-shaped Spherical Crustal Blocks

We used the global digital topographic model JGP95E (Arabelos, 1999) for outer-zone reduction (outside the 166.7-km Hayford radius), and shipboard bathymetric data and the 2'×2' seafloor topographic data (Smith & Sandwell, 1994) for inner-zone reduction (inside the Hayford radius). Topographic data need to be mapped into a new ring-shaped grid model around each computing point. Contributions from each fan-shaped block are summed up and used to correct the geoid undulation and gravity anomaly at the computing point. We used the Airy isostatic compensation model in the isostatic reduction with a compensation depth of 30 km. We first corrected the geoid undulation and gravity anomaly for the global topography to obtain the Bouguer geoid undulation and gravity anomaly. We then corrected for the outer-zone isostasy to obtain the Glenni geoid undulation and gravity anomaly (Glennie, 1932). Finally, we corrected for the inner-zone isostasy to obtain the isostatic geoid undulation and gravity anomaly. These models are all presented as matrices with 4'×4' grid spacing. The isostatic gravity anomaly, Glenni geoid undulation and isostatic geoid undulation are shown in Figs. 6, 7 and 8 respectively.

3.2 Inversion of Moho Depth from Glenni Geoid Undulation

The fact that the Glenni geoid undulation (Fig. 7) is almost inversely proportional to the topography implies that the Moho depth is approximately a mirror image of the topography because of the requirement for isostatic equilibrium of the crust. In the study region, the Glenni geoid ascends broadly from northwest to southeast, being less than -120 m in Tibet and larger than 130 m in the southern Philippine Sea. There are some local uplifts in the continental margin seas. Obviously, the model appears as gradient belts corresponding to tectonic boundaries, and facilitates inversion of the Moho depth. The lateral heterogeneities of the mantle affect the recovery of the subtle variations of the crustal structure and needs to be removed. According to Bowin's formula for an equivalent point-mass depth for each degree- n harmonic term of the geopotential field (Bowin, 1986), $d_n = R/(n-1)$, we removed the geoid undulation represented by EGM96 harmonic terms up to degree 36 from the Glenni geoid undulation (Fig. 7) to reduce the mantle heterogeneity effects.

The corrected Glenni geoid undulation displays high-frequency components related to local structures reasonably well. But information about

regional structures may have been lost. We solved this problem by removing and restoring a reference model of the Moho depth with large- and medium-scale structures, as we did in the remove-restore procedure of EGM96 to recover the gravity anomaly. Combining data from the 2°×2° global crust and mantle model 3SMAC (Nataf & Richard, 1996), the 1°×1° global sediment thickness model (Laske & Masters, 1997) and the 5'×5' global topographic model JGP95E (Arabelos, 1999), we built a 2°×2° Moho reference model. After the removal of mantle heterogeneity and the Moho reference model, the Glenni geoid undulation mainly reflects the high-frequency fluctuation of the Moho depth. Similar to the recovery of gravity anomaly, we achieved convergence after 2 to 3 times of iteration with an FFT inversion algorithm (Parker, 1973) available in a small shifted window (32×32). Because the high-frequency fluctuation h of the Moho discontinuity is much smaller than its average depth z_0 , in the frequency domain K , where k represents the amplitude of K , the forward and the inverse formulae are

$$\tilde{N}(K,0) = \frac{2\pi G \sigma}{g} e^{-2\pi k z_0} \sum_{n=1}^{\infty} \frac{(2\pi k)^{n-2}}{n!} \tilde{h}^n(K) \quad (5)$$

$$\tilde{h}(K) = \frac{g}{G \sigma} k e^{2\pi k z_0} \tilde{N}(K,0) \quad (6)$$

Here σ is density difference between the mantle and the crust. The inversed Moho depth is shown in Fig. 9. Compared with the 2°×2° Moho reference model, the inversed model contains more subtle tectonic features.

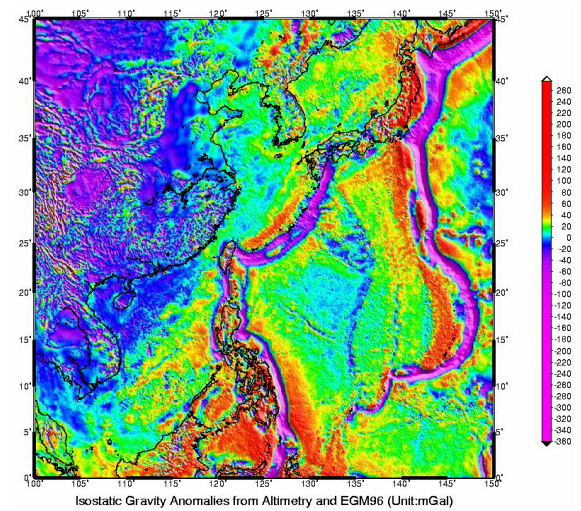


Fig. 6 Isostatic Gravity Anomaly (Unit: mGal)

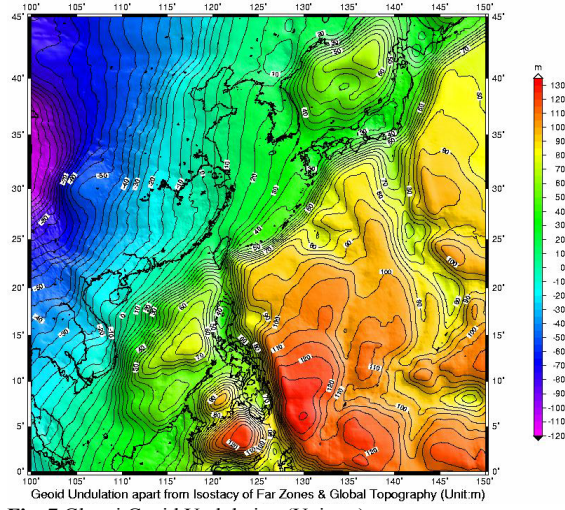


Fig. 7 Glenni Geoid Undulation (Unit: m)

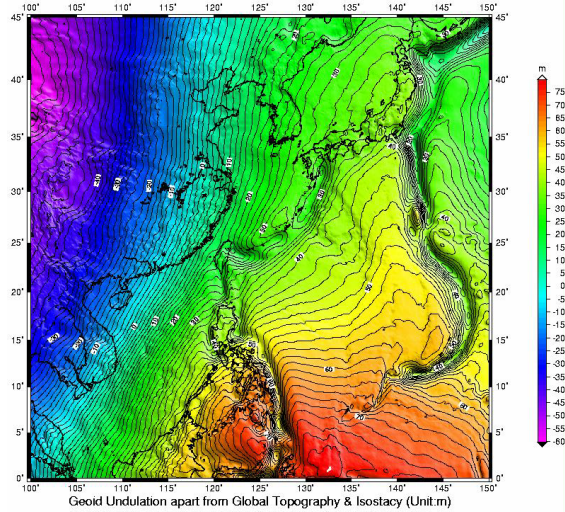


Fig. 8 Isostatic Geoid Undulation (Unit: m)

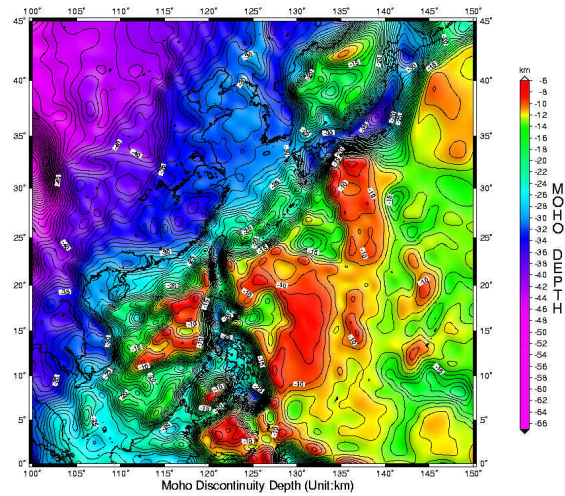


Fig. 9 Depth to Moho Discontinuity (Unit: km)

3.3 Calculation of Stress Fields of Mantle Convection

The isostatic geoid undulation in Fig. 8 displays larger geopotential nonequilibrium than that in the geoid undulation in Fig. 2. This fact indicates that the topographic and isostatic effects related to a hydrostatic equilibrium between the crust and the upper mantle play a roll in mitigating mantle dynamic nonequilibrium, which dominates the global geodynamic regime and drives lithospheric tectonic movements through mantle convections. Runcorn (1967) established a basic relationship between the geopotential model and the mantle thermal convection. From this relationship, the stress fields beneath the lithosphere can be calculated. To better characterize the stress field of the mantle convection, we define a new concept — the pseudo-potential field of the mantle convection, the horizontal gradient of which being the stress field of the mantle convection. This new field is introduced to represent the distribution of mantle geodynamic energy. With the introduction of this new field, we can make stress calculations directly from the arithmetic rule of the geoid and the vertical deflection from the harmonic coefficients. Corresponding to harmonic coefficients $(\bar{C}_{nm}, \bar{S}_{nm})$ of the geopotential model, the pseudo-potential field of the mantle convection and its harmonic coefficients are given as

$$V(\rho, \theta, \lambda) = \frac{GM}{\rho} \sum_{n=2}^{\infty} \left(\frac{R}{\rho} \right)^n \sum_{m=0}^n (\bar{C}'_{nm} \cos m\lambda + \bar{S}'_{nm} \sin m\lambda) \bar{P}_{nm}(\cos \theta) \quad (7)$$

$$\bar{C}'_{nm} = -\frac{g}{4\pi G} \frac{R}{\rho} \frac{2n+1}{n+1} \bar{C}_{nm} \quad \bar{S}'_{nm} = \frac{g}{4\pi G} \frac{R}{\rho} \frac{2n+1}{n+1} \bar{S}_{nm} \quad (8)$$

Here (ρ, λ, θ) are the spherical coordinates of an arbitrary point on the outer surface of the mantle convection system. M and R are the mass and average radius of the Earth respectively. If the harmonic coefficients $(\bar{C}_{nm}, \bar{S}_{nm})$ of the geopotential model are replaced by $(\bar{C}'_{nm}, \bar{S}'_{nm})$ as like in Formula (7), the “disturbed potential” beneath the lithosphere becomes the pseudo-potential field of the mantle convection. The “vertical deflection” of this field multiplied by $-g$ is then the stress of the mantle convection. Following this approach, we calculated the pseudo-potential and stress fields for the long- and medium-wavelength mantle convection. The results are given in Figs. 10 and 11. Fig. 10 corresponds to degrees 4~11 and Fig. 11 corresponds to degrees 12~36.

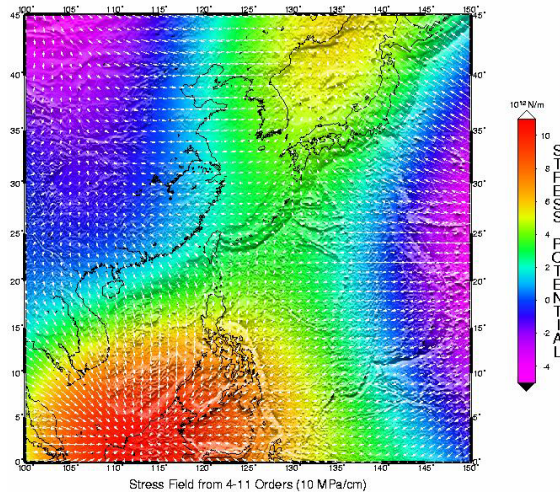


Fig. 10 Stress from Long-wavelength Mantle Convection

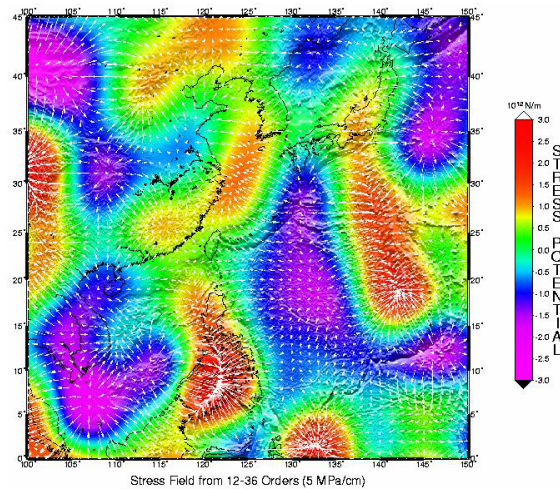


Fig. 11 Stress from Medium-wavelength Mantle Convection

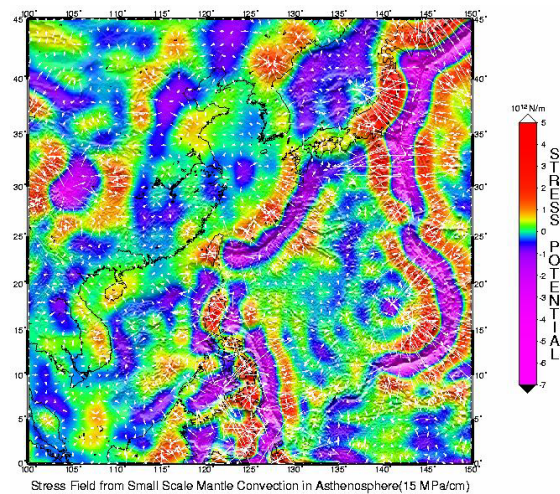


Fig. 12 Stress from Short-wavelength Mantle Convection

From Formulae (7) and (8) it can be seen that the pseudo-potential field of the mantle convection is approximately proportional to the Earth's disturbed potential, or the geoid undulation. The higher the degree is, the better the proportional relationship is. For higher-degree terms ($n \rightarrow \infty$), Formula (7) shows that, from the Earth's surface down to the bottom of the lithosphere, the pseudo-potential field of the mantle convection can be approximately obtained by multiplying the geoid undulation by a scaling factor $g^2/2\pi G$. Short-wavelength components of the geoid are seriously affected by the topography and density anomaly in the lithosphere. The isostatic geoid undulation (Fig.8), which is corrected for topography and isostasy, therefore, is the ideal model for calculating short-wavelength mantle convections. The model needs to be corrected for long- and medium-wavelength contributions represented by EGM96 harmonic terms up to degree 36, and the components with wavelengths shorter than 50 km need to be filtered out as well. With this approach and in the Cartesian coordinate system, we reduced the isostatic geoid undulation and continued it from the Earth's surface down to the bottom of the lithosphere to obtain the pseudo-potential field for short-wavelength mantle convection and the corresponding stress field. The stress field is shown in Fig. 12.

4 Implications of the Models for Submarine Tectonics and Geodynamics

4.1 Submarine Tectonic Features

The altimetry-derived geoid undulation (Fig.2) features convergence of the Eurasian plate and the Pacific plate. Appearing in the east of the Philippine Sea, the geoid uplifting belt continues to rise and expand towards the Equator. The global geoid is the highest at New Guinea ($5^{\circ}20.4'S$, $142^{\circ}40.2'E$). The geoid abruptly drops about 8~10 m between the northern extending ends of the Luzon arc and the Gagua ridge east of Taiwan, where, accompanied by a deficit of mass, the Philippine Sea plate changes from subduction to obduction. The geoid depresses more distinctly along the Philippine trench and the Izu — Bonin — Mariana trench than along the Ryukyu trench and the Palau — Yap trench. In terms of a double trench-arc system, subduction may be more active along the outer belt than along the inner belt. Furthermore, along the outer belt, subduction may be more active in the north than in the south. Along the inner belt, it is the opposite: subduction

may be more active in the south than in the north. These features are also prominently displayed in the altimetry gravity anomaly model (Fig. 4), in the isostatic gravity anomaly model (Fig. 6), in the Glenni geoid undulation model (Fig. 7) and in the isostatic geoid undulation model (Fig. 8).

The residual geoid (Fig. 3) derived solely from altimeter data reveals many intraplate-structure features such as fore-arc basins and out-trench rises. The positively beaded belt over the Kyushu—Palau ridge forms a boundary between different lineation trends in the west and east Philippine Seas. One set of parallel lineations is nearly perpendicular to the central basin ridge in the west Philippine Sea Basin, and another set is across the central ridge in the Parece Vela Basin. In the Japan Sea, the residual geoid lineations coincide with the NE trending structures. In the East China Sea, three positive belts trending NNE are located in the middle of the shelf basin, along the outer-shelf uplift and the Okinawa Trough, respectively. In the southwestern sub-basin of the South China Sea, a negative belt trending NE is located in the center of the basin, where there may be a spreading ridge. On both sides of the spreading ridge, the lineations appear to be in the parallel direction. In the eastern sub-basin of the South China Sea, the lineations are dominated by N—S trending, except for those along the Huangyan seamount chain.

Tectonic features indicated by the residual geoid are present in the altimetry gravity anomaly (Fig.4) as well. However, more features of trench and arc structures can be found from the gravity anomaly. The gravity anomaly is lower along the Luzon trough than along the Manila trench. Anomalies at both places are not the same as those along other arcs and trenches. A negative belt of gravity anomaly along the Philippine trench has two thin branches nearly reaching the Ryukyu trench, one on the east side of the Luzon arc, and the other on the east side of the Gagua ridge. Similarly, another negative belt along the Ryukyu trench is separated into two branches at its northern end, one stretching to the Japan Sea, and the other into the Shikoku Basin. Both positive and negative belts of gravity anomaly with large amplitudes occur on the north of the Japan Sea and on the south of the South China Sea. Similar to those accompanied by known trenches and arcs, they may be related to relict trenches and arcs. In addition, the altimetry gravity anomaly shows many continental shelf basins.

The isostatic gravity anomaly (Fig. 6) is generally positive over seas, except along trenches. The anomaly has higher amplitude outside trenches

and behind arcs. There are two belts of higher amplitude (about 40~80 mGal) anomaly, one along the Mariana Trough, and the other along the Okinawa Trough. They may be the result of mass surplus caused by the intrusion of the asthenospheric material during the trough forming. The amplitude and the area of the positive isostatic gravity anomaly increase from South China Sea in the north toward Celebes Sea in the south with Sulu Sea in the middle. They may also be related to the intrusion of the asthenospheric material. The isostatic gravity anomaly outside of the Philippine trench is the largest in the study region. These large-value positive belts or areas indicate that the troughs and seas in this area have had intensive tectonic activities.

From Fig. 7, Fig. 9 and Table 1 it is evident that in the Philippine Sea from west to east, the Glenni geoid undulation and Moho depth correlate to the age of the basins with lower Glenni geoid undulations and deeper Moho depths corresponding to younger basins. Crustal thickness decreases to about 14~16 km in the southern Okinawa Trough (Fig.9), where there occurred a process of evolution from continental crust to oceanic crust. From Fig. 7, Fig. 9 and Table 2 it can be seen that, different from what is observed in the Philippine Sea, in marginal seas from north to south, higher Glenni geoid undulations and shallower Moho depths correspond to younger basins. The phenomenon implies that the strengths of the tectonic activities are almost the same on both sides of the Ryukyu—Taiwan—Philippine arc, but their patterns are different.

Table 1. Glenni geoid and Moho depth in Philippine Sea

Basins	Glenni geoid(m)	Moho depth(km)
West Philippine Sea	<130~135	8~14
Parece Vela	<110~115	10~12
Shikoku	<90~95	9~13
Mariana Trough	<80~85	10~12

Table 2. Glenni geoid and Moho depth in marginal seas

Basins	Glenni geoid(m)	Moho depth(km)
Japan Sea	<65~70	12~20
South China Sea	<75~80	9~18
Sulu Sea	<95~100	9~16
Celebes Sea	<120~125	8~14

4.2 Marine Geodynamic Features

According to Bowin's formula for an equivalent point-mass depth (Bowin, 1986), stress fields from

long-, medium- and short-wavelength mantle convections shown in Figs. 10, 11 and 12 are applicable to the lower mantle, the upper mantle and the asthenosphere, respectively.

Consistent with the southeastward movement of the Eurasian plate and the northwestward movement of the Pacific plate, the stress field from the long-wavelength mantle convection (Fig. 10) shows that the lower-mantle dynamic energy slips in the middle and converges toward both north and south ends. The stress convergences appear as sinistral compresso-shear on the north and dextral compresso-shear on the south. This is because the stress from the Pacific plate (about 3~7.5 MPa) is larger than that from the Eurasian plate (about 1~6 MPa). The energy convergence on the south (about 12×10^{12} N/m) is twice that on the north (about 6×10^{12} N/m). It may be attributed to the northward movement of the India-Australian plate and would shed light on why there are more basins with large areas and young ages in the south. The stresses from the long-wavelength mantle convection are generally perpendicular to the Japan trench-arc, the Philippine trench-arc and the Isu—Bonin—Mariana trench-arc. It implies that these trench-arcs are dominated by the lower-mantle convection.

There appears to be a medium-wavelength (upper) mantle convection in the Philippine Sea region (Fig. 11). It may be related to some kind of energy adjustment for the lower-mantle convection. In the Ryukyu trench-arc region, the stress from upper-mantle convection reaches 4 MPa and is perpendicular to the trench-arc. On the other hand, the stress from lower-mantle convection in the same region is no more than 1 MPa and is parallel to the trench-arc. Same feature also exists along the Luzon trough-arc. The phenomenon implies that both the Ryukyu trench-arc and the Luzon trough-arc might be driven by the upper-mantle convection. In the Philippine trench-arc region, however, stresses from both the upper-mantle convection and the lower-mantle convection are perpendicular to the trench-arc, though the magnitude of the stress from the upper-mantle convection is 6 MPa whereas that of the stress from the lower-mantle convection is no more than 3.5 MPa. It indicates that the subduction along the Philippine trench-arc is driven by both the lower-mantle and the upper-mantle convections.

There are two ascending and diverging centers of the upper-mantle convection, one on the north of the Japan Sea, and the other on the southwest of the South China Sea (Fig. 11). The center on the south is stronger and has a larger scale. These centers may be related to the energy convergence of the

lower-mantle convection. The locations of these centers are in good agreement with the existence of relict trenches and arcs indicated by the gravity anomaly.

Stress fields from short-wavelength asthenospheric current (Fig. 12) are generally divergent in back-arc basins and along trenches, and convergent along arcs and outside of trenches. Although convergent stress from short-wavelength asthenospheric current is weaker outside of trenches than along arcs, it indicates certain obstruction experienced by the subduction slab. The stress outside the Japan—Isu—Bonin—Mariana trench is much larger than that outside the Ryukyu—Philippine trench. This is consistent with conclusions drawn from the geoid undulation model analysis that the subduction may be more active along the outer belt than along the inner belt.

There are two belts of convergent stress from short-wavelength asthenospheric current along the Okinawa Trough and the Mariana Trough (Fig. 12). Both belts seem to contradict the trough rifting. However, they coincide with the two belts of higher-amplitude isostatic gravity anomaly (Fig. 6). It may be an indication of the beginning stage of a developing back-arc basin. At this stage, the asthenospheric current would be convergent and would not have flown back to the mantle entirely. Part of the magma would intrude into the lithosphere and produce lava diapirism in the crust. The lithosphere and the crust would then be melted, thinned and rifted to provide space for asthenospheric current to gradually change from convergence to divergence.

5 Discussions and Conclusions

(1) Collinear stacking average may degrade the resolution of the altimeter data as repeated orbits do not follow the same track exactly. Therefore, we did not use the method in modeling the geoid undulation. The along-track slope and direction for single orbit may contain high-frequency errors, which can be reduced by collinear stacking average to improve the accuracy of the gravity anomaly recovered from the vertical deflection. The use of the generalized topographic and isostatic reduction method is necessary for extracting tectonic and geodynamic information from the gravity anomaly and the geoid undulation, such as the Moho discontinuity and the stress field from the short-wavelength mantle convection. The remove-restore procedure of reference model and FFT available in the shifted window can achieve

high precision and resolution of the gravity anomaly and the Moho discontinuity calculated from the altimeter data. The pseudo-potential field of mantle convection is introduced to make the arithmetic rule of the stress field to be coherent with that of the geopotential field. With the introduction of the pseudo-potential field of the mantle convection, the physical relationship between the stress field and the geopotential field becomes clearer and more straightforward.

(2) Consistent with the convergence of the Eurasian plate and the Pacific plate, the geoid rises toward the Equator in the Philippine Sea. The long-wavelength lower-mantle convection converges in at the northwest of the Japan Sea and the southeast of the South China Sea, where strongly converging energy is attributed to the northward movement of the Indian-Australian plate. The movement has created some basins with large areas and young ages. For the west Pacific double-trench-arc system, the geoid undulation and the gravity anomaly show that subduction may be more active along the outer belt than along the inner one. Moreover, along the outer belt subduction may be more active on the north than on the south. They may be related to the change of spreading direction of the Pacific plate from NNW to NWW during 50-40 Ma (Hild, 1977).

(3) Along the inner belt of the west Pacific double-trench-arc system, slipping of the stress from the long-wavelength lower-mantle convection may be responsible for the weak subduction in the region, which is mainly driven by the medium-wavelength upper-mantle convection beneath the Philippine Sea. Subduction along the Philippine trench, which is stronger than that along the Ryukyu trench, is partly reinforced by the Pacific NWW trending motion. That is why main structures in the eastern sub-basin of the South China Sea are parallel to the Philippine trench-arc. Stress fields from the long- and medium-wavelength mantle convections may jointly lead to the northwestward rift of the Japan Sea and the southeastward rift of the South China Sea, which are signaled by gravity anomalies.

(4) Except for the Okinawa Trough, a southward migration and enhancement of tectonic activities of the marginal sea basins resulted in deep seafloors and thin crust. This is in contrast to the eastward migration and enhancement of tectonic activities in the Philippine Sea. This observation, along with the corresponding stress field patterns from long- and mediums-wavelength mantle conventions, indicates that the marginal seas have their unique tectonic and geodynamic regime, different from that of the

Philippine Sea and the Pacific Ocean. The strength of the tectonic activity of the marginal seas is comparable to those of the Philippine Sea and the Pacific Ocean tectonic activities.

(5) Large positive values of the isostatic gravity anomaly at the Okinawa Trough may imply that the convergent asthenospheric current has not fully descended, but partly ascended into the lithosphere and produced lava diapirism in the crust. As the beginning stage of a back-arc basin development, the trough may also be affected significantly by lithospheric activation (Gao et al., 2002). The plate subduction would cause the lithosphere to fold and uplift along the island-arc, then tilt down and extend behind the arc. In addition, shear-opening structures in the lithosphere could occur in the trough, which are probably accompanied by the strongly compressional structures in Taiwan. Coupling of the lithospheric activation and asthenospheric current would be the main reason for the trough to rift and depress.

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