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An Experimental Study of Regional Heterogeneities of Gravity Tides in Central Japan

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Abstract

Tidal gravity observations have been conducted at ten stations in Central Japan with two LaCoste & Romberg gravimeters for the purpose of investigating regional variations of gravity tides. Analyses of the obtained data have revealed remarkable regional heterogeneities of tidal gravity parameters. These heterogeneities cannot be eliminated by ocean tide corrections. After ocean tide corrections the obtained δ-factors show a tendency to decrease from the south-eastern part of this area toward the north-western part by 1.5% for M_2 constituent and by 3% for O_1 one. The obtained phase differences after ocean tide corrections also show systematic variations associated with the δ-factor variations. Such large lateral variations in the δ-factor, occurring within a range of about a few hundred kilometers, have never been clearly obtained in previous studies.

We have examined several possible causes of these δ-factor variations, such as correction errors of ocean tide effects, meteorological disturbances, groundwater perturbations and topographic effects, but none of them can explain the δ-factor variations. The area of a lower δ-factor corresponds approximately to the center of negative Bouguer gravity anomaly and beneath that area exists an attenuative and low velocity layer. Therefore, the lateral heterogeneity of the earth which has both elastic and anelastic properties seems to be a likely cause, but a satisfactorily quantitative explanation cannot be given at present. Improvements of an earth tide theory including effects of lateral heterogeneities are required for quantitative explanations.

Contents

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1. Introduction

Forced oscillations of the earth are caused by tide-generating forces of the moon and sun. Since the period of the fundamental free oscillation of the earth (about 54 minutes) is considerably less than that of the earth-tidal waves (diurnal or semi-diurnal period), earth tides generally are considered to be of the equilibrium type. Then we ordinarily assume that the earth reaches equilibrium with the tide-generating forces instantaneously, so that the phase difference of each tidal constituent is zero. We further assume that the earth is spherical and that its physical properties vary only as a function of radius. The ratio of the observed tidal amplitudes on such an ideal earth to the theoretical equilibrium tidal amplitudes, which we call a tidal parameter, would be the same at all points on the earth's surface for all tidal constituents. However, results of modern observations have shown that assumptions of such an ideal earth are not fulfilled.

Observations of earth tides generally do not reveal the tidal deformation of the solid earth alone. Various secondary effects or indirect effects are superimposed on the primary field of the tidal deformation. They are effects of ocean tides, those of geologic and tectonic origin or those of meteorological disturbances. These effects cause regional variations of tidal parameters.

The tidal deformation of the earth is observed by various methods on the earth's surface. Observations of tidal gravity are less affected by local conditions than observations of tidal tilt and strain. Therefore, for investigations of regional variations of earth tides, observations of tidal gravity are considered superior to those of either tidal tilt or strain. Many experimental studies about regional variations of the tidal gravity parameters have been made in the past twenty years. Extensive efforts were made during the International Geophysical Year (IGY) in earth-tide observations on a worldwide basis. HARRISON *et al.* (1963) clearly showed the existence of regional variations of tidal gravity parameters and inferred that both ocean tides and regional geologic differences contribute to such parametric variations. On the other hand, KUO *et al.* (1970) established a transcontinental tidal gravity profile across the United States and demonstrated that the observed regional variations of tidal gravity parameters can be explained only by the ocean tide effects. In these studies, however, the knowledge of ocean tide effects was somewhat insufficient for detailed interpretations about the regional variations of tidal gravity parameters.

The establishment of a load deformation theory (FARRELL, 1972) and the improvement of ocean tide models (e.g. HENDERSHOTT, 1972) made it possible to estimate ocean tide effects on tidal gravity observations more accurately. These investigations have much encouraged experimental studies of gravity tides (BRETREGER and MATHER, 1978; MELCHIOR, 1978; BAKER, 1980; MELCHIOR *et al.*, 1981). They estimated contributions of body tides by assuming the δ-factor to be 1.16 uniformly over the entire earth without any phase-lag and interpreted residuals of the observed tides from the theoretical body tides as ocean tide effects. Their interpretations can be roughly accepted in the first approximation theory, but more detailed treatments have found their assumption to be inconsistent with observational results.

MELCHIOR *et al.* (1981) found that there exist good agreements between observed residuals and calculated ocean tide effects for most of the island and coastal stations but not for the many continental stations. This result may suggest the existence of effects on tidal gravity other than those of ocean tides. PARIYSKIY (1978) pointed out that the δ-factors obtained at Eurasian stations are systematically 2% smaller than those at stations in the European part of the U.S.S.R. SOURIAU (1979) examined statistically the spatial distribution of gravity and tilt tides in Europe and found the spatial trend of the δ-factors. Recently MELCHIOR and BECKER (1983) have indicated a correlation between lithosphere's thicknesses and tidal gravity parameters obtained at Alpine stations. All the above-mentioned studies suggest the possibility that tidal gravity parameters may vary regionally due to causes other than ocean tide effects. To resolve such problems, spatially wide and dense observations of earth tides are required.

Japan's main islands are located on the plate-subducting zone including a large heterogeneity in the crust-mantle structure. Therefore, it is geophysically interesting to investigate regional variations of tidal parameters across the island arc. A pioneering study was made by NAKAGAWA (1962a) in Japan. He suggested the existence of some relationship between the δ-factors and the Bouguer gravity anomalies. Afterwards tidal gravity observation of long duration have been followed up to obtain precise values of tidal gravity parameters at various stations (NAKAGAWA, 1962b; NAKAGAWA *et al.*, 1975; HOSOYAMA, 1977; SHIMADA, 1979). These studies clarified the significance of ocean tide effects in Japan. However, by these available data only we cannot draw a conclusion on the problem of regionalities of tidal gravity parameters. The point is whether their observed regionalities in Japan can be fully explained by ocean tide effects or not. The problem has still remained unresolved up to now. The present study aims at reexamining this problem with tidal gravity observations in Central Japan. There exists a complex and faulted geological structure which may cause regional heterogeneities of the earth tides.

2. Gravity Tides on the Earth

Tidal gravity variations consist of the sum of two components. In addition to the earth's body tide signal due to the direct astronomical forces there is a further signal arising from the surface loading of the ocean tides. The latter component is referred to as "load tides" in this paper. We will briefly describe the standard formulas for both body tides and load tides which are used in the interpretation of results of tidal gravity observations.

2-1 Body Tides

Body tides consist of two parts. One is the direct acceleration of the sun and moon and is derived from the tide-generating potential

$$\Psi_t = g\bar{\zeta}_t\left(\frac{r}{a}\right)^2 \tag{2-1}$$

where g is the gravitational acceleration on the earth's surface, $\bar{\zeta}_t$ the equilibrium tide expressed by the spherical harmonics of degree 2, r the distance from the center of the earth to the point of observation and a is the mean radius of the earth. If the earth were rigid, the acceleration derived from the potential Ψ_t would be observed on the earth. The other part is derived from the gravitational potential Ψ_d which arises from the tidal deformation of the earth.

Observational data of the radial displacement u_r and the potential perturbation Ψ on the earth's surface for tidal deformation are given as

$$u_r=\frac{h}{g}\Psi_t; \quad \Psi=\Psi_t+\Psi_d=(1+k)\Psi_t, \tag{2-2}$$

where h and k are the Love numbers.

An expression for gravity variations on the deformed surface of the earth can be obtained as given below. The potential variation on the deformed surface is

$$W=u_r\frac{dV_0}{dr}+\Psi, \tag{2-3}$$

where V_0 is the gravitational potential of an undeformed earth and

$$\frac{d^2V_0}{dr^2}=-\frac{2}{r}\frac{dV_0}{dr}=\frac{2g}{r}. \tag{2-4}$$

The gravity variation Δg just outside the surface $(r=a^+)$ is

$$\Delta g=\frac{\partial W}{\partial r}\bigg|_{r=a^+}=u_r\frac{d^2V_0}{dr^2}+\frac{\partial \Psi}{\partial r}\bigg|_{r=a^+}. \tag{2-5}$$

For the space outside the surface we have

$$\Psi_d=gk\bar{\zeta}_t\left(\frac{a}{r}\right)^3 \quad r\geqq a. \tag{2-6}$$

Putting eqs. (2-1), (2-2), (2-4) and (2-6) into (2-5) yields

$$\Delta g=\left(1+h-\frac{3}{2}k\right)\frac{\partial \Psi_t}{\partial r}. \tag{2-7}$$

The factor in the bracket of the eq. (2-7) is the ratio of the tidal gravity variation on the earth to that on the rigid earth and is termed as the δ-factor. It should be noted that the δ-factor takes the same

Table 1. Theoretical δ-factor.

Constituent	MOLODENSKY (1961)		SHEN and MANSINHA (1976)	WAHR (1981)	
	Model I	Model II		ϕ=35°	38°
M_2, S_2	1.160	1.165		1.154	1.153
K_1	1.136	1.142	1.138	1.135	1.133
O_1	1.159	1.164	1.160	1.155	1.153

ϕ: latitude

value on the earth's surface as far as the earth can be assumed to be spherical symmetric and non-rotating. Table 1 shows theoretical values of the δ-factors obtained by several studies. In addition to the amplitude modification expressed by the δ-factor, there exists a phase difference between the observed and the theoretical gravity tides. Phase difference is zero for the perfectly elastic earth.

As Ψ_t can be very accurately calculated by celestrial mechanics, the δ-factor and phase difference are determined by tidal gravity observations.

2-2 Load Tides

We consider gravity variations due to a surface mass load of ocean tide $\zeta(\vartheta, \phi)$. We are concerned with the term ζ_n of degree n in the surface spherical harmonic expansion of ζ. The tide ζ_n exerts a normal stress $-\rho_w g\zeta_n$ across the surface of the earth, where ρ_w denotes the density of sea water. In addition, the tide ζ_n produces a potential

$$\Psi_l=\begin{cases}\dfrac{4\pi Ga}{2n+1}\rho_w\zeta_n\left(\dfrac{a}{r}\right)^{n+1} & r\geqq a\\[2ex] \dfrac{4\pi Ga}{2n+1}\rho_w\zeta_n\left(\dfrac{r}{a}\right)^{n} & r\leqq a,\end{cases} \tag{2-8}$$

where G is the Newtonian gravitational constant. The total perturbation in potential Ψ is again given in terms of Ψ_l and Ψ_d. The latter, in terms of the load Love numbers k'_n, is written as

$$\Psi_d=k'_n\frac{4\pi Ga}{2n+1}\rho_w\zeta_n\left(\frac{a}{r}\right)^{n+1} \quad r\geqq a. \tag{2-9}$$

The load Love numbers, k'_n, h'_n will be defined by relations identical to eq. (2-2).

Using eqs. (2-8) and (2-9) in (2-5), the gravity variations Δg_n just outside the surface ($r=a^+$) for the load deformation is given by

$$\begin{aligned}\Delta g_n&=u_r\frac{d^2V_0}{dr^2}+\frac{\partial\Psi}{\partial r}\bigg|_{r=a^+}\\ &=\{2h'_n-(n+1)k'_n\}\frac{\Psi_l(a)}{a}.\end{aligned} \tag{2-10}$$

Here we exclude the direct attraction of the sea water.

The gravity variation due to $\zeta_n(\vartheta, \phi)$ is obtained by summing up eq. (2-10) for degree n as,

$$\Delta g=4\pi G\sum_{n=0}^{\infty}\{2h'_n-(n+1)k'_n\}\frac{\rho_w\zeta_n}{2n+1}. \tag{2-11}$$

Since we have

$$\frac{4\pi}{2n+1}\zeta_n(\vartheta, \phi)=\int_0^{2\pi} d\phi' \int_0^{\pi} \zeta(\vartheta', \phi')P_n(\cos\alpha)\sin\vartheta' d\vartheta'$$
$$\cos\alpha=\cos\vartheta\cos\vartheta'+\sin\vartheta\sin\vartheta'\cos(\phi-\phi'), \tag{2-12}$$

eq. (2-11) is rewritten as follows

$$\begin{aligned}\Delta g &=\rho_w a^2\int_0^{2\pi} d\phi' \int_0^{\pi} \zeta(\vartheta', \phi')g(\alpha)\sin\vartheta' d\vartheta' \\ &=\rho_w \iint_S \zeta(\vartheta', \phi')g(\alpha)dS \\ g(\alpha)&=\frac{g}{m}\sum_{n=0}^{\infty}\{2h'_n-(n+1)k'_n\}P_n(\cos\alpha),\end{aligned} \tag{2-13}$$

where m is the earth's mass.

$g(\alpha)$ is a function of α which is determined from the earth's physical properties. FARRELL (1972) calculated $g(\alpha)$ for the Gutenberg-Bullen earth model. Then if we know the distribution of ocean tides around the world, we can obtain its gravity effect by the formula (2-13).

3. Observations

3-1 Stations and Periods of Observations

Central Japan is located at the junction of the East Japan island arc and the West Japan island arc. The geological structure there, which is supposed to be complicated, is of special interest for investigating island arc systems. This region is characterized by high-altitude mountain systems (see Fig. 1) and also by a negative Bouguer gravity anomaly amounting to -80 mgals at maximum. This negative anomaly is considered as an isostatic balance of terrain masses. Seismic explosions and surface wave studies also support the existence of a concave surface of the Mohorovičić discontinuity. Isostasy is undoubtedly prevailing in this region but a detailed underground structure in Central Japan has never completely been clarified. Recently FUKAO and YAMAOKA (1983) have insisted that the compensation inferred from the gravity anomalies is not large enough to support the surface topography.

The geological structure of Central Japan is supposed to be anomalous since it is located in the arc-arc junction. Therefore Central Japan, if any, is one of the most promising regions for discovering a lateral heterogeneity of earth-tidal parameters. As shown in Fig. 1, we have chosen locations of gravity stations so as to cover this region as uniformly as possible. The position coordinates of the stations and their detailes are given in Table 2. The basement or observation vault has

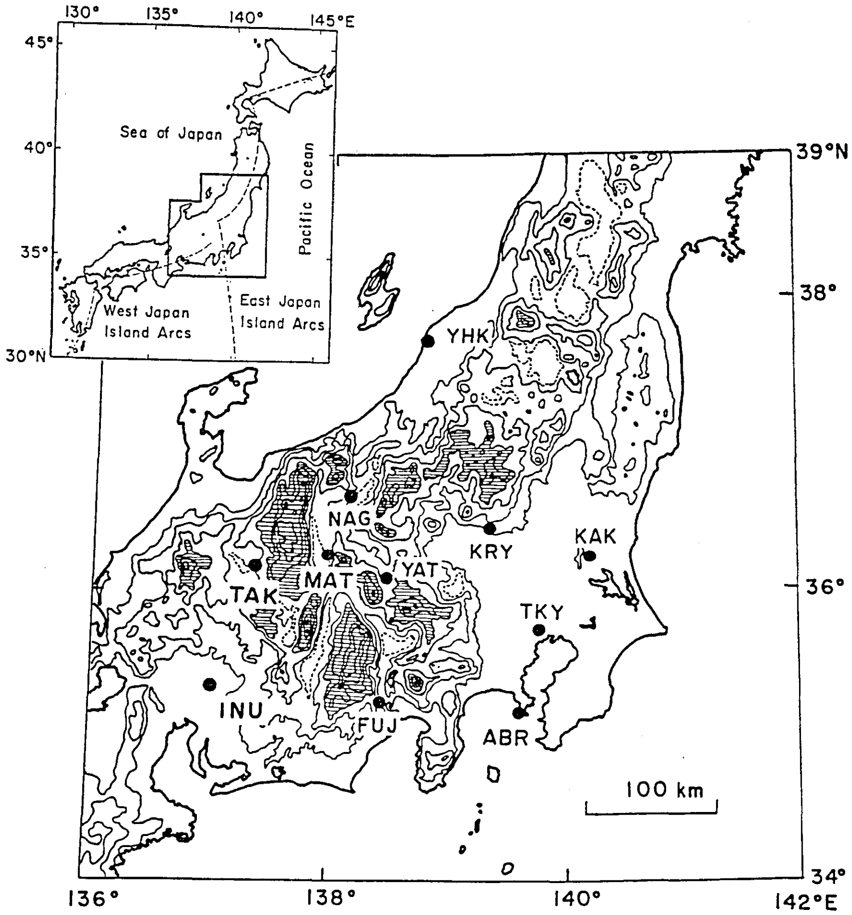


Fig. 1. Locations of tidal gravity stations. Surface topography is shown with a contour interval of 400 m. Hatching indicates areas higher than 1600 m above sea level. TKY: Tokyo, FUJ: Fujigawa, YHK: Yahiko, YAT: Yatsugatake, NAG: Nagano, KRY: Kiryu, KAK: Kakioka, INU: Inuyama, MAT: Matsumoto, TAK: Takayama, ABR: Aburatsubo.

been selected as an observation site. At the Aburatsubo station (ABR) observations have been made by SHIMADA (1979). We will include his results in this paper. Observation periods are three to four months for three stations (TKY, FUJ and YAT), but about five to six months for the other stations (ABR, YHK, NAG, KRY, KAK, INU, MAT and TAK) as shown in Table 3.

3-2 General Situations of Observations

The tidal gravity observations have been carried out with two LaCoste & Romberg gravimeters (G34 and G447). The gravimeter has been installed on the concrete block or the concrete floor of the room.

Table 2. General description of tidal gravity stations.

Station	Abbreviations	Location	Height above sea level (m)	Gravimeter	Remarks
Tokyo	TKY	35.716°N 139.763°E	20	G 34 G447	Earthquake Research Institute, underground room
Fujigawa	FUJ	35.228°N 138.420°E	640	G 34	Fujigawa Crustal Movement Observatory, recording house
Yahiko	YHK	37.733°N 138.801°E	30	G 34	Yahiko Crustal Movement Observatory, observation vault
Yatsugatake	YAT	36.094°N 138.486°E	850	G 34	Yatsugatake Geomagnetic Observatory, basement
Nagano	NAG	36.665°N 138.186°E	490	G 34	Hokushin Microearthquake and Crustal Movement Observatory, basement
Kiryu	KRY	36.419°N 139.352°E	120	G 34	Gumma University, Faculty of Technology, basement
Kakioka	KAK	36.228°N 140.194°E	30	G447	Geophysics Research Laboratory, University of Tokyo, basement
Inuyama	INU	35.349°N 137.028°E	134	G447	Inuyama Crustal Movement Observatory, observation vault
Matsumoto	MAT	36.248°E 137.981°N	619	G 34	Shinsyu University, Faculty of Science, basement
Takayama	TAK	36.133°N 137.184°E	700	G447	Takayama Seismological Observatory, observation vault
Aburatsubo	ABR	35.156°N 139.616°E	6.9	G447	Aburatsubo Crustal Movement Obsevatory, observation vault

Table 3. Observation periods.

Period	G34		G447	
	Station	Observation period	Station	Observation period
I	Tokyo	Mar. 30, 1980–Jul. 23, 1980	Tokyo	Mar. 30, 1980–Jul. 23, 1980
II	Fujigawa	Aug. 22, 1980–Dec. 2, 1980	—	
III	Yahiko	Dec. 9, 1980–May 8, 1981	Tokyo	Dec. 28, 1980–May 8, 1981
IV	Yatsugatake	May 9, 1981–Aug. 26, 1981	Tokyo	May 9, 1981–Aug. 26, 1981
V	Tokyo	Aug. 29, 1981–Nov. 18, 1981	Tokyo	Aug. 29, 1981–Nov. 18, 1981
VI	Nagano	Mar. 24, 1982–Sep. 7, 1982	—	
VII	Kiryu	Sep. 8, 1982–Feb. 9, 1983	Kakioka	Aug. 28, 1982–Feb. 14, 1983
VIII	Matsumoto	Mar. 29, 1983–Oct. 3, 1983	Inuyama	Mar. 15, 1983–Sep. 9, 1983
IX			Tokyo*	Jan. 22, 1984–Mar. 19, 1984
X			Takayama*	Apr. 25, 1984–Jul. 27, 1984 Sep. 28, 1984–Nov. 28, 1984

* The gravimeter was implemented with a electrostatic feedback.

The mechanical sensitivity of the gravimeter has been adjusted according to the manufacturer's manual. Therefore the offset angle from the astatic point has become about 100″. However, for the observation by the gravimeter G447 at the Aburatsubo station, offset angle from the astatic point has been adjusted to 240″ (SHIMADA, 1979). Because the instrumental phase lag depends on the offset angle, the result of phase differences at the Aburatsubo station cannot be treated in the same way as that of the other stations by the gravimeter G447. The levels of the gravimeter have been settled in the position corresponding to that of minimum sensitivity to tilting because the gravimeter output exhibits a parabolic dependence on the extent of tilting.

The output signal from the gravimeter has been recorded on a chart with a feeding speed of 25 mm $hour^{-1}$ through an active low-pass filter with a time constant of 100 seconds. Temperature and pressure at the observation point have also been measured simultaneously. Gravity variations have been recorded with a normal sensitivity of 0.3 mm μgal^{-1} and digitized for every thirty minutes by using a X-Y digitizer with a resolution of 0.1 mm. Fig. 2 is a sample of the original data obtained at the Fujigawa station.

During observations at the Matsumoto station, the gravimeter G34 showed anomalous behavior several times. We therefore exclude the Matsumoto's data from analyses. In January of 1984, we have implemented the electrostatic feedback with the gravimeter G447 after the method of HARRISON and SATO (1984). The implementation improves a stability of instrumental sensitivities and greatly reduces an instrumental phase lag. The method will be briefly described in the Appendix. By this feedback modified gravimeter observations have been made at both the Tokyo and the Takayama station.

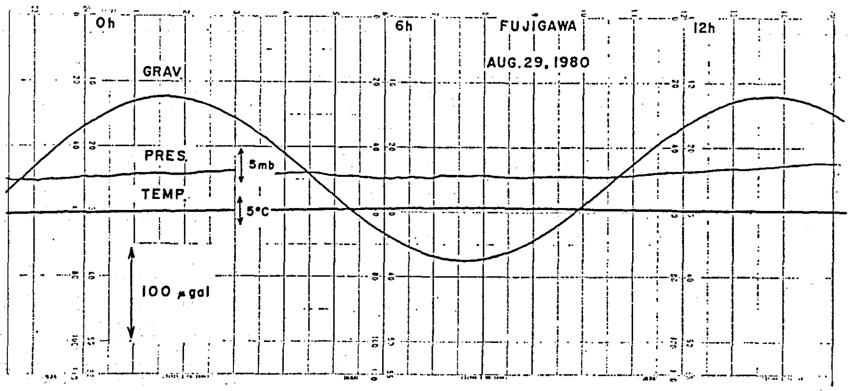


Fig. 2. A sample of the original data. GRAV.: Gravity, PRES.: Pressure, TEMP.: Temperature.

4. Data Analyses

4-1 Pre-processing Method

Due to the astatization of a spring system, the mechanical sensitivity of a LaCoste & Romberg gravimeter is very sensitive to its tilting. It has been pointed out that tidal observations by a LaCoste & Romberg gravimeter are largely affected by short-term sensitivity changes which occur possibly due to its tilting. We cannot follow such changes by occasional experimental calibrations. Such changes might cause systematic errors of tidal gravity parameters if they are not corrected properly.

In order to estimate gravimetric sensitivity changes, we apply here the pre-processing method introduced by NAKAI (1977). In this method, tidal gravity data of a few-day's duration are fitted to the following formula with time t.

$$Y(t)=\frac{1}{S}R(t)+\frac{\Delta t}{S}\frac{dR}{dt}+\sum_j k_j t^j, \qquad (4\text{-}1)$$

where $Y(t)$ is the reading value of gravity; S the relative sensitivity; $R(t)$ the theoretical tide that would be observed on a rigid earth without oceans; Δt the mean time lag and k_j the coefficient of j-th order of the drift polynomial. In eq. (4-1) unknown parameters are S, Δt and k_j. These unknown parameters can be determined by the least squares method if the order of the polynomial is assumed and the appropriate number of the data are available. In the present analysis, we assume a cubic polynomial, that is $j=0$, 1, 2 and 3, and determine daily values of the above unknown parameters by using the corresponding three-days' data set. By applying this procedure successively, we finally obtain daily variations of the unknown parameters throughout the observation period.

On the other hand, the gravimetric sensitivity has been experimentally determined each month. In principle, alternate deflections of the gravimeter-beam which correspond to approximately 300 μgals have been produced by turning the measuring dial in the direction of both increasing and decreasing values by 30 dial scale divisions. This operation has been repeated almost every hour during several hours of calibration. The gravimeter-beam may not reach its exact equilibrium even after one hour. However, relative accuracy will be retained so long as calibrations are made in the same way at each observation station. Therefore we can compare tidal gravity parameters at the different stations. As far as tidal gravity parameters are time-invariant, a relative variation of a calculated sensitivity in the pre-processing

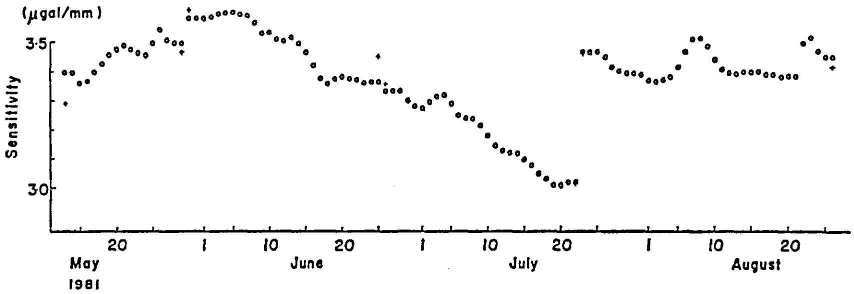


Fig. 3. Variations in gravimetric sensitivity (open circles) obtained by applying Nakai's pre-processing method for the record at the Yatsugatake station (YAT). Crosses (+) indicate sensitivity obtained by the experimental calibration.

method must agree with an experimentally determined one. That is to say, a calculated sensitivity must be proportional to a measured one. On such an assumption, we can determine the mean ratio of the measured sensitivity to the calculated one as f_0, and its standard deviation as $f_0\Delta\varepsilon$ by the least squares method. $\Delta\varepsilon$ can then be regarded as a relative error of calibrations and Sf_0 becomes a gravimetric sensitivity at an arbitrary epoch through the observation period.

For example, let us determine gravimetric sensitivities for the gravity data at the Yatsugatake station (YAT). By the above-described procedure, the mean ratio f_0 was obtained as 1.220 with a standard deviation of $f_0\Delta\varepsilon=0.008$. The time variation of Sf_0 is shown in Fig. 3. We see in this figure that such estimated sensitivities (open circles) agree well with the ones determined experimentally (crosses). A large offset at July 23, 1981 was caused by re-leveling of the gravimeter.

4-2 Venedikov's Method

We use the Venedikov least squares method (VENEDIKOV, 1966) to obtain tidal gravity parameters. We will give an outline of the method in the following text. VENEDIKOV (1966), MELCHIOR and VENEDIKOV (1968) devised the numerical filters which eliminate the drift components and at the same time separate diurnal, semi-diurnal and ter-diurnal tides from each other.

The hourly reading values of earth tides at epoch T_i+t are expressed as follows:

$$Y(T_i+t)=\sum_j H_j \cos(\omega_j t+\phi_j(T_i))+D(T_i+t), \qquad (4\text{-}2)$$

where H_j is the observed amplitude of constituent of frequency ω_j, ϕ_j is the observed phase of the same constituent at epoch T_i and $D(T_i+t)$ is the drift components at epoch T_i+t. For each tidal frequency band, the following two filters are applied on a sequence of 48 hourly readings. One is an even filter (C) and the other is an odd filter (S). The diurnal filters $C^{(1)}$, $S^{(1)}$ are constructed to amplify the diurnal constituents and eliminate the semi-diurnal ones. An opposite property is obtained with the semi-diurnal filters $C^{(2)}$, $S^{(2)}$. The amplification factors of those filters can be obtained by:

$$\left.\begin{aligned} c^{(\tau)} &= \sum_{t=-23.5}^{+23.5} C^{(\tau)} \cos \omega_j t, \\ s^{(\tau)} &= \sum_{t=-23.5}^{+23.5} S^{(\tau)} \sin \omega_j t, \end{aligned}\right\} \quad \tau=1,\ 2. \tag{4-3}$$

The filters eliminate the drift components as an arbitrary combination of Chebychef polynomials of order 3 for C filters and of order 2 for S filters.

Both filters (C and S) are applied to every available recording interval covering 48 hourly readings without any gap in order to obtain the two values, M_i and N_i (i being the number of the recording intervals). Using the eq. (4–2), for every central epoch T_i of 48 hours we can write

$$\left.\begin{aligned} M_i &= \sum_{t=-23.5}^{+23.5} Y(T_i+t)C_t = \sum_j c_j H_j \cos \phi_j(T_i), \\ N_i &= \sum_{t=-23.5}^{+23.5} Y(T_i+t)S_t = -\sum_j s_j H_j \sin \phi_j(T_i), \end{aligned}\right\} \quad i=1,\ 2,\ \cdots,\ n. \tag{4-4}$$

Here we assume the $D(T_i+t)$ in the eq. (4–2) is completely eliminated by the filters.

From geophysical considerations we assume that tidal parameters are equal for constituents having very near frequencies within the same group, with the exception of K_1 group. In the frequency band of K_1 group exist variations of tidal parameters associated with liquid-core resonance effects. However data of longer duration (more than 1 year) than those of the present observation are required to obtain the frequency dependence of tidal parameters. Therefore we assume the same tidal parameters also within K_1 group.

Venedikov distributes the tidal constituents into p groups and, denoting by α_k and β_k the index number of the first and the last constituent of one group he writes:

$$\left.\begin{aligned} & H_j = \delta_k h_j, \\ & \phi_j(T_i) = \varphi_j(T_i) + \kappa_k, \\ & \xi_k = \delta_k \cos \kappa_k, \\ & \eta_k = -\delta_k \sin \kappa_k, \\ & \alpha_k \leqq j \leqq \beta_k, \end{aligned}\right\} \quad k=1,\ 2,\ \cdots,\ p, \tag{4-5}$$

where $\varphi_j(T_i)$ is the theoretical phase of constituent j for the epoch T_i and h_j is the theoretical amplitude of the same constituent.

Then one obtains a system of $2n$ equations $(i=1, 2, \cdots, n)$ with $2p$ unknowns $(k=1, 2, \cdots, p)$:

$$
\begin{aligned}
M_i&=\sum_{k=1}^{p}\left(\xi_k \sum_{j=\alpha_k}^{\beta_k} c_j h_j \cos\varphi_j(T_i)+\eta_k \sum_{j=\alpha_k}^{\beta_k} c_j h_j \sin\varphi_j(T_i)\right),\\
N_i&=\sum_{k=1}^{p}\left(-\xi_k \sum_{j=\alpha_k}^{\beta_k} s_j h_j \sin\varphi_j(T_i)+\eta_k \sum_{j=\alpha_k}^{\beta_k} s_j h_i \cos\varphi_j(T_i)\right),
\end{aligned}
\tag{4-6}
$$

with $2n>2p$.

The least squares method is applied to obtain unknowns, δ_k, κ_k $(k=1, 2, \cdots, p)$.

$$\delta_k=(\xi_k^2+\eta_k^2)^{1/2}, \qquad \kappa_k=\tan^{-1}(-\eta_k/\xi_k). \tag{4-7}$$

The associated errors can be calculated from the unit weight error given by the least squares method and the diagonal elements of the error matrix.

4-3 Results

Results of the analyses are presented in Tables 4a, 4b and 4c for the gravimeters G34, G447, and modified G447 respectively. The δ-factor is the ratio of an observed amplitude of the constituent to the amplitude that would be observed on a rigid earth without oceans. The standard error of the δ-factor is estimated by the following formula.

$$\sigma_\delta=\delta((\Delta\delta/\delta)^2+(\Delta\varepsilon)^2)^{1/2}, \tag{4-8}$$

Table 4a. δ-factors (δ) and phase differences (κ) obtained by the gravimeter G34.

Station	M_2		S_2		K_1		O_1	
	δ	κ (degree)	δ	κ (degree)	δ	κ (degree)	δ	κ (degree)
Tokyo*	1.257	−0.70	1.273	−1.16	1.265	−0.99	1.278	−0.18
	±0.005	±0.06	±0.005	±0.11	±0.006	±0.12	±0.006	±0.16
Fujigawa	1.261	−0.93	1.296	−1.84	1.266	−3.30	1.289	−0.97
	±0.010	±0.09	±0.011	±0.15	±0.012	±0.23	±0.012	±0.27
Yahiko	1.251	−0.54	1.258	−1.15	1.246	−1.71	1.267	−0.62
	±0.005	±0.05	±0.006	±0.09	±0.005	±0.09	±0.006	±0.12
Yatsugatake	1.236	−1.08	1 231	−0.86	1.242	−0.40	1.240	+0.17
	±0.010	±0.04	±0 010	±0.11	±0.010	±0.14	±0.011	±0.24
Nagano	1.229	−1.22	1.231	−1.65	1.230	−1.72	1.228	−2.17
	±0.006	±0.10	±0.008	±0.24	±0.007	±0.21	±0.010	±0.36
Kiryu	1.237	−0.67	1.258	−0.32	1.265	−2.33	1.265	−0.73
	±0.009	±0.09	±0.010	±0.17	±0.010	±0.15	±0.010	±0.23

* The mean values of the results of two analyses (periods I and V in Table 6).

Table 4b. δ-factors (δ) and phase differences (κ) obtained by the gravimeter G447.

Station	M_2		S_2		K_1		O_1	
	δ	κ (degree)	δ	κ (degree)	δ	κ (degree)	δ	κ (degree)
Tokyo*	1.215	−1.04	1.227	−1.05	1.221	−1.60	1.242	−0.42
	±0.004	±0.04	±0.005	±0.08	±0.004	±0.09	±0.005	±0.13
Kakioka	1.209	−0.70	1.238	−0.47	1.228	−1.22	1.228	+0.22
	±0.011	±0.07	±0.011	±0.15	±0.012	±0.18	±0.013	±0.27
Inuyama	1.212	−1.31	1.219	−1.96	1.203	−1.59	1.232	−1.08
	±0.004	±0.03	±0.004	±0.06	±0.004	±0.05	±0.004	±0.06
Aburatsubo	1.246	0.70	1.254	−0.15	1.229	−0.63	1.271	+0.72
	±0.002	±0.09	±0.004	±0.20	±0.004	±0.17	±0.006	±0.26

* The mean values of the results of two analyses (periods I and V in Table 6).

Table 4c. δ-factors (δ) and phase difference (κ) by the modified gravimeter G447.

Station	M_2		S_2		K_1		O_1	
	δ	κ (degree)	δ	κ (degree)	δ	κ (degree)	δ	κ (degree)
Tokyo	1.202	0.58	1.213	−0.38	1.208	−0.52	1.217	0.88
	±0.005	±0.08	±0.006	±0.15	±0.006	±0.12	±0.006	±0.17
Takayama	1.184	0.08	1.189	−0.87	1.166	−0.03	1.188	0.32
	±0.002	±0.03	±0.002	±0.06	±0.002	±0.03	±0.002	±0.05

where $\Delta\delta$ is the error obtained from the Venedikov harmonic analysis. The above formula combines an error estimate from the harmonic analysis with that ($\Delta\varepsilon$, see 4–1) from the processing method. The phase differences in Tables 4a, 4b and 4c are given with respect to the equilibrium tide in the local meridian (phase lead is taken as positive). The standard error or the phase difference is obtained from the Venedikov harmonic analysis.

5. Effects of Ocean Tides

Tidal gravity variations are greatly influenced by ocean tides. An accurate estimation of earth tides depends on the method of correcting ocean tide effects on the gravity field. For calculating ocean tide effects the entire ocean area is divided into three zones according to the distance from tidal gravity stations: a global ocean area of an angular distance $\vartheta \gtrsim 15°$, a sea surface area around the Japan islands ($1 \lesssim \vartheta \lesssim 15°$) and the zone adjacent to the gravity stations ($\vartheta \lesssim 1°$).

For tidal data in the global ocean the numerical ocean tidal models of PARKE and HENDERSHOTT (1980) and SCHWIDERSKI (1980) are employed.

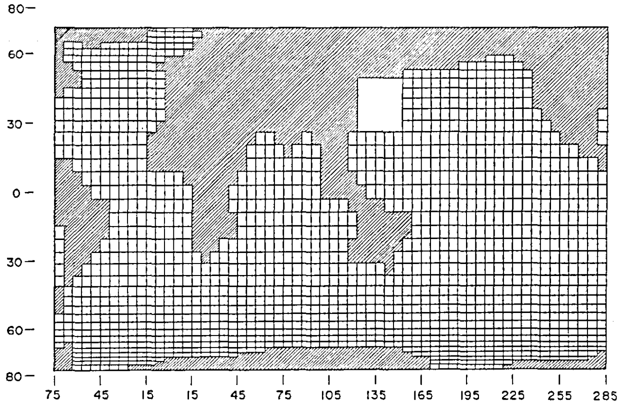


Fig. 4. Divisions of the global ocean for the Parke-Hendershott model used for calculating ocean tide effects.

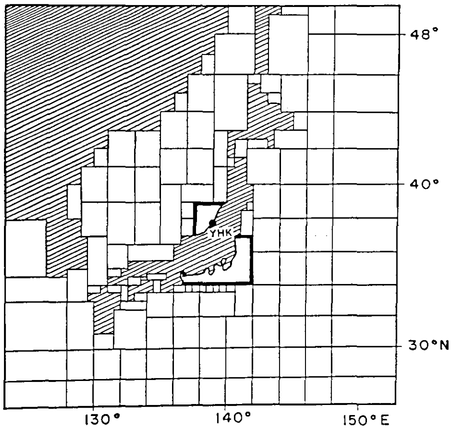


Fig. 5. Divisions of the sea around Japan for calculating ocean tide effects. Finer grids are used inside the bold lines.

Both the models were originally constructed to correct satellite-altimetry data for tidal variations in sea surface elevations. They are considered to be more accurate than any other ocean tide models available at present. The Parke-Hendershott model gives amplitudes and phases of ocean tides at 6° (latitude) $\times$ 6° (longitude) grid points and the Schwiderski model at $1° \times 1°$ grid points in digital forms. These digital data can be directly used for the present calculations. Fig. 4 shows grids for calculations by the Parke-Hendershott model.

A cotidal chart in the sea around Japan was given by TSUKAMOTO and NAKAGAWA (1978, 1980). In an area of $1° \lesssim \vartheta \lesssim 15°$, we adopt their chart for the present calculations by digitizing amplitudes and phases of ocean tides at $2° \times 2°$ grid points. Fig. 5 shows grids for calculations in the sea around Japan. A local cotidal chart in the sea adjacent to the gravity stations is drawn by taking tidal constants of the coastal tide-gauge stations into account (MIYAZAKI *et al.*, 1967). In the case of $\vartheta \lesssim 1°$, we use finer grids according to the distance from the gravity stations. Fig. 6 shows such subdivisions of the grids for the Yahiko

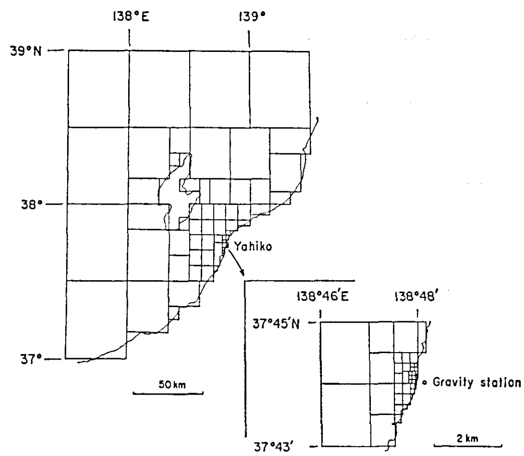


Fig. 6. Divisions of the sea near the Yahiko station (YHK) for calculating ocean tide effects.

Table 5a. Calculated ocean tide effects on gravity for the M_2 constituent.

Station	Model	Global ocean		Around Japan		Adjacent to the station		Total	
		Amplitude (μgal)	Phase (degree)	Amplitude (μgal)	Phase (degree)	Amplitude (μgal)	Phase (degree)	Amplitude (μgal)	Phase (degree)
Tokyo	PK*	0.56	16.32	2.16	27.81	0.04	12.38	2.75	25.26
	SW**	0.36	331.89					2.42	20.48
Fujigawa	PK	0.60	12.45	2.35	24.55	0.25	9.07	3.18	21.08
	SW	0.40	330.72					2.85	16.70
Yahiko	PK	0.45	27.67	1.65	39.35	0.23	82.40	2.26	41.02
	SW	0.27	334.14					1.93	36.74
Yatsugatake	PK	0.54	16.91	2.04	29.22	—	—	2.57	26.65
	SW	0.35	331.59					2.25	21.66
Nagano	PK	0.52	18.45	1.81	32.18	0.02	83.63	2.33	29.53
	SW	0.33	331.30					2.00	24.35
Kiryu	PK	0.50	23.22	2.02	33.40	—	—	2.51	31.39
	SW	0.30	334.70					2.19	26.68
Kakioka	PK	0.50	27.35	2.14	34.43	0.15	46.28	2.78	33.80
	SW	0.27	338.52					2.45	29.91
Inuyama	PK	0.64	8.33	2.20	22.52	0.39	355.94	3.19	16.55
	SW	0.45	328.90					2.87	11.73
Takayama	PK	0.58	11.96	1.90	27.24	0.01	85.63	2.47	23.89
	SW	0.40	329.32					2.14	18.37
Abratsubo	PK	0.57	17.72	2.39	27.48	1.33	23.99	4.28	25.10
	SW	0.35	333.82					3.94	22.20

* PK denotes the Parke-Hendershott model.

** SW denotes the Schwiderski model.

Table 5b. Calculated ocean tide effects on gravity for the S_2 constituent.

Station	Model	Global ocean		Around Japan		Adjacent to the station		Total	
		Amplitude (μgal)	Phase (degree)	Amplitude (μgal)	Phase (degree)	Amplitude (μgal)	Phase (degree)	Amplitude (μgal)	Phase (degree)
Tokyo	PK*	0.31	35.64	0.92	359.61	0.02	346.47	1.20	8.11
	SW**	0.20	353.21					1.14	358.26
Fujigawa	PK	0.31	33.56	1.00	356.69	0.12	342.65	1.37	3.25
	SW	0.20	349.86					1.32	354.39
Yahiko	PK	0.31	39.47	0.70	9.78	0.08	52.40	1.05	21.20
	SW	0.18	0.72					0.94	11.36
Yatsugatake	PK	0.31	35.89	0.87	0.87	—	—	1.14	9.86
	SW	0.19	353.77					1.06	359.60
Nagano	PK	0.30	36.62	0.76	3.63	0.01	53.63	1.03	13.17
	SW	0.19	355.13					0.95	2.40
Kiryu	PK	0.32	38.19	0.87	4.48	—	—	1.15	13.36
	SW	0.19	357.98					1.06	3.32
Kakioka	PK	0.34	39.28	0.94	5.12	0.08	16.18	1.32	14.14
	SW	0.20	0.27					1.22	5.05
Inuyama	PK	0.29	30.48	0.91	354.63	0.18	331.42	1.31	358.95
	SW	0.20	346.05					1.28	350.10
Takayama	PK	0.29	33.28	0.78	359.17	—	—	1.03	8.23
	SW	0.19	349.96					0.97	357.37
Abratsubo	PK	0.33	32.27	1.03	359.56	0.63	354.62	1.93	3.81
	SW	0.20	354.10					1.86	357.30

Table 5c. Calculated ocean tide effects on gravity for the K_1 constituent.

Station	Model	Global ocean		Around Japan		Adjacent to the station		Total	
		Amplitude (μgal)	Phase (degree)	Amplitude (μgal)	Phase (degree)	Amplitude (μgal)	Phase (degree)	Amplitude (μgal)	Phase (degree)
Tokyo	PK*	0.89	10.68	1.52	354.41	0.02	353.88	2.41	0.35
	SW**	0.63	7.12					2.16	358.09
Fujigawa	PK	0.90	9.17	1.61	353.60	0.15	353.63	2.64	358.86
	SW	0.62	4.72					2.37	356.49
Yahiko	PK	0.83	16.22	1.14	356.21	0.19	191.20	1.75	3.91
	SW	0.61	14.51					1.54	1.51
Yatsugatake	PK	0.87	11.54	1.41	354.30	—		2.26	0.86
	SW	0.62	8.15					2.02	358.52
Nagano	PK	0.84	12.94	1.23	354.21	0.02	191.81	2.02	1.70
	SW	0.60	9.90					1.80	359.20
Kiryu	PK	0.88	12.94	1.47	355.96	—		2.33	2.30
	SW	0.64	10.55					2.10	0.37
Kakioka	PK	0.91	12.93	1.61	356.69	0.11	4.08	2.61	2.61
	SW	0.66	10.98					2.37	0.98
Inuyama	PK	0.86	8.72	1.42	351.91	0.15	352.97	2.41	357.91
	SW	0.59	3.11					2.15	355.04
Takayama	PK	0.83	10.92	1.23	352.27	0.01	192.82	2.02	359.70
	SW	0.58	6.46					1.79	356.72
Abratsubo	PK	0.93	9.69	1.73	354.83	0.86	356.38	3.50	359.12
	SW	0.65	6.23					3.23	357.52

* PK denotes the Parke-Hendershott model.

** SW denotes the Schwiderski model.

Table 5d. Calculated ocean tide effects on gravity for the O_1 constituent.

Station	Model	Global ocean		Around Japan		Adjacent to the station		Total	
		Amplitude (μgal)	Phase (degree)	Amplitude (μgal)	Phase (degree)	Amplitude (μgal)	Phase (degree)	Amplitude (μgal)	Phase (degree)
Tokyo	PK*	0.67	26.48	1.19	15.65	0.02	15.21	1.87	19.50
	SW**	0.48	26.44					1.68	18.70
Fujigawa	PK	0.68	24.98	1.26	14.95	0.11	13.60	2.04	18.20
	SW	0.48	23.92					1.85	17.19
Yahiko	PK	0.63	31.89	0.90	16.82	0.19	216.20	1.33	21.15
	SW	0.48	33.60					1.18	20.48
Yatsugatake	PK	0.66	27.31	1.10	15.41	—	—	1.75	19.87
	SW	0.48	27.33					1.57	19.02
Nagano	PK	0.64	28.66	0.96	15.09	0.02	216.81	1.57	20.31
	SW	0.47	28.91					1.40	19.38
Kiryu	PK	0.67	28.79	1.15	17.06	—	—	1.81	21.34
	SW	0.49	30.02					1.63	20.92
Kakioka	PK	0.69	28.73	1.26	18.06	0.09	23.39	2.03	21.90
	SW	0.52	30.73					1.86	21.83
Inuyama	PK	0.65	24.46	1.11	13.10	0.11	12.97	1.86	17.04
	SW	0.45	21.61					1.67	15.38
Takayama	PK	0.63	26.63	0.96	13.24	—	—	1.58	18.54
	SW	0.45	25.04					1.40	17.00
Abratsubo	PK	0.70	25.55	1.35	16.47	0.67	17.80	2.71	19.13
	SW	0.50	25.94					2.51	18.70

* PK denotes the Parke-Hendershott model.
** SW denotes the Schwiderski model.

station (YHK) which is located only 300 m from the coast.

The method used for the present calculations is basically the same as that described by FARRELL (1972). The contribution due to the attraction of sea water is calculated by assuming water masses as rectangular prisms for an ocean area within 100 km of the station and as point masses for the remote zone. The contributions of the earth by the tidal loading is calculated by using Farrell's load Green function of the Gutenberg-Bullen A earth model. The results of the calculations are listed in Tables 5a, 5b, 5c and 5d. Fig. 7 shows the amplitudes and phases of ocean tide effects calculated by using the Schwiderski model for M_2 constituent at the observation stations. The phase is referred to the local meridian with a positive value indicating a lead relative to the equilibrium tide. We observe in Fig. 7 that the Aburatsubo station (ABR) has a maximum ocean tide effect of about 4 μgals and the Yahiko station (YHK) a minimum of about 2 μgals for M_2 constituent. The phases of ocean tide effects for M_2 constituent show a lead relative to the equilibrium tide at all the stations. Ocean

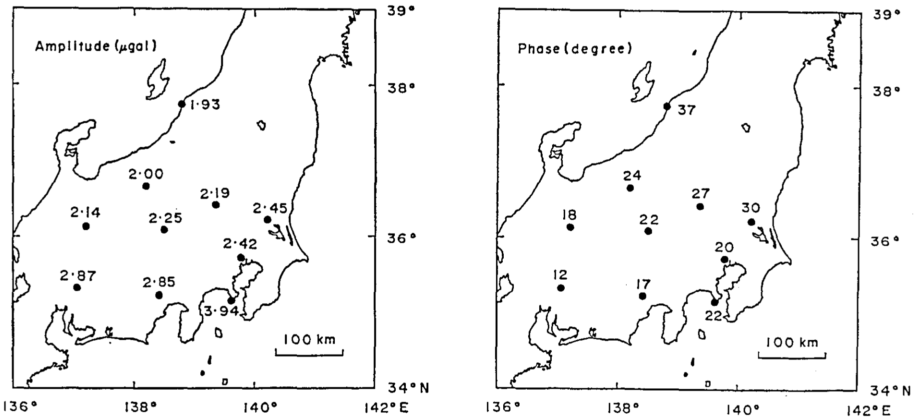


Fig. 7. Ocean tide effects for M_2 constituent in Central Japan calculated by using the Schwiderski model. Phase is referred to the local meridian.

tide effects for the other constituents also decrease gradually from the Pacific coast to the Japan sea coast.

Fig. 8 shows the sum of ocean tide effects of four major constituents at the Yahiko station as an example. Theoretical gravity tides of the solid earth at the station are also shown in Fig. 8 as a reference. Theoretical tides are calculated by assuming the δ-factor to be 1.2 without any phase difference. We observe that the ocean tide effect at the station amounts to about 5% of the earth tide and the former shows a slight phase lead relative to the latter.

The results obtained from the Parke-Hendershott model for four major constituents show that the total effects are systematically about

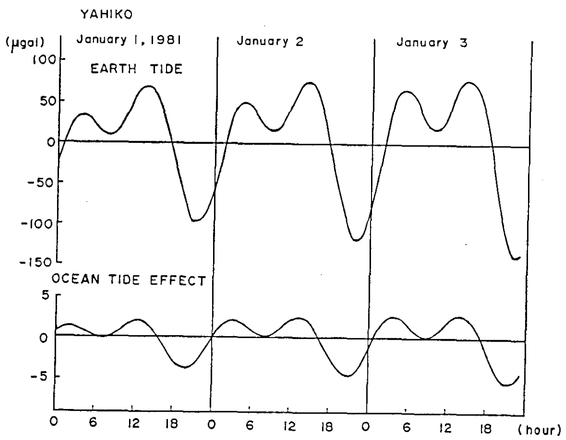


Fig. 8. Theoretical gravity tides (upper) and ocean tide effects on gravity calculated by the Schwiderski model (lower) at the Yahiko station (YHK).

10% larger than those obtained from the Schwiderski model. However, station-to-station differences in ocean tide effects for four constituents are almost the same within $5\times10^{-2}\,\mu$gal for both models. Such an agreement is quite natural because the distribution of the gravity stations is limited to an area within a few hundred kilometers. Therefore, the choice of global ocean tide models is not important for the present discussion. Tidal gravity parameters corrected with the Schwiderski model will be used in the following discussions.

6. Tidal Gravity Parameters at the Tokyo Station

6-1 Instrumental Problems

In Tables 4a and 4b, the results at the Tokyo station are given as mean values for two data series (periods I and V). The results for each data series are given in Tables 6a and 6b, respectively. We see that the δ-factors obtained by G34 are systematically about 3.5% larger than those obtained by G447 for both the periods. The δ-factors obtained by the same gravimeter agree with each other with observational errors for both the periods. The relation between phase differences obtained by G34 and G447 is not as clear as that for the δ-factors. In period I, G447 shows 0.42° phase lag relative to G34 for M_2 constituent and 0.47° phase lead for O_1 one. In period V, G447 shows 0.27° phase lag

Table 6a. δ-factors (δ) and phase differences (κ) obtained from the simultaneous observations at the Tokyo station for period I.

Constituent	G34		G447	
	δ	κ (degree)	δ	κ (degree)
M_2	1.260±0.008	−0.81±0.06	1.215±0.006	−1.23±0.06
S_2	1.274±0.008	−0.75±0.13	1.222±0.007	−1.07±0.13
K_1	1.262±0.008	−0.77±0.10	1.219±0.007	−1.61±0.14
O_1	1.280±0.009	−0.43±0.17	1.241±0.008	+0.04±0.23

Table 6b. δ-factors (δ) and phase differences (κ) obtained from the simultaneous observations at the Tokyo station for period V.

Constituent	G34		G447	
	δ	κ (degree)	δ	κ (degree)
M_2	1.254±0.007	−0.58±0.10	1.215±0.005	−0.85±0.04
S_2	1.272±0.007	−1.57±0.17	1.231±0.008	−1.02±0.08
K_1	1.267±0.008	−1.21±0.22	1.225±0.005	−1.59±0.11
O_1	1.276±0.009	+0.08±0.27	1.243±0.006	−0.87±0.13

relative to G34 for M_2 constituent and 0.95° phase lag for O_1 one. In mean values of the results for both the periods, G447 shows about 0.3° phase lag relative to G34 for M_2 constituent and 0.24° phase lag for O_1 one. These differences originate from different characteristics of the gravimeters employed. Several experiments have pointed out these kinds of differences among LaCoste & Romberg gravimeters (TSUBOKAWA *et al.*, 1977). One of the possible causes is suspected to be errors of scale value of the gravimeter or different rheological properties of its spring systems, but definitive conclusions have not yet been obtained. We must consider these systematic errors in an interpretation of the observational results.

6-2 Temporal Variations of Tidal Gravity Parameters

The present observations have been performed by the two gravimeters, G34 and G447, transported from station to station. In order to discuss the regional variations of tidal gravity parameters, we assume that the tidal parameters are time-invariant. To check this assumption, we have carried out the following test analyses.

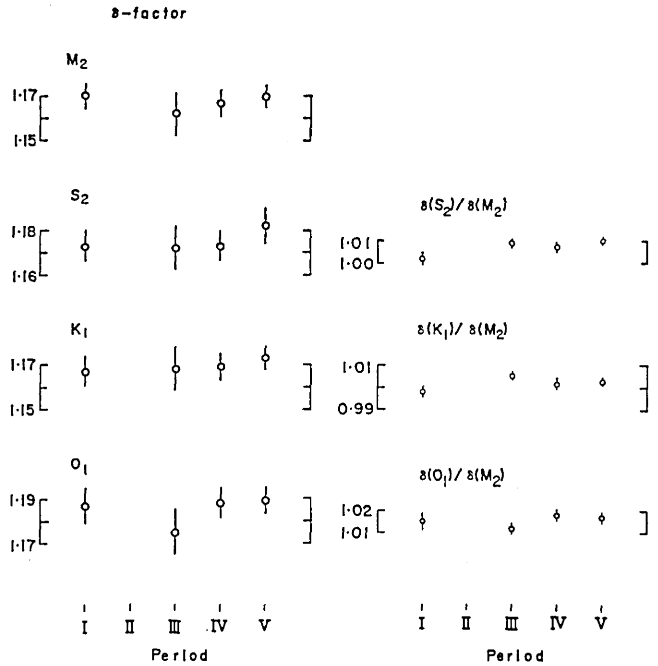


Fig. 9. δ-factors corrected for ocean tides at the Tokyo station in different observation periods, together with the ratios of the δ-factors for S_2, K_1 and O_1 constituents to that for M_2 constituent. Error bars estimated in the present analyses are also shown.

We analyze four series of gravity records obtained continuously by G447 at the Tokyo station during 1980 and 1981. The δ-factors obtained after the ocean tide corrections are plotted with standard error bars in Fig. 9. The ratios of the δ-factors of S_2, K_1 and O_1 constituents to that of M_2 constituent, i.e. $\delta(S_2)/\delta(M_2)$, $\delta(K_1)/\delta(M_2)$ and $\delta(O_1)/\delta(M_2)$, are also presented in Fig. 9. They are considered to be quantities independent of instrumental sensitivities. As can be seen from these results, the δ-factors of M_2 and O_1 constituents seem to decrease during period III. The deviations reach 0.7% of the mean value for M_2 and 1.0% for O_1 constituent in these cases. However, we can see that the estimated standard errors for period III are also larger than these for the other periods. In addition, we observe that the change in the ratio of the δ-factors has no clear correlation with that in the δ-factor itself and is smaller than 1%. Then we consider that the change of the δ-factors during period III is spurious and unlikely to be significant. As we do not find a clear indication of temporal variations of the δ-factors in these results, we may presume the δ-factors are time-invariant. Therefore we can expect that the accuracies of the δ-factors are expressed by the standard errors of the present analyses.

The obtained phase differences are plotted with the standard errors bars in Fig. 10. We observe that deviations of the phase differences from the mean value amount to $\pm 0.2°$ for M_2, $\pm 0.3°$ for S_2, $\pm 0.1°$ for K_1 and $\pm 0.8°$ for O_1 constituents. These are generally a few times larger than the standard errors obtained from the harmonic analysis. The observed deviations must result from both random and systematic errors. The former errors stem from random noises which have no correla-

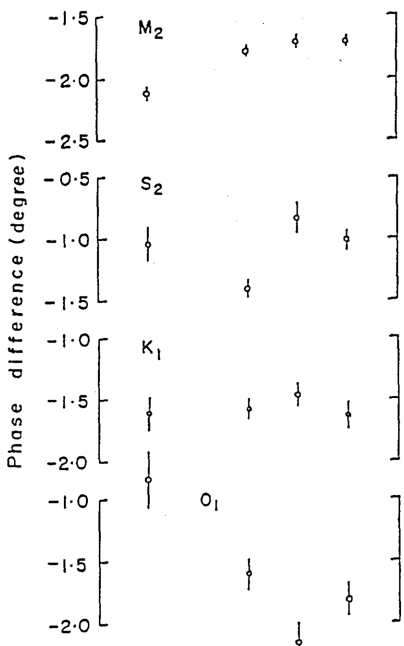


Fig. 10. Phase differences corrected for ocean tides at the Tokyo station in different observaion periods. Error bars estimated from the harmonic analyses are also shown.

tion with the tides. On the other hand, the latter errors stem from tide-correlated noises. Harmonic analyses estimate errors by assuming noises to be random and include tide-correlated noises in the estimation of tidal gravity parameters. Therefore the observed deviations of the phase differences suggest the existence of some systematic errors. There are several possibilities for the systematic errors. SATO (1977) demonstrated a dependency of instrumental phase lags on mechanical sensitivities. Temporal change of the inclination of a gravimeter might cause these deviations of phase differences since the mechanical sensitivity changes with the instrumental inclination. In addition, atmospheric disturbances could cause tide-correlated noises.

7. Regional Variations of Tidal Gravity Parameters in Central Japan

There exist station-to-station differences in the δ-factors which amount to 3% and 5% for M_2 and O_1 constituents, respectively, as listed in Tables 4a, 4b and 4c. These percentages exceed the observational errors and are regarded as statistically significant. The validity of the observational errors of the δ-factors estimated in the present

Table 7a. δ-factors (δ) and phase differences (κ) by the gravimeter G34 after ocean tide corrections.

Station	M_2		S_2		K_1		O_1	
	δ	κ (degree)	δ	κ (degree)	δ	κ (degree)	δ	κ (degree)
Tokyo	1.212	−1.54	1.224	−1.14	1.213	−0.95	1.224	−1.04
Fujigawa	1.207	−1.75	1.240	−1.67	1.208	−3.29	1.229	−1.89
Yahiko	1.219	−1.71	1.216	−1.59	1.210	−1.81	1.230	−1.28
Yatsugatake	1.194	−1.93	1.185	−0.88	1.193	−0.36	1.190	−0.66
Nagano	1.192	−2.08	1.189	−1.79	1.187	−1.75	1.184	−3.01
Kiryu	1.197	−1.66	1.211	−0.46	1.215	−2.44	1.214	−1.69

Table 7b. δ-factors (δ) and phase differences (κ) by the gravimeter G447 after ocean tide corrections.

Station	M_2		S_2		K_1		O_1	
	δ	κ (degree)	δ	κ (degree)	δ	κ (degree)	δ	κ (degree)
Tokyo	1.170	−1.92	1.178	−1.02	1.169	−1.59	1.188	−1.32
Kakioka	1.166	−1.95	1.185	−0.72	1.171	−1.33	1.170	−0.91
Inuyama	1.156	−1.95	1.164	−1.59	1.151	−1.44	1.177	−1.87
Aburatsubo	1.173	−0.71	1.174	+0.02	1.150	−0.50	1.190	−0.56

Table 7c. δ-factors (δ) and phase differences (κ) by the modified gravimeter G447 after ocean tide corrections.

Station	M_2		S_2		K_1		O_1	
	δ	κ (degree)	δ	κ (degree)	δ	κ (degree)	δ	κ (degree)
Tokyo	1.156	−0.24	1.164	−0.32	1.156	−0.46	1.163	0.02
Takayama	1.142	−0.61	1.147	−0.80	1.123	0.09	1.142	−0.36

analysis has already been shown in the preceding chapter. The differences in the δ-factors are considerably reduced by the ocean tide corrections, but there still remain distinct differences after these corrections as listed in Tables 7a, 7b and 7c.

Fig. 11 shows the relative differences in the δ-factors in percent for the four major constituents, respectively. We adopt the value of the δ-factor at the Tokyo station as the reference and subtract it from the values obtained at the other stations. This procedure is necessary for eliminating the instrumental differences as discussed in the preceding chapter. We find that the δ-factor variations are rather similar for the four major constituents. The δ-factor differences obtained at the five coastal stations (YHK, KAK, TKY, ABR and FUJ) are smaller than 0.6% in magnitude for M_2 and 1.5% for O_1 constituent, but those obtained at the stations in the north-western part of Central Japan (NAG, YAT and TAK) show distinct negative values. That is to say, the δ-factors obtained at those stations are about 1.2–1.6% smaller than those at the five coastal stations for M_2 and 1.8–3.2% smaller for O_1 constituent. The δ-factors obtained at the Kiryu (KRY) and the Inuyama stations (INU) are about 1% smaller for both M_2 and O_1 constituents. We find that the δ-factor variations for O_1, S_2 and K_1 constituents are generally larger than those for M_2 constituent. We consider the cause of such differences among constituents is that the accuracies of tidal parameters and ocean tide corrections for those three constituents are inferior to those for M_2 constituent. Here we take the average of the δ-factors for the four major constituents with a weight according to their amplitudes and observe the spatial variations of such averaged values (Fig. 12). We can see a similar tendency as that of the δ-factor itself more clearly. The averaged δ-factors obtained at the three stations (NAG, YAT and TAK) are about 2% smaller than those at the five coastal stations. The values obtained at the other stations (KRY and INU) in the intermediate zone between the two groups of stations are about 1% smaller than those at the five coastal stations. These facts may indicate that a zone of a lower δ-factor exists locally in the north-western part and the δ-factor decreases

Fig. 11 (a)

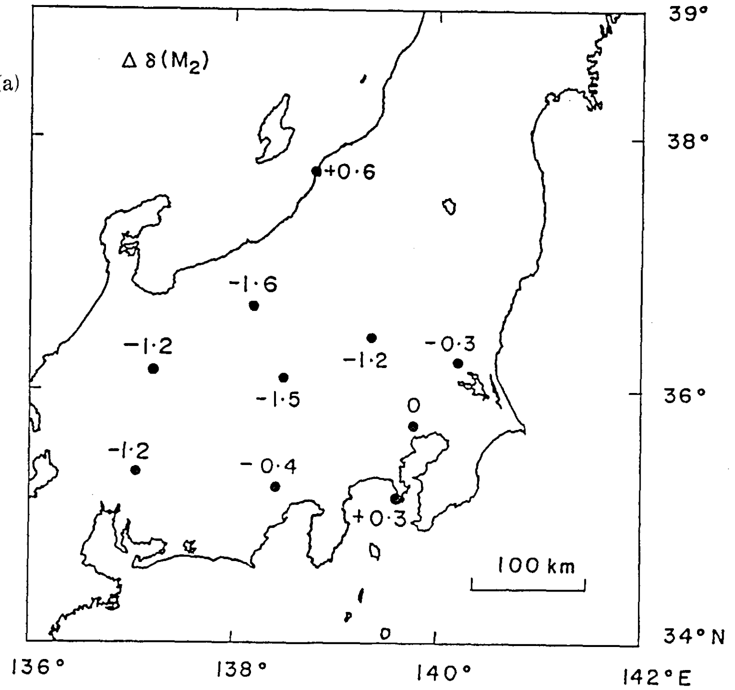


Fig. 11 (b)

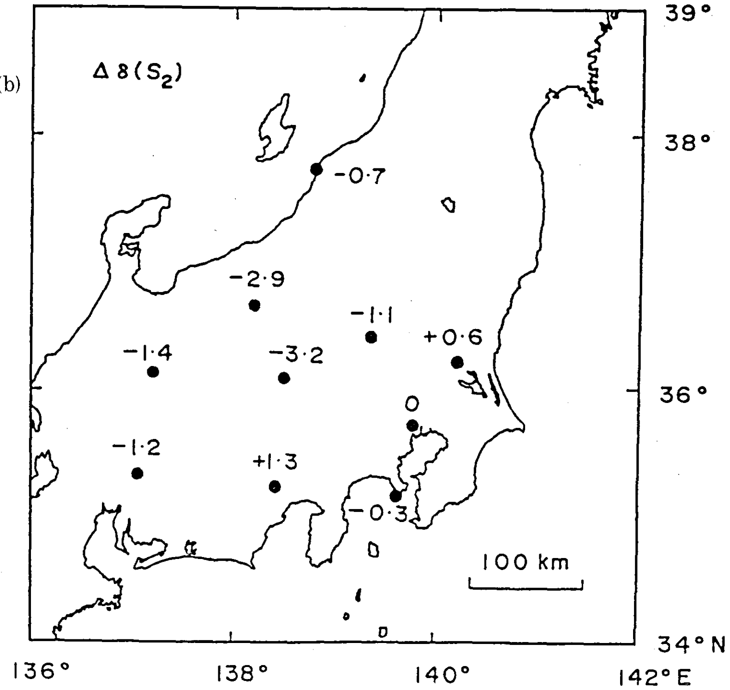


Fig. 11 (c)

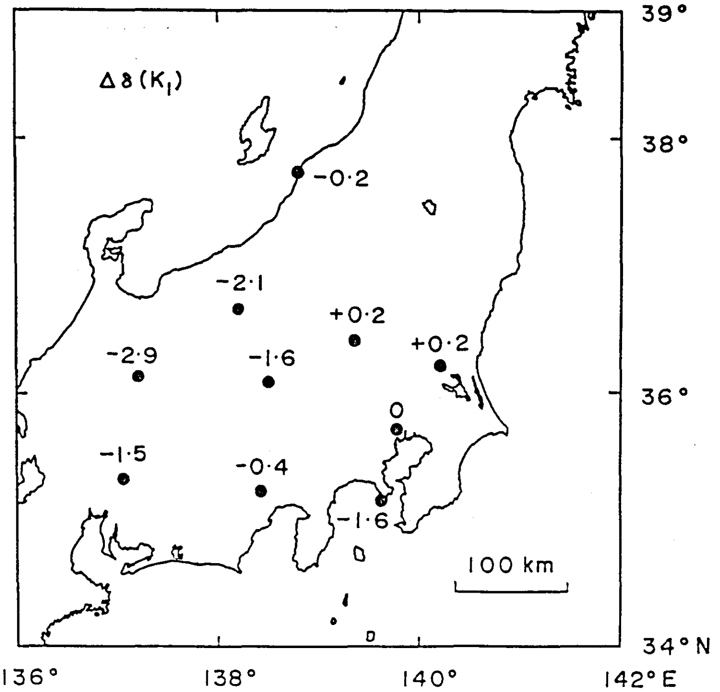


Fig. 11 (d)

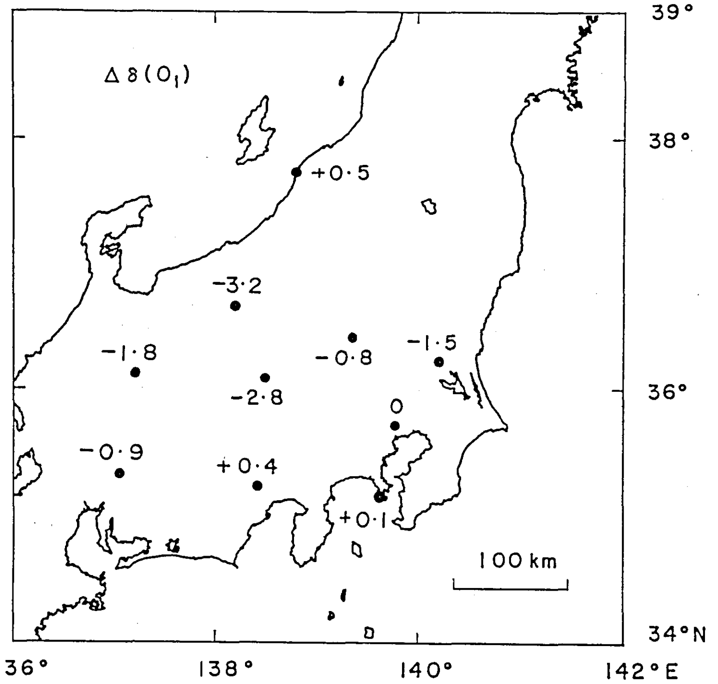


Fig. 11. The relative differences of the δ-factor corrected for ocean tides for the four major constituents from the values at the Tokyo station (in percent).

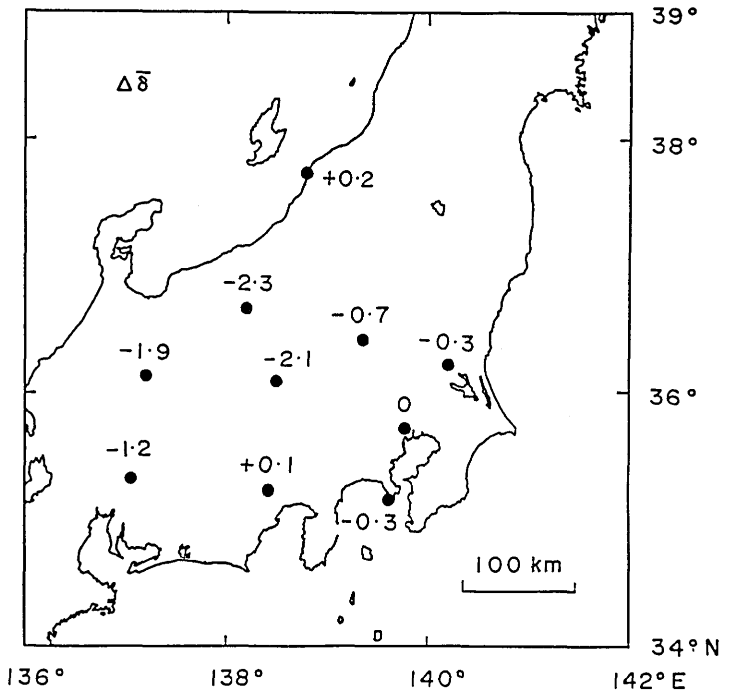


Fig. 12. The relative differences of averaged value obtained from the δ-factors corrected for ocean tides for the four major constituents from the values at the Tokyo station (in percent).

gradually from the coastal zone toward the north-western inland part. The discovery of the existence of the regional variation of the δ-factor in Central Japan is the most important result in the present observations.

In order to investigate a regional variation of phase differences, we obtain statistical variances of the phase difference as follows,

$$\sigma_\kappa=\left(\sum_{i=1}^{n}(\kappa_i-\bar{\kappa})^2/(n-1)\right)^{1/2}, \tag{7-1}$$

where κ_i is the phase difference obtained at i-th observation station; $\bar{\kappa}$ the mean value of the phase differences obtained at all the stations and n the number of stations. In the present case we have nine stations' data. Then $n=9$. We have excluded the data of the Aburatsubo station because the instrumental phase characteristic at the station has been different from that at the other stations. Actually the phase difference at the Aburatsubo station is systematically positive compared with the value at the Tokyo station. $\kappa(\text{ABR})-\kappa(\text{TKY})$ is 1.22° and 0.77° for M_2 and O_1 constituents, respectively. σ_κ for M_2 constituent is reduced

Fig. 13 (a)

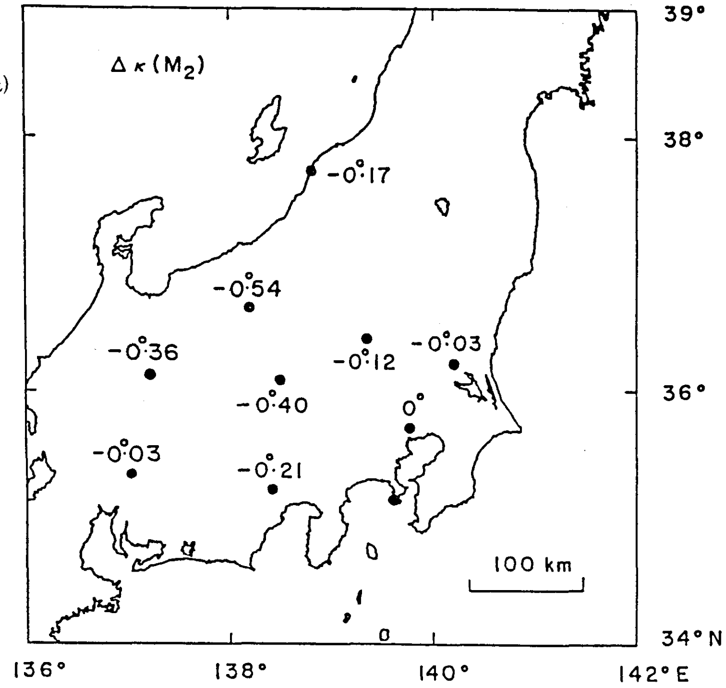


Fig. 13 (b)

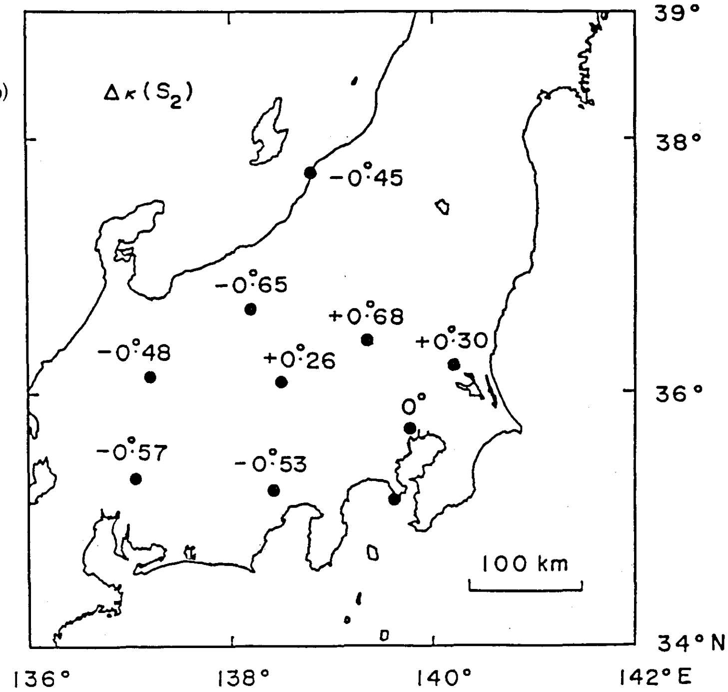


Fig. 13 (c)

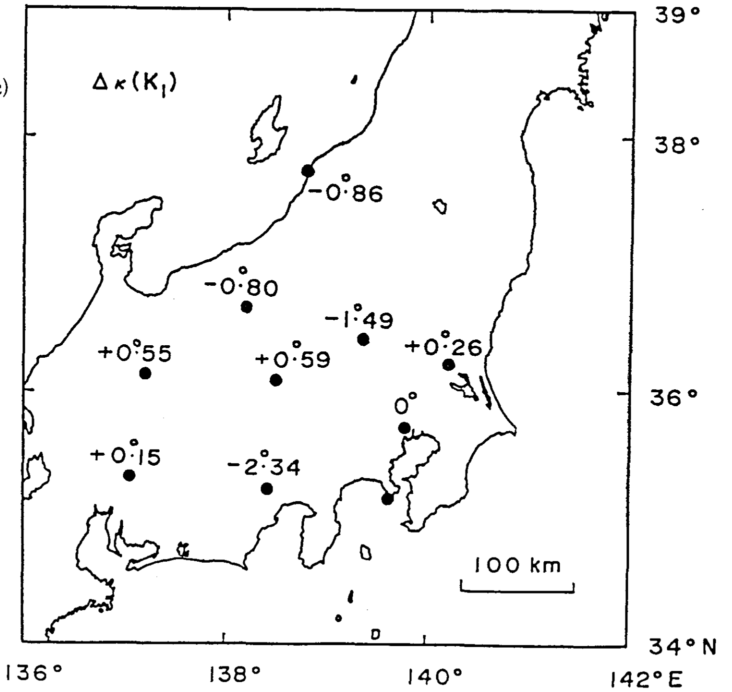


Fig. 13 (d)

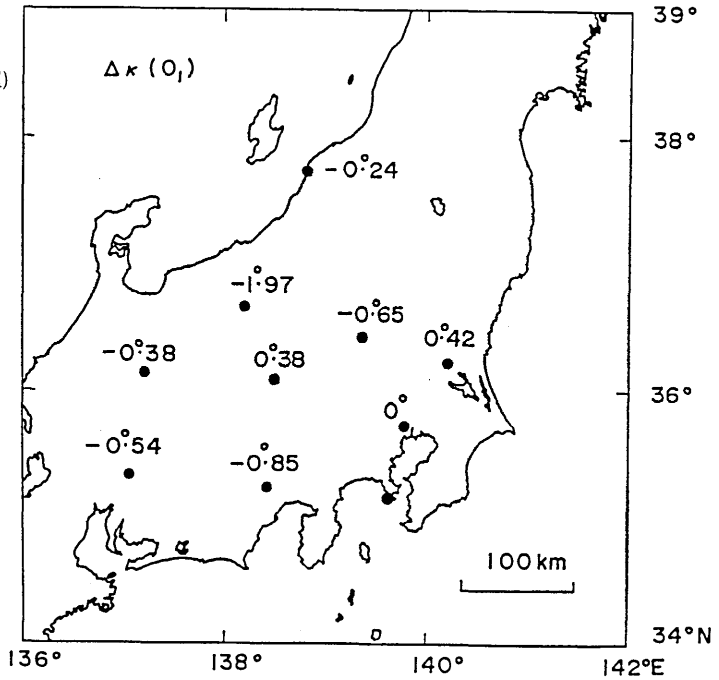


Fig. 13. The phase differences corrected for ocean tides for the four major constituents relative to the values at the Tokyo station (in degrees).

from 0.32° to 0.19° after the ocean tide corrections. This fact suggests the existence of ocean tide effects.

The corrected phase differences for M_2 constituent still take such significant negative values as $-1.5°$ to $-2°$ at all the stations except for the Aburatsubo station as listed in Tables 7a and 7b. On the other hand, the modified gravimeter G447 has shown such a small phase lag as $-0.24°$ after the ocean tide correction at the Tokyo station as listed in Table 7c. We therefore infer that those observed phase differences amounting to $-1.5°$ to $-2°$ result mainly from instrumental phase lags of the gravimeters, G34 and G447. Moreover, we consider those phase differences result partly from errors of the ocean tide corrections.

Fig. 13 shows phase differences with ocean tide corrections for the four major constituents relative to the observed value at the Tokyo station. We cannot observe such a clear tendency as that of the δ-factor in these maps. We have observed in the preceding chapter that the phase difference is more severely affected by noises than the δ-factor and that their fluctuations reach $\pm 0.2°$ even for M_2 constituent at the Tokyo station. However phase differences for M_2 constituent show systematic variations amounting to 0.5° which seem to associate with the δ-factor variations. Fig. 14 suggests such a correlation between the δ-factor and the phase difference variations, that is to say, a lower δ-factor corresponds to a more negative phase difference. SOURIAU (1979) also found a zone of a low δ-factor with a negative phase difference and a high δ-factor with a positive phase difference in Europe. This correspondence may be a clue to resolve a cause of the variations of tidal gravity parameters.

Hereafter we will test the relation between the observed δ-factor variations and other geodetic data. Fig. 15(a) shows the relation between the δ-factor variations for M_2 constituent and the heights of the station above sea level. The heights of the station range from 6.9 m (ABR) to 850 m (YAT). We observe a negative correlation between the two quantities (the linear-cor-

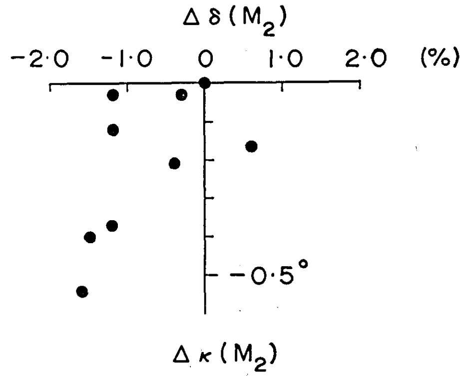


Fig. 14. Relation between the relative differences of the δ-factor corrected for ocean tides from the value at the Tokyo station (in percent) and the phase differences corrected for ocean tides relative to the value at the Tokyo station (in degrees) for M_2 constituent.

relation coefficient is -0.63). Fig. 15(b) shows the relation between the δ-factor variations for M_2 constituent and the gravity values at the Tokyo station. We observe a positive correlation between the two quantities (the linear-correlation coefficient is 0.51). Fig. 15(c) shows

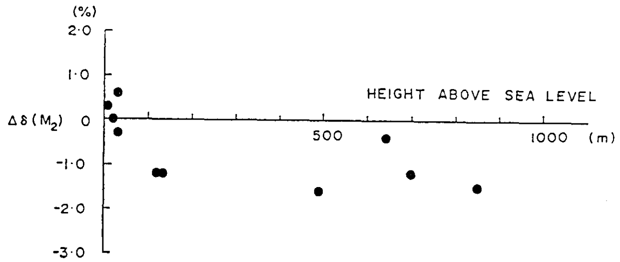


Fig. 15(a). Relation between the relative differences of the δ-factor corrected for ocean tides from the value at the Tokyo station for M_2 constituent (in percent) and the heights above sea level.

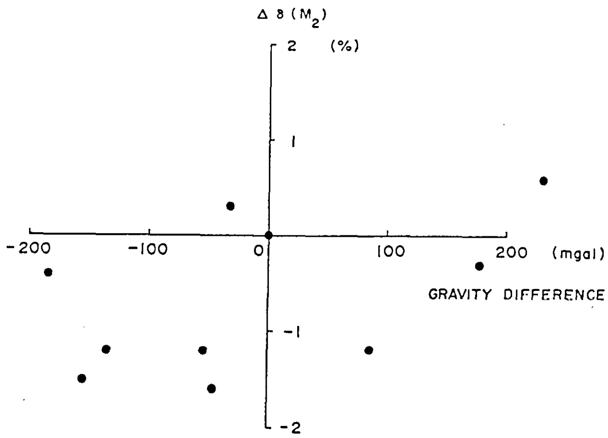


Fig. 15(b). Relation between the relative differences of the δ-factor corrected for ocean tides from the value at the Tokyo station for M_2 constituent (in percent) and the gravity values relative to the value at the Tokyo station.

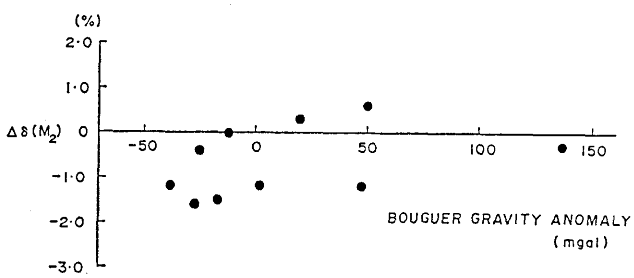


Fig. 15(c). Relation between the relative differences of the δ-factor corrected for ocean tides from the value at the Tokyo station for M_2 constituent (in percent) and the Bouguer gravity anomalies.

the relation between the δ-factor variations for M_2 constituent and the Bouguer gravity anomalies. We observe a positive correlation between the two quantities (the linear-correlation coefficient is 0.39). Since the three quantities (height, gravity value and Bouguer gravity anomaly) are not mutually independent, we find it difficult to draw a conclusion on the cause of the δ-factor variation from the above-described results. We have also tested the relation between the δ-factor variations for M_2 constituent and the long wavelength topographies and the relation between the δ-factor variation of M_2 constituents and the long wavelength Bouguer gravity anomalies. We take the long wavelength topographies and Bouguer gravity anomalies at the observation stations from the map obtained by HAGIWARA (1967). The cut-off wavelength adopted by him is about 150 km which is supposed to be an appropriate wavelength to separate the crustal effects from the spatial broad variations due to the mantle structure. The linear-correlation coefficients are -0.77 and 0.47 for topography and Bouguer gravity anomaly, respectively. The high correlation of δ-factor variations with the long wavelength topographies suggests that the observed variations of the δ-factor may reflect regional underground structures.

We notice an isolated region of a lower δ-factor which is located in the midst of the Central Japan area. The center of this region approximately coincides with the center of the area of the negative Bouguer gravity anomaly covering the north-western part of Central Japan. The coincidence of the center of the δ-factor variations with that of the negative Bouguer gravity anomaly may indicate that such a variation of the δ-factor has some relation to the undulation of the surface of the Mohorovičić-discontinuity located beneath the middle of Central Japan.

From the results of microearthquake observations, FUKAO and YAMAOKA (1983) suggested the existence of an attenuative structure beneath the Hida mountain range in Central Japan. That area almost corresponds to the middle of that with a lower δ-factor. A low velocity structure was also obtained in that area by HIRAHARA (1981). Some relation may exist between such seismological structures and the δ-factor variations. Possible causes, including the regional variation in the crust-mantle structure, will be examined one by one in the following chapter.

8. Discussion

We have found that the δ-factors are about 1-3% smaller in the north-western part of Central Japan than in the south-eastern part after correcting the ocean tide effects on the gravity field. The regional variation of the δ-factor must be attributed either to the incompleteness of ocean tide corrections or to other effects such as a lateral heterogeneity of the earth's structure, atmospheric disturbances and so on. We will discuss possible causes of the δ-factors variation in the following sections.

8-1 Errors in Ocean Tide Corrections

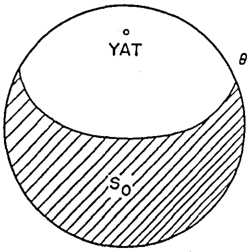
At first sight one could have suspected the incompleteness of ocean tide corrections to be a cause of the δ-factor variation, since it seems to have a correlation with a land-sea distribution. Therefore it is extremely important to check the accuracy of ocean tide corrections. We consider that errors in ocean tide corrections mostly originate from uncertainties in ocean tide data. SCHWIDERSKI (1983) compared his modelled ocean tide with 32 tide data obtained from distant offshore, deep-sea tide gauge stations in the north Atlantic, the northeast Pacific and the southeast Indian Oceans. The differences range from 0-5 cm in amplitudes and 0-6° in phases. He claimed that these results fully validata the 5 cm tide prediction accuracy estimated by him. At present we may consider the accuracy of the Schwiderski model is about 5 cm in open oceans. There exists at any rate no definitive evidence that errors in the tidal data of open oceans are larger than that we have considered. In this section we assume that errors of modelled ocean tide data are about 10 cm. We will estimate the error which results from uncertainties in the assumed ocean tide model.

Ocean tides in the global ocean generate tidal variations in the gravity field. In the present observations, however, the distribution of the gravity stations is limited to a few hundred kilometers, so that the contribution from the global ocean is considered to be almost equal for each of the gravity stations. As recognized in Tables 5a-5b, the obtained spatial variations of ocean tide effects are governed chiefly by the contribution from the sea around Japan. It is interesting to estimate what contribution come from certain parts of the ocean areas. In order to estimate such contributions of the ocean tide effects, we will make calculation as described below.

Supposing that the sea level increases worldwide by 10 cm, we consider its effect on gravity for each station. Although this is an unrealistic and extreme case, it seems useful for an estimation of errors in ocean tide corrections. In order to know the relative contribution of the ocean tide effects, the ocean area is divided into two zones according to distances from the Yatsugatake station (YAT). One is a spherical cap centering on the Yatsugatake station and the other is the remaining part. The reason why we adopt the Yatsugatake station as an origin in the present calculation is that it is located almost in the middle of the observation area. The effects of ocean tides in the remaining part, which is denoted here by S_O, on gravity at an arbitrary station is expressed by a function of ϑ, angular radius of the spherical cap centering on the Yatsugatake station:

$$O(\vartheta)=\int_{S_O}(g_A+g_E)dS, \quad (8\text{-}1)$$

where g_A and g_E are gravity effects due to the attraction of ocean topography and the earth's loading deformations, respectively. For example, we show $O(\vartheta)$ for the Yatsugatake station in Fig. 16. It is seen in this figure that $O(\vartheta)$ decreases smoothly as ϑ increases.



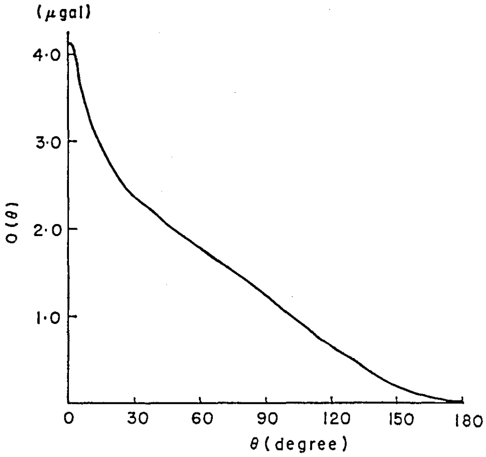


Fig. 16. Ocean tide effects on gravity at the Yatsugatake station as a function of ϑ (eq. (8-1) in the context) assuming that the sea level increases worldwide by 10 cm.

A station-to-station difference in ocean tide effects is important in the present discussion. Then we define a quantity as follows,

$$\Delta_{\text{station}}(\vartheta)=O_{\text{station}}(\vartheta)-O_{\text{YAT}}(\vartheta). \quad (8\text{-}2)$$

This formula represents a difference in ocean tide effects between an arbitrary station and the Yatsugatake station (YAT) arising from a uniform elevation of sea surface by 10 cm. This quantity is expected to decrease rather sharply as ϑ increases. We calculate $\Delta_{\text{station}}(\vartheta)$ for four coastal stations (FUJ, TKY, YHK and KAK). The results are shown in Fig. 17, in which we see that the differences in the ocean tide effects are restricted only to a few degrees of angular distance from the Yatsugatake station (YAT). The differences in ocean tide effects between the Yatsugatake and other stations, except for the Yahiko station (YHK), amount to 0.1 μgal at most, so that their influence

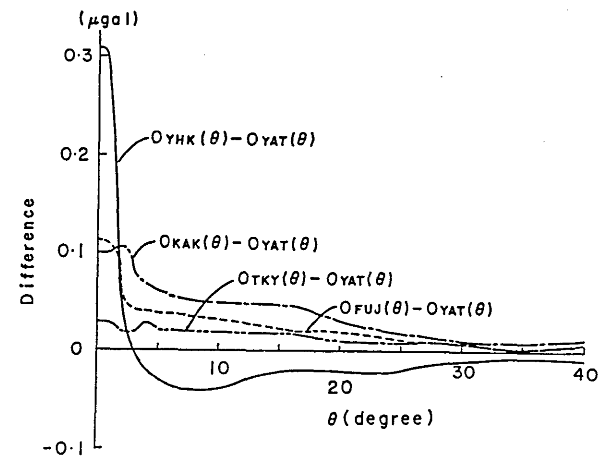


Fig. 17. Differences of ocean tide effects at coastal stations from that at the Yatsugatake station as a function of ϑ (eq. (8-2) in the context) assuming that sea level increases worldwide by 10 cm.

on the δ-factor may become only 0.2–0.3%.

Although the present estimation is made by assuming an extreme case where the sea surface elevates 10 cm over the whole ocean area, the quantity of $\Delta_{\mathrm{station}}(\vartheta)$ is proved to be negligibly small. In an actual case, the correction errors of station-to-station differences in ocean tide effects on the gravity stations may be much smaller than in this extreme case. We therefore conclude that the regional variation of δ-factors cannot be explained by errors in the ocean tide corrections. This conclusion suggests the possibility of other causes, such as geological structure, atmospheric disturbances and so on.

8–2 Atmospheric Disturbances

Atmospheric pressure variations have tidal components. In regarding atmospheric pressure effects on gravity, many experimental studies have been carried out (NAKAI, 1975; WARBURTON and GOODKIND, 1977; SHIMADA, 1979). These studies have obtained the admittance ($\Delta g/\Delta p$) of a gravity change (Δg) per a pressure variation (Δp) as 0.3–0.4 μgal mb^{-1}. The observed admittance is almost equal to what we expect theoretically from the gravitational attraction of an air-mass change and the air-mass loading corresponding to a pressure change.

Fig. 18 shows the amplitude spectra of the atmospheric pressure variations observed at the Fujigawa station (FUJ). The duration of the analyzed data is from August 23, 1981 to November 11, 1981. Dominant peaks exist in diurnal, semi-diurnal and ter-diurnal bands. We estimate the atmospheric pressure effects on gravity at the Fujigawa

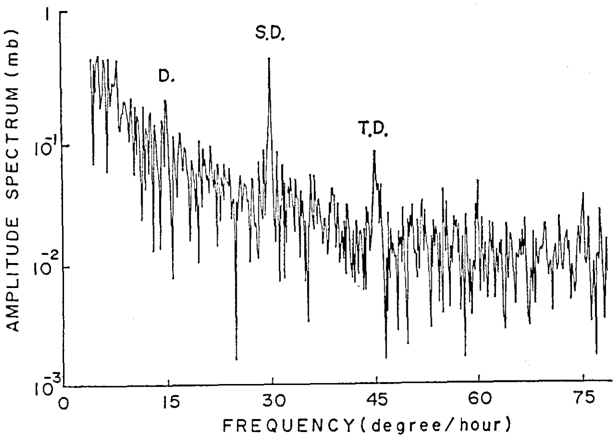


Fig. 18. Amplitude spectra of atmospheric pressure variations at the Fujigawa station. D., S.D. and T.D. denote peaks at the diurnal, semi-diurnal and ter-diurnal frequency bands, respectively.

Table 8. Atmospheric pressure variations (Δp) of tidal frequencies at the Fujigawa station, their effects on gravity (Δg) and those on the δ-factor ($\Delta\delta$).

Constituent	Δp (mb)	Δg=0.35 (μgal mb^{-1})$\cdot\Delta p$ (μgal)	$\Delta\delta$
M_2	0.09	0.03	0.0006
S_2	0.52	0.18	0.008
K_1	0.23	0.08	0.002
O_1	0.06	0.02	0.0007

station by assuming as $\Delta g/\Delta p$=0.35 μgal mb^{-1}. The results are listed in Table 8. The effects are 0.1-0.2% for M_2, K_1 and O_1 constituents but considerable larger (about 1%) for S_2 one. Except for S_2 constituent, the effect of atmospheric pressure distubances on gravity tides may be less than the observational errors.

Since the present tidal observations have been made at various heights (6.9-850 m), we should also consider a direct influence of air pressure on a gravimeter. For example, the LaCoste & Romberg gravity sensor is affected by buoyant force of air. Although the sensor is designed to be compensated for buoyant force, it is probably that its compensation is slightly out of balance. This uncompensated buoyant force may cause a systematic error which depends on atmospheric pressure values. However, any pressure-dependent error should also affect field gravity surveys in mountain areas. TAJIMA *et al.* (1978) showed the instrumental differences of two gravimeters' results are within ±0.02 mgl at observation sites with height differences amounting to 600 m (gravity differences are about 200 mgal). NAKAGAWA *et al.*

(1983) reexamined this problem in the Andes mountain areas and showed the instrumental differences of eight gravimeters' results are within ± 0.1 mgal in observation sites with height differences amounting to 3000 m (gravity differences are about 1000 mgal). From these studies, we estimate that a pressure-dependent error does not exceed a few parts in 10^{-4} in relative accuracy. We may say that it is less than the errors of the present observations (0.5-1.0%). Therefore we may not consider pressure-dependent errors of instrumental origin in interpretations of the observational results.

8-3 Effects of Groundwater

As is well known, water-levels in wells are much affected by earth tides (MELCHIOR, 1978). Tidal fluctuations in groundwater levels might affect gravity tides. In a first-order approximation, the water-level effects on gravity can be expressed as $2\pi G\rho_W h\phi$, where h is the change of groundwaterlevel, G the Newtonian gravitational constant, ρ_W the density of groundwater and ϕ the porosity of rocks.

If we choose $\rho_W=1.0\ \mathrm{g\,cm^{-3}}$, $h=10^{-2}$ m and $\phi=0.2$, the waterlevel effect becomes 0.08 μgal. This is about 0.2% of the total amplitude for M_2 constituent and less than the observational errors. Even in such an unrealistic case, the expected effect of change in groundwater level is too small to be observed, so that this cannot be responsible for the observed δ-factor variations.

8-4 Effects of Topographies

In chapter 7 we have shown the correlation between the δ-factor variations and the heights of the stations. Therefore we should consider topographic effects on gravity tides. Gravity has close relation to vertical displacement whose horizontal derivative approximately corresponds to tilt. HARRISON (1976) calculated topographic effects on tilt and strain by finite element techniques. According to his results, 1 in 10 (5.7°) slope which is rather steep for long wavelength topography produces strain-induced tilts amounting to 0.1 to 0.4 times the regional strain in its vicinity. Since tidal strain amplitude is about 10^{-8}, tidal tilt perturbation becomes $(1\text{-}4)\times 10^{-9}$ radian. Even if such tilt perturbation occurs in the horizontal dimension of 100 km, perturbation of vertical displacement does not exceed 0.4 mm which corresponds 0.3% in terms of the δ-factors. Therefore we consider that topographic effects are not sufficient to explain the observed δ-factor variations.

8-5 Effects of Lateral Heterogeneities of the Earth's Structure

It is an obvious fact that the earth's structure is not laterally uniform. The tidal phenomena observed on the earth's surface may depend on its ununiformity. Hereafter we first discuss an effect of the earth's lateral heterogeneous structures on load tides and then on body tides.

BEAUMONT (1978) calculated a load deformation due to annular modelled loads by taking structural changes at the continental margin into consideration by finite element techniques. He intended that the results should be illustrative of a load deformation of laterally heterogeneous earth. He concluded that on either side of the continental margin the vertical displacement would be adequately modelled by laterally homogeneous models, and that directly over the continental margin the displacement perturbation is at most 10% and trends smoothly from the calculated value by adopting the oceanic structure to that by the continental structure. Based on his results, the maximum displacement perturbation of 10% is about 0.6 mm for ocean tide amplitude of 30 cm which is approximately an amplitude of M_2 ocean tides around Japan. Therefore the perturbation of gravity tides reach only 0.4% of the total amplitude at maximum. We may say that effects of lateral heterogeneities on load tides are below the observational errors of the present observations. However, those effects are important for tilt tides as shown by Beaumont.

The effects of lateral heterogeneities on body tides have not been fully clarified. OKUBO and SAITO (1983) showed that tidal Love numbers are very insensitive to the upper mantle and crustal structures for a spherical symmetric earth model. It has generally been thought that lateral variations in tidal Love numbers are too small to be detected with the present accuracy of observations. BEAUMONT and BERGER (1974) calculated the deformation due to body tides for a laterally heterogeneous structure by finite element techniques. They assumed seismic velocity anomalies to be 10–30% in a dilatant zone of forty kilometers across. Their calculated results show that the perturbation for vertical tidal displacements does not reach 1% of the total tidal amplitude in contrast to the perturbation for both tilt and strain tides of 50–100%. They consider that the tidal displacement perturbation is too small to be detected with a gravimeter. ZÜRN *et al.* (1976) estimated the influence on vertical displacements of the plate-subducting structure as 0.8% at maximum. However, their estimations were made for some special cases. A general treatment using the spherical harmonic expansions was attempted by MOLODENSKII and KRAMER (1980). We use their formulas to estimate effects of lateral heterogeneities for a simple model shown in Fig. 19. As a preliminary approach, we adopt a homogeneous incompressible earth model with a lateral heterogeneity.

The homogeneous part of that model has a density $\rho_0=5.52\,\mathrm{g\,cm^{-3}}$ and a rigidity $\mu_0=3.45\times10^{11}\,\mathrm{dyn\,cm^{-2}}$. The heterogeneous part is cylindrically shaped and extends vertically from the surface to an assumed depth with the same contrast of rigidity, $\Delta\mu$. The heterogeneity is located

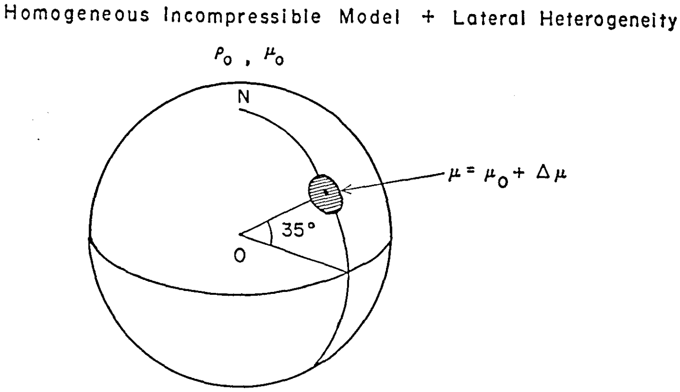


Fig. 19. Lateral heterogeneous earth model constructed from the homogeneous incompressible earth model (the density $\rho_0=5.52\,\mathrm{g\,cm^{-3}}$, the rigidity $\mu_0=3.45\times10^{11}\,\mathrm{dyn\,cm^{-3}}$, the radius $a=6.371\times10^{3}\,\mathrm{cm}$) and a locally distributed heterogeneity (hatched area) which is cylindrically shaped and which extends vertically from the surface to an assumed depth with the same contrast of rigidity, $\Delta\mu$. The heterogeneity is located at the latitude $\varphi=35°$ and the longitude $\lambda=0°$.

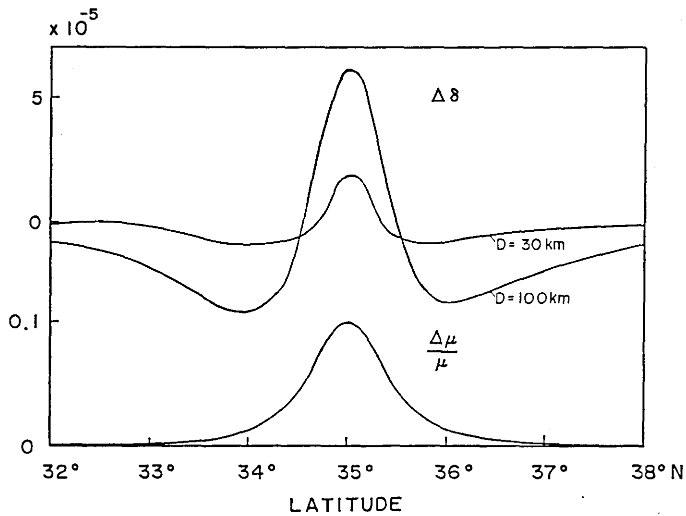


Fig. 20. Results of calculations for the model shown in Fig. 19. The upper figure indicates δ-factor variations along the meridian of the longitude $\lambda=0°$ for the two cases. One case is that the heterogeneity extends vertically to a depth of 30 km and the other to a depth of 100 km. The lower figure indicates the assumed rigidity perturbation along the same meridian.

at the latitude $\varphi=35°$. We give a laterally smoothed variation of rigidity which simulates a cylindrical shape for calculations by spherical harmonic series. Their horizontal extents are about 2° which approximately corresponds to that of the observation area. We calculate for two cases, one in which the heterogeneity extends to a depth of 30 km and the other to a depth of 100 km. The result is shown in Fig. 20. We see that the calculated effects on the δ-factor reach only 10^{-4} and are by two orders of magnitude smaller than the observed variations of the δ-factor. Therefore the adopted model cannot explain the observed δ-factor variations. However, that model is very simplistic, so that we should not conclude that effects of lateral heterogeneities are not important from these results. We should make calculations for a more complex model. Block structures or anelastic properties of the earth seem necessary for the model. A local response of the earth must be significantly different from that of the spherical symmetric model in order to cause large δ-factor variations.

We have discussed possible causes of the regional variation of tidal gravity parameters observed in Central Japan. Although there exist some ambiguities about ocean tide models in open oceans at present, there are not enough data by which we can check their accuracies. However, judging from limited data, errors of the modern ocean tide model seem within ± 5 cm (SCHWIDERSKI, 1983). If modelled ocean tide data in the sea around Japan have also the same accuracy we cannot attribute the cause of the δ-factor variations to ocean tide effects as discussed in the first section of this chapter. As discussed in chapter 7, the observed δ-factor variation seem to have relation to large scale structural changes. We should further investigate effects of underground structures. Since that problem is a complicated one, we cannot fully clarify it in this study.

9. Conclusions

The regional variation of tidal gravity parameters has been found by tidal gravity observations at ten stations in Central Japan. The δ-factors obtained at the Yatsugatake and Nagano stations, are about 1.5% smaller for M_2 constituent and 3% smaller for O_1 constituent than those at the five coastal stations: Yahiko, Kakioka, Tokyo, Aburatsubo and Fujigawa. The δ-factors obtained at the other stations, Kiryu, Inuyama and Takayama are about 1% smaller for both M_2 and O_1 constituents than those at the five coastal stations. The regional variations of the averaged δ-factor obtained from values for the four major constituents show a rather clearer tendency than those of the

δ-factor itself. The averaged δ-factors obtained at the three stations, Takayama, Nagano and Yatsugatake are about 2% smaller than those at the five coastal stations. The averaged δ-factors obtained at the other stations in the middle zone between the two groups of stations are about 1% smaller than those at the five coastal stations. These facts may confirm that the zone with a lower δ-factor is isolated locally in the north-western part of Central Japan and the δ-factor decreases gradually from the south-western part of this area to the north-western part. In addition, the phase differences for M_2 constituent also show systematic variations associated with the δ-factor variations. That is, as the δ-factor decreases, the phase lag increases.

Similar observational results on regional variation of tidal gravity parameters have been previously obtained by PARIYSKIY (1978) and MELCHIOR and BECKER (1983). The variation observed in the present study is comparable in magnitude with those observed in their studies, but its horizontal scale is much smaller than theirs. The present result may prove the possibility of a small-scale δ-factor variation of a few hundred kilometers in diameter.

We have examined five cases in section 8 as possible causes, which are as follows:

(1) Incompleteness of the ocean tide corrections
(2) Atmospheric disturbances
(3) Groundwater
(4) Topographies
(5) Lateral heterogeneities of the earth's structure

We have estimated the influences of (1), (2), (3) and (4) to be less than observational errors and considered that those of (5) need further investigation.

At first sight one could have suspected the incompleteness of ocean tide corrections to be a cause of the δ-factor variation, since it has a correlation with a land-sea distribution. However, we have estimated the correction error of ocean tide effects to be at least one order of magnitude smaller than the δ-factor variations.

We should further investigate the effects of the lateral heterogeneity of the earth which has both elastic and anelastic properties. Those effect cannot be fully clarified in this study. The reason why we consider those effects important is that the variations of tidal gravity parameters seem to have relation to the regional underground structures as discussed in chapter 7 and moreover, the area with a lower δ-factor approximately corresponds to the center of the negative Bouguer gravity anomaly and several seismological studies suggested that an attenuative and low velocity structure exists beneath that area. The existing tidal theory of a laterally heterogeneous earth can be applied only to very limited cases, so that we cannot apply it directly to the present case. Its improvement is necessary for investigating the physical mechanism which causes an anomalous tidal deformation.

In addition to tidal gravity observations, the regional variation of earth-tidal parameters can be detected by observations of tilt and strain tides. Although tilt and strain are much more affected by local conditions and noises than gravity, the observed tidal parameters of tilt and strain tides depend much more on the lateral heterogeneity of the crust and upper mantle structures than those of gravity tides. Therefore these observations may provide a clue to resolve the cause of the δ-factor variation obtained through the present investigations in the future.

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Appendix Implementation of electro-static feedback with a LaCoste & Romberg gravimeter

The LaCoste & Romberg gravimeter was originally designed for measurements by the null method. However, in tidal gravity observations, the instrument has been widely used by the "deflection method", that is, recording the output of the capacitive position indicator (CPI) with which the instrument is equipped. HARRISON and SATO (1984) pointed out the disadvantages of this deflection method and devised a simple method for implementing electro-static feedback with a LaCoste & Romberg gravimeter. Clearly there are important advantages to observing earth tides with feedback gravimeters in which the beam is maintained at its null position. Hysteresis effects are then eliminated as the length of the spring remains constant. That is, the instrumental phase lag becomes much smaller (almost zero) than that by the deflection method. In addition, over reasonable tilts, the sensitivity of the meter is independent of tilt (although the observed value of gravity will, of course, change as the cosine of the tilt angle). We employed this method in observations at the Takayama and Tokyo station in 1984. We will describe the outline of their method in the following text.

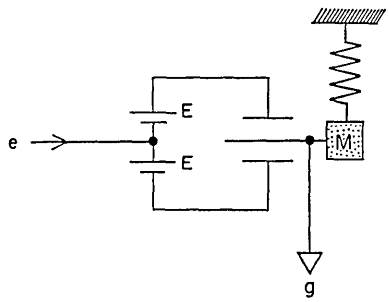

Fig. A-1. Electrostatic force transducer.

In electrostatic feedback techniques the output of the beam position indicator is amplified and applied, together with a differential bias voltage, to the CPI plates in such a manner as to develop an electrostatic force returning the beam to its null position as shown in Fig. A-1. In Fig. A-1 the fixed plates are biased $\pm E$ volts, and a feedback voltage e is applied; the fixed plates are spaced at a distance $2d$ apart and the moving plate is placed at distance δ from the center position.

Harrison and Sato further modified the situation in Fig. A-1 as the feedback voltages applied to the two fixed plates are not equal in order to attain the feedback force proportional to the feedback voltage. We will suppose them to be $(1+k)e$ and $(1-k)e$ respectively where k is a small number. The feedback force in this case given by

$$\frac{2C_0}{d}\left[\left\{1+\frac{2\delta k}{d}\right\}eE+\left\{\frac{\delta}{d}(1+k^2)+k\right\}e^2+\frac{\delta}{d}E^2\right]$$

to terms in first order in δ/d, where C_0 is the capacitance between

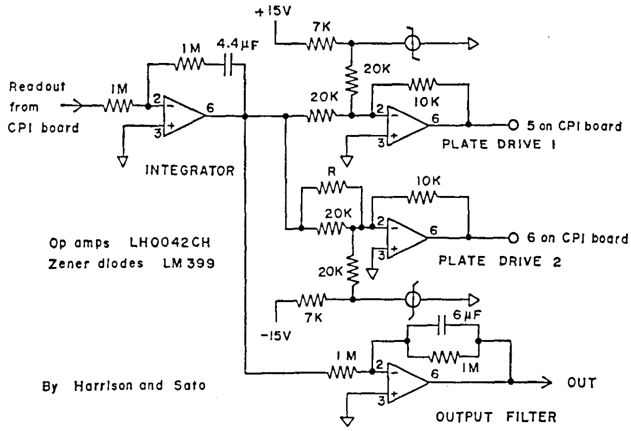

Fig. A-2. Circuit used for electrostatic feedback of LaCoste & Romberg gravimeter.

either fixed plate and the center plate when the latter is centrally located. We can adjust k so that the coefficient of the term in e^2, $\{(\delta/d)(1+k^2)+k\}$ is 0.

The unequal feedback is provided with a circuit of the type shown in Fig. A-2. The error signal from the beam is integrated and the output of the integrator summed with bias voltages derived from Zener diodes at the summing junction of the two separate plate drive operational amplifiers 1 and 2. The feedback is linearized by the addition of an appropriate resistor R which has the effect of adjusting the fraction of the integrated beam position voltage summed into the output of the plate drive operational amplifier. R may, of course, have to be added in the corresponding position of plate drive 1, depending on the sign of δ.

中部日本における重力潮汐の地域的変化に関する観測的研究

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地球潮汐は月，太陽の引力に伴い，地球が変形して生ずる現象であり，地球上では重力，傾斜，歪等の潮汐変化として観測される．それらの観測量の潮汐変化の振幅，位相を記述する潮汐パラメータは地球内部構造に依存する量であり地球物理学的に重要である．過去 20 年ほど，特に重力潮汐パラメータの地域分布を観測で決定する研究が世界中で精力的に行なわれてきた．これらの研究を通じて海洋潮汐が重力潮汐に顕著な影響を与えており，その地域変化の原因の1つであることは明らかになった．そして海洋潮汐の影響をとり除いた潮汐パラメータは観測誤差の範囲内で地球上で一様であると考えられるようになった．しかし過去においては海洋潮汐の影響を正確に除去することが困難であり，詳しい議論をすることは難しかった．最近は詳細な海洋潮汐モデルが作られ，海洋潮汐の影響の補正の精度が向上した．その結果，生じてきた問題は潮汐パラメータの地域変化はすべて海洋潮汐の影響で説明できるのかどうかという事である．潮汐パラメータの地域変化は地球の水平方向の不均質構造を反映しているのではないかという考えも存在する．近年，MELCHIOR and BECKER により海洋潮汐の影響では説明できない重力潮汐パラメータの地域変化がヨーロッパのアルプス付近に存在することが指摘されている．彼らはその変化を地下の地震学的構造と関連づけている．ただし地域変化の議論を進めていくためには，まだまだ，データが十分ではないと考えられる．

本研究は特に地学的に興味のある中部日本において，海洋潮汐の影響以外の有意な潮汐パラメータの地域変化が存在するのかどうかを明らかにするためにラコスト重力計2台による潮汐観測を行なったものである．観測点は中部日本をおおうように分布する 10 地点である．データの解析の結果，重力潮汐パラメータの顕著な地域変化が明らかになった．その変化は海洋潮汐の影響を補正しても依然として残る．海洋潮汐の影響の補正後，求められた δ-factor は M_2 潮については 1.5% 程，O_1 潮については 3% 程，この地域の南東部から北西部に向かって減少する傾向を示す．数 100 km の範囲におけるそのような大きな δ-factor の変化は過去の研究では，はっきりとは，とらえられていない．この点は本研究の最も重要な結果である．

次にこれらの観測された δ-factor の変化の要因について考察した．それらは，(1) 海洋潮汐の影響の補正誤差，(2) 気象擾乱，(3) 地下水，(4) 地形，(5) 地球の水平方向の不均質性である．(1) は求められた δ-factor の変化が計算による海洋潮汐の影響量の変化と類似した傾向をもつ点から，まず考えられるべき要因である．しかし単純な考察の結果，あり得る海洋潮汐データの誤差を考慮しても本研究で観測された程の 1% に及ぶ δ-factor の地域変化は説明できないことがわかった．また (2), (3) および (4) の影響はその変化より1オーダー程小さいこともわかった．そこで現段階では，地球の水平方向の不均質構造が可能性の高い要因と考えられる．その理由としては観測された δ-factor の地域変化が地下構造の変化と関連をもつようにみえることおよび，相対的に小さな δ-factor の観測されている地域では地下における地震波速度が小さく，地震波の減衰が大きいと推定されていることが挙げられる．ただし本研究では定量的に満足できる説明を与えることはできない．この仮説を検討するためには，この観測地域における重力だけではなく，傾斜および歪潮汐の観測，そして水平方向の不均質構造を考慮した地球潮汐理論の研究が必要であろう．