Earth Planets Space, **52**, 579–586, 2000

Amplification of gravity and Rayleigh waves in a layered water-soil model

Tatyana Novikova[1], Kuo-Liang Wen[2], and Bor-Shouh Huang[1]

[1]*Institute of Earth Sciences, Academia Sinica, Taiwan*
[2]*Institute of Geophysics, NCU, Taiwan*

(Received August 23, 1999; Revised August 18, 2000; Accepted August 25, 2000)

The coupled seismic to gravitational surface wave fields are analyzed in a liquid layer lying on the gravitating elastic, low-rigidity half-space. Solution is obtained within the framework of the normal mode formalism applied to the flat ocean-solid Earth model. From the theory of propagation of coupled surface waves (Rayleigh and Love) in layered media, we find the individual multipliers that determine the surface wave spectrum over the entire frequency range. Spectra of excitation functions are investigated for dip-slip point source in the half-space. Main results can be summarized as follows. When the half-space is filled with sediments, dip-slip excitation functions of gravity and Rayleigh waves are one order of magnitude larger than for the half-space composed of hard rocks. Including gravity in the elastic medium essentially changes the character of gravity wave spectrum, leading to an appearance of the third maximum. At the deepening of the source amplitude of this maximum increases. Theoretical marigrams show that including gravity in the half-space also increases period of the gravity wave excited by deep sources by a factor of two, up to 10 minutes. At the same time, presence of gravity force in the half-space has no effect on the spectrum of the Rayleigh wave.

1. Introduction

Considerable effort has been taken to develop both the analytical description of the propagation process of long period gravity (tsunami) and ocean Rayleigh waves for the case of uniform depth (Yamashita and Sato, 1974; Ward, 1980; Comer, 1984a, b; Okal, 1988; *et al.*) and the numerical one for actual bathymetry (e.g. Hwang, 1972; Satake, 1985).

Yamashita and Sato (1974, 1976), using the fully coupled ocean-solid Earth model, analyzed the influence of a number of focal parameters-dip-angle, fault length and focal depth and the rise time of the source time function- on the tsunami (long period gravity wave) and Rayleigh waves. Earthquake source was modeled by the point dislocation and a finite-moving source. Numerical computations have been performed for the dip-slip source model, because large submarine earthquakes near Japan are of this type. Their results can be summarized as follows. In the case of tsunami wave, such parameters as dip-angle, fault length and depth all play an important role in generation process. The maximum amplitudes in the wave trains at azimuths $\varphi = 90^\circ$ and $\varphi = 270^\circ$ become large for large dip-angles. In the case of $\varphi = 0^\circ$, the maximum amplitude is the largest when dip-angle is 45°. The fault length also significantly affects the directivity. When the ratio- of 'fault width to its length increases, the directivity at the azimuth $\varphi = 90^\circ$ and $\varphi = 270^\circ$ become very strong. Velocity of rupture propagation and rise time do not affect much the tsunami amplitude and waveform. In the case of Rayleigh wave, the spectral amplitude decreases monotonously, with as the increase of focal depth. The dip-angle factor has the entirely opposite effect compared to the tsunami wave. The spectral amplitude of Rayleigh wave predominates for smaller dip-angle. The rise time parameter also affects strongly the Rayleigh wave, which become essentially suppressed-as rise time increases.

Ward (1980) considered the tsunami generation problem in the context of a spherically symmetric self-gravitating and elastic ocean-Earth model using the normal mode formalism. He formulated and solved full linearized equations of motion in a manner similar to that originally applied to the Earth's free oscillations. Using moment tensor representation for both point and line seismic source he has derived expressions for tsunami wave displacement in near- and far-field zones.

Consequently Comer (1984a, b) introduced a solution for the tsunami mode excitation in the flat Earth by a point source. He emphasized that the excitation problem in the flat Earth differs substantially from the corresponding problem for the spherical Earth. There is a continuum of frequencies and wavenumbers for the flat Earth but frequencies and angular orders are discrete on the spherical Earth. Also, the normal modes of a finite body form a complete basis for the small oscillations of the body but those of an infinite body, like the flat Earth, do not. In common with Ward, Comer assumed the ocean to be non-viscid and only considered linearized equations of motion and boundary conditions. Alike the earlier authors investigations (Pod"yapol'sky, 1968, 1970; Alexeev and Gusiakov, 1976) he has shown that far-field tsunami in the fully coupled ocean-solid Earth model depend strongly on the source depth, duration, moment and mechanism.

Okal's study (1988) was largely based on Ward formalism considering tsunami wave as the superposition of the free oscillations of an elastic self-gravitating Earth, excited by a seismic sources. Extending the range of source depths to 250 km he showed that this parameter plays only a limited role in controlling the tsunami amplitude. More important are the effects of directivity due to rupture propagation along the fault and the possibility of enhanced tsunami excitation in an elastically weak material. Analyzing tsunami and Rayleigh wave generation by non-double couple sources, he found, in particular, that a landslide involving weak sediments could result in very large tsunami.

In the description of tsunami and Rayleigh waves propagation in most models, discussed above, the gravity force was included only in the liquid layer. Since in the present study we take gravity into account also in the solid Earth, we briefly review the previous works in this direction.

Surface wave propagation in the elastic medium under the influence of gravity was considered by a number of authors (Matuzawa, 1925; Gilbert, 1967; Lomnitz, 1970, 1990, 1991; De and Sengupta, 1976). Gilbert (1967) described a gravitationally perturbed Rayleigh wave in a low-rigidity medium which forms a linear continuum ranging from a solid rocks to water. In this approach everything is attributed to the variation of a single parameter- the rigidity. Gilbert has made energy estimations for surface waves and showed that in a low rigidity medium such as unconsolidated sediments the energy of gravity is comparable with the shear strain energy. In the case of Rayleigh wave propagated in the non-gravitating half-space the $\overline{S}$ is more dominant pulse. In the present of gravity, as β decreases, the $\overline{S}$ pulse becomes insignificant, and the $\overline{P}$ pulse becomes dispersive and approaches the behavior of the classical gravity wave.

Lomnitz (1990), analyzing the behavior of sediments during the Mexico 1985 earthquake, has supposed that a nonlinear mechanism could make the rheology of the Mexico City soft clay nearer to a fluid than to a solid. At large amplitudes, when the stress-strain relation becomes strongly nonlinear sediments may liquefy, and in such conditions the gravity wave propagates as in a liquid. It is worth mentioning here that originally Matuzawa (1925) proposed the idea of existence and propagation of waves of hydrodynamic origin in soils. Based on the study of Tokyo 1923 earthquake, he speculated that sediments could behave like solids at high frequencies and like fluids at low frequencies.

Dealing with gravity wave propagation in the system consisting of the liquid layer and elastic half-space it is possible to neglect the gravity in an elastic medium, since it is small in comparison with the elastic forces. General effect of gravity on the wave propagation in such a system is determined by gravity effect in the liquid. When the ocean bottom consists of rigid rocks the above assumption is reasonable. Although in this case too, including the gravity force in the motion equations of the elastic substratum, changes slightly the dispersion curve at the 10000 seconds period (Pod"yapol'sky, 1968, 1970; Alexeev and Gusiakov, 1974). In the present work, however, we focus on the wave excitation by seismic sources located in low-rigidity media. Therefore it is appropriate not to omit gravity from the description of this medium, since its effects can be evaluated to some extent even within the framework of linear theory.

Present study is focused on the developing an analytical model of gravity and Rayleigh waves excitation by a realistic earthquake source located in the gravitating low-rigidity half-space. We do not deal in detail with the effects of focal parameters on the gravity and Rayleigh wave generation, because it was already done in previous studies mentioned above. It is worth noting that in the case of gravity wave we analyze the wave spectrum in the whole range of frequencies, not limited by its long-period component (tsunami). We will describe the seismic response in function of source location in the half-space and compare the results obtained for a liquid layer and a gravitating half-space model with a liquid layer plus a non-gravitating half-space.

2. Statement of the Problem

Consider gravity and Rayleigh waves in a compressible liquid layer of uniform thickness H over a half-space. Gravity acts in both media. We assume: (1) both media are homogeneous; (2) in the solid medium stress and deformation are related through Hooke's law and in the liquid pressure is proportional to the degree of compression; (3) deformations are small and displacements in the liquid are small compared to the liquid layer thickness and characteristic wavelengths. The latter condition allows us to neglect the distinction between Eulerian and Lagrangian variables in the equations of motion (Pod"yapol'sky, 1968, 1970; Comer, 1984a).

Linear approximation is used to formulate the equations of motion. This means that we neglect the variation of the displacement field on distances comparable with displacement. We also neglect the displacements of the boundary, since these displacements are very small compare to the characteristic wavelengths.

Based on the above approximation and on the main postulates of the fluid dynamics (Landau and Lifschitz, 1980) the linearized equations for the dynamic displacement field in liquid and in elastic medium will be respectively (after Pod"yapol'sky, 1968, 1970):

$$c_f^2 \nabla \mathrm{div}\mathbf{u}_1 - g\mathbf{e}_z \mathrm{div}\mathbf{u}_1 = \frac{\partial \mathbf{u}_1}{\partial t^2} \qquad 0 < z < H \quad (1)$$

$$a^2 \nabla \mathrm{div}\mathbf{u}_2 - b^2 \nabla \times (\nabla \times \mathbf{u}_2) - g\mathbf{e}_z \mathrm{div}\mathbf{u}_2 = \frac{\partial^2 \mathbf{u}_2}{\partial t^2} \qquad z > H, \quad (2)$$

Boundary conditions at the ocean surface ($z = 0$) and the bottom ($z = H$) are:

$$c_f^2 \mathrm{div}\mathbf{u} - gW \big|_{z=0} = 0, \quad (3)$$

$$W_1(H) = W_2(H), \quad (4)$$

$$p_f\left[c_f^2 \mathrm{div}\mathbf{u} - gW_1 \big|_{z=H}\right] = \rho\left[\left(a^2 - 2b^2\right)\mathrm{div}\mathbf{u} + 2b^2\frac{\partial W_2}{\partial z} - gW_2 \big|_{z=H}\right]$$

$$\mu\left(\frac{\partial W_2}{\partial x} + \frac{\partial U_2}{\partial z}\right)\bigg|_{z=H} = 0,$$

$$\mu\left(\frac{\partial V_2}{\partial z} + \frac{\partial W_2}{\partial y}\right)\bigg|_{z=H} = 0, \quad (5)$$

At infinity the displacement vanishes:

$$\mathbf{u} \to 0 \quad \text{at} \quad z \to 0. \quad (6)$$

Here c_f is the velocity of the acoustic wave in the liquid layer, ρ_f is the density of the liquid, H is the thickness of the liquid layer, λ and μ are the Lame coefficients in the elastic half-space, ρ is the density of the elastic medium, a and b are the velocities of P- and S-waves, $\mathbf{u}_i = (U_i, 0, W_i)$ is the displacement vector ($i = 1$ liquid layer, $i = 2$ elastic half-space), g is gravity, is the angular frequency, and z is the vertical distance from the initial (undisturbed) position of the free surface (positive downwards).

3. Solution

The solution of Eqs. (1) and (2) may be written in the form of a stationary plane wave propagating along the x axis

$$\mathbf{u}_i(x, z, \omega, t) = \mathbf{v}_i(z, \omega) \exp\left[i\omega(t - x/c)\right]. \quad (7)$$

Substituting (7) into Eqs. (1) and (2) and introducing the boundary condition at $z \to \infty$, the wave amplitudes as a function of depth are found as

$$U_1(z, \omega) = \frac{ic_f^2}{\omega c}\big[-B\exp(-\eta_2 z/c_f) + C\exp(-\eta_1 z/c_f)\big], \quad (8)$$

$$W_1(z, \omega) = -\frac{c_f}{\omega^2}\big[-\eta_1 B\exp(-\eta_2 z/c_f) + \eta_2 C\exp(-\eta_1 z/c_f)\big], \quad (9)$$

where, for the liquid layer,

$$\eta_1 = -\omega\gamma - g/2c_f \quad (10)$$

$$\eta_2 = -\omega\gamma - g/2c_f \quad (11)$$

$$\gamma^2 = \frac{c_f^2}{c} - 1 + \frac{g^2}{4c_f^2\omega^2}, \quad (12)$$

and for the half-space,

$$U_2(z, \omega) = \frac{ia^2}{\omega c} D\exp(-\omega\alpha(z - H)/a) - \frac{ib\beta}{\omega} F\exp(-\omega\beta(z - H)/b), \quad (13)$$

$$W_2(z, \omega) = \frac{a\alpha}{\omega}\left(1 + \frac{g}{2\omega\alpha a}\right) \cdot D\exp\left(\left(-\omega\alpha + \frac{g}{2a}\right)(z - H)/a\right) - \frac{b^2}{\omega c} F\exp(-\omega\beta(z - H)/b), \quad (14)$$

where

$$\alpha^2 = a^2/c^2 - 1 + \frac{g^2}{4a^2\omega^2}$$

and

$$\beta^2 = b^2/c^2 - 1.$$

Equations (3) to (14) yield a set of four homogeneous equations in four unknown coefficients, namely B, C, D and F, which has a non-trivial solution when the determinant is zero.

The dispersion equation is obtained as follows:

$$\begin{aligned} m\Big[& -\gamma\cosh\left(\omega\gamma H/c_f\right) + \frac{g}{2c_f\omega}\left(\frac{2c_f^2}{c^2} - 1\right)\sinh\left(\omega\gamma H/c_f\right)\Big] \\ & \cdot\left[-a^2\left(1 - \frac{2b^2}{c^2} - \frac{g^2}{2\omega^2 a^2}\right) + 4b^3 a\alpha\beta/c^2\right] \\ & - c_f\left(a\alpha - \frac{g\beta^2}{2\omega}\right) \\ & \cdot\left[\frac{g\gamma}{c_f\omega}\cosh(\omega\gamma H/c_f) - \left(1 - \frac{g^2}{2c_f^2\omega^2}\right) \cdot\sinh\left(\omega\gamma H/c_f\right) - \frac{g\gamma}{c_f\omega}\exp(-gH/2c_f^2)\right] \\ & - \frac{ga\alpha}{\omega^2\left(1 - \frac{c_f^2}{c^2}\right)} \\ & \cdot\Big[\omega\gamma\exp\left(-gH/2c_f^2\right) - \omega\gamma\cosh\left(\omega\gamma H/c_f\right) - \frac{g}{2c_f}\left(\frac{2c_f^2}{c^2} - 1\right)\sinh\left(\omega\gamma H/c_f\right)\Big] = 0, \end{aligned} \quad (15)$$

where

$$m = \frac{\rho}{\rho_f}.$$

We follow the approach of surface-wave theory. Consider the displacement of a stationary surface wave excited by a point source in a homogeneous half-space (Aki, 1980; Keilis-Borok, 1989):

$$\mathbf{u}(r, \varphi, \omega, t) = \frac{\exp(-i\pi/4)}{\sqrt{8\pi}}\frac{\exp(i\omega(t - r/c))}{\sqrt{\omega \quad r/c}} \cdot \frac{\mathrm{U}(z, \omega)}{\sqrt{cuI_0}}\frac{\mathrm{Q}(h, \varphi, \omega)}{\sqrt{cuI_0}}; \quad (16)$$

for a constant thickness of the liquid layer, or

$$\mathbf{u}(r, \varphi, \omega, t) = \frac{\exp(-i\pi/4)}{\sqrt{8\pi}} \cdot \frac{\exp\left(i\omega\left(t - \int_M^N \frac{dx}{c(\omega, x)}\right)\right)}{\sqrt{\omega J(x)/c}} \cdot \left(\frac{\mathbf{U}(z, \omega)}{\sqrt{cuI_0}}\right)_N \left(\frac{\mathrm{Q}(h, \omega)}{\sqrt{cuI_0}}\right)_M, \quad (17)$$

for a variable thickness of the liquid layer, where $\mathbf{U}(z, \omega) = U(z, \omega)\mathbf{e}_r + W(z, \omega)\mathbf{e}_z$, $\mathrm{Q}(h, \varphi, \omega) = m_{rs}(\omega)B_{rs}(h, \varphi, \omega)$ is the excitation function, $m_{rs}(\omega)$ is the spectrum of the seismic moment tensor, $B_{rs}(h, \varphi, \omega)$ is a tensor which can be expressed via the eigenfunctions $U(z, \omega)$ and $W(z, \omega)$ and their derivatives and which depends on the axis orientation of the source,

$$I_0 = \int_0^\infty \rho\left[U^2(z, \omega) + W^2(z, \omega)\right]dz$$

Table 1.

Media density [g/cm^3]		Wave velocities [km/s]		Layer thickness (H-for liquid) [km.]; depth of source location (h-for half-space) [km.]
Model I				
Liquid layer	1	Acoustic wave	1.45	$H = 4$
Half-space	2.03	P-wave	2.0	$h = 10–30$
		S-wave	1.15	
Model II				
Liquid layer	1	Acoustic wave	1.45	$H = 4$
Half-space	2.5	P-wave	3.9	$h = 10–30$
		S-wave	2.3	
Model III				
Liquid layer	1	Acoustic wave	1.45	$H = 4$
Half-space	3.1	P-wave	7.15	$h = 10–30$
		S-wave	4.1	

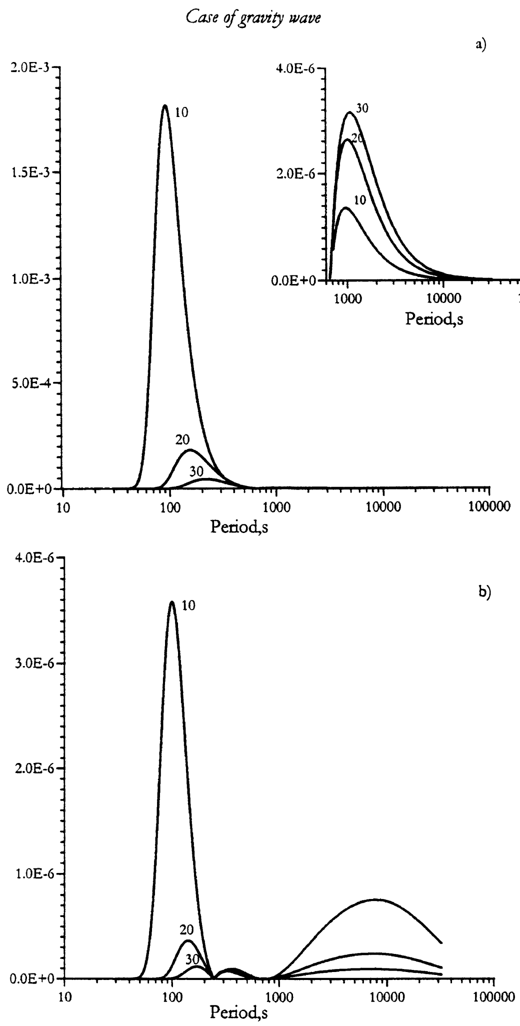

Fig. 1. The dependence of $|q(h, \omega, \varphi)|$ on period for the dip-slip source located a) in non-gravitating half-space, b) in gravitating half-space. Numbers at the curves indicate the source depth in half-space.

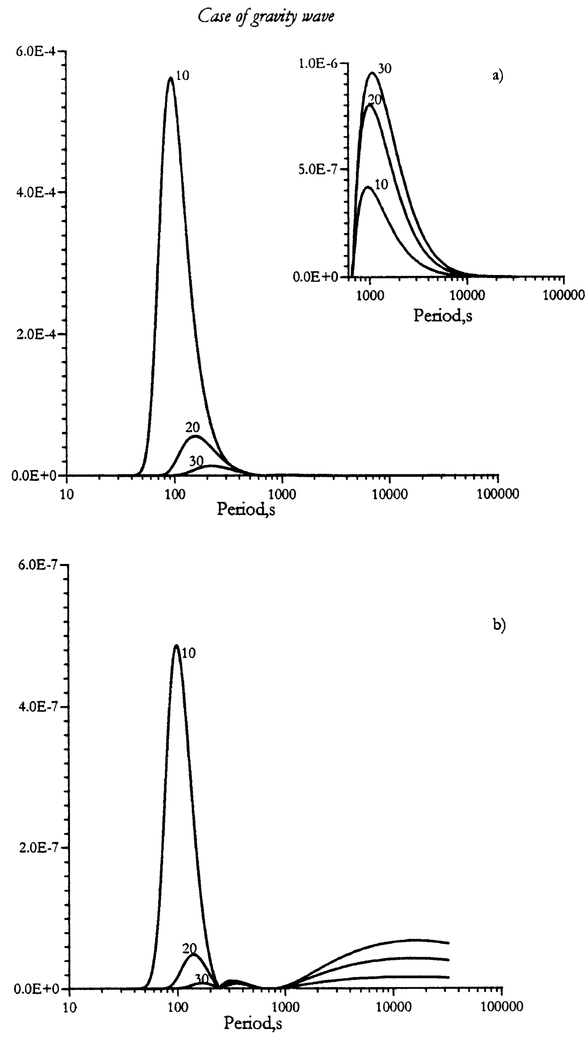

Fig. 2. The dependence of $|q(h, \omega, \varphi)|$ with period for dip-slip source located a) in non-gravitating half-space, b) in gravitating half-space. Numbers at the curves indicate the source depth in half-space.

is the energy integral, u is the group velocity and $J(x)$ is the geometrical divergence in the x-y plane.

In Eqs. (16) and (17), the second term describes the effect of geometrical divergence of the energy flow on the wave propagation. The third term depends on the depth of the receiver, and the fourth term depends on depth, focal mechanism and radiation spectrum in the half-space.

As we shall see, the presence of gravity in the half-space leads to new properties of the gravity wave spectrum. In the following we consider the case of surface waves excited by a dip-slip point source.

4. Dip-Slip Source

Let the fault plane be orthogonal to the x-axis and let displacement be vertical. Then the seismic moment tensor has two non-zero components:

$$m_{xz} = m_{zx} = M_0 F(\omega), \tag{18}$$

where M_0 is the seismic moment and $F(\omega)$ is the seismic moment spectrum. The corresponding components of tensor **B** are

$$B_{xz} = B_{zx} = \frac{i\cos\varphi}{2}\left(-W_2 + \frac{dU_2}{dz}\right). \tag{19}$$

Thus

$$Q(h,\omega,\varphi) = iM_0\cos\varphi\left(-\xi W_2 + \frac{dU_2}{dz}\right)F(\omega), \tag{20}$$

$$q(h,\omega,\varphi) = \frac{i\cos\varphi\left[-\xi W_2(h,\omega) + \left.\dfrac{dU_2(z,\omega)}{dz}\right|_{z=h}\right]}{\sqrt{cuI_0}}, \tag{21}$$

where $\xi = \omega/c$ is the wave number.

To estimate the source radiation function, some value of the seismic moment M_0 and its spectrum $F(\omega)$ must be assumed. We assume values of the radiation function up to the factor M_0 and for that azimuth along which seismic radiation is maximal. Thus

$$F(\omega) = \frac{1}{i\omega(i\omega\tau_0 + 1)}, \tag{22}$$

where τ_0 is the rise time.

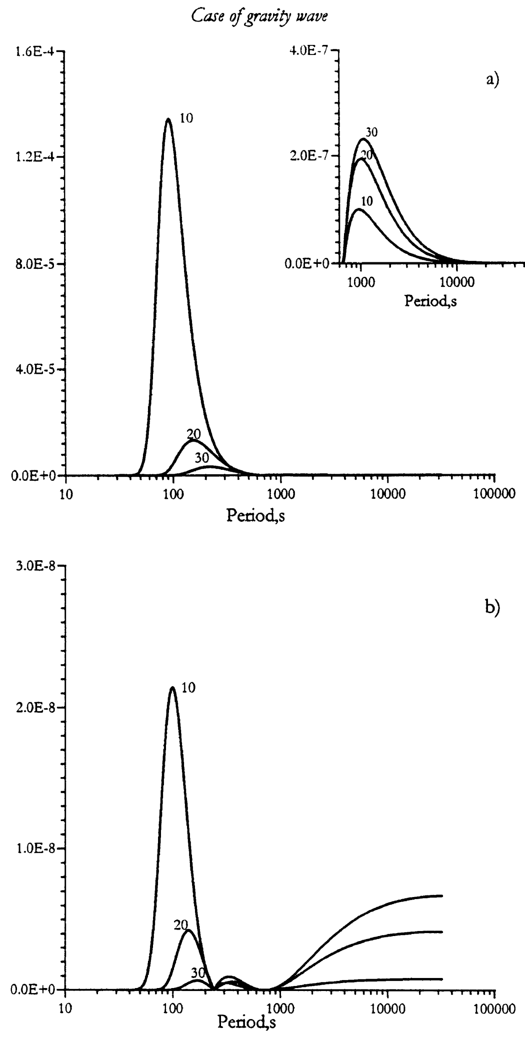

Fig. 3. The dependence of $|q(h,\omega,\varphi)|$ with period for dip-slip source located a) in non-gravitating half-space, b) in gravitating half-space. Numbers at the curves indicate the source depth in half-space.

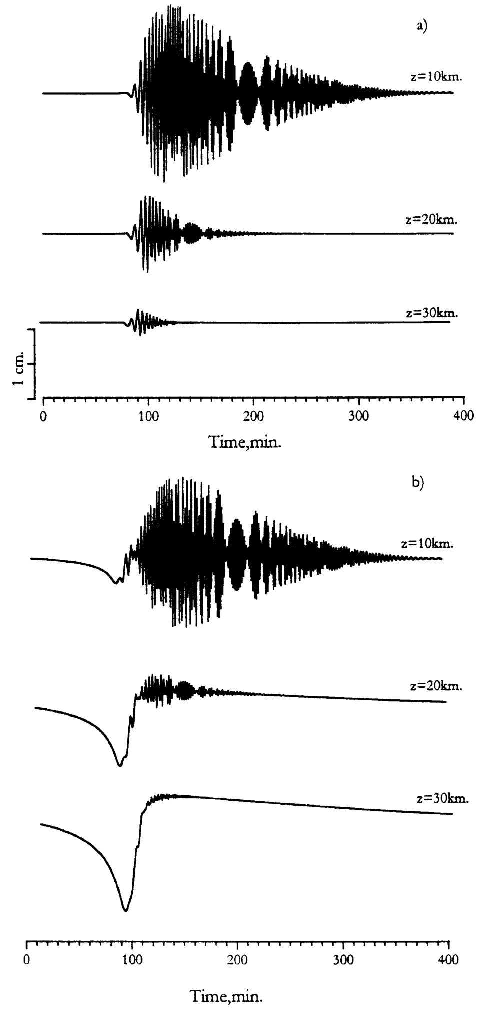

Fig. 4. A comparison of theoretical marigrams of gravity wave from the dip-slip source for model I a)- with non-gravitating half-space, b)- with gravitating half-space. The right numbers at the curves indicate the source depth in half-space.

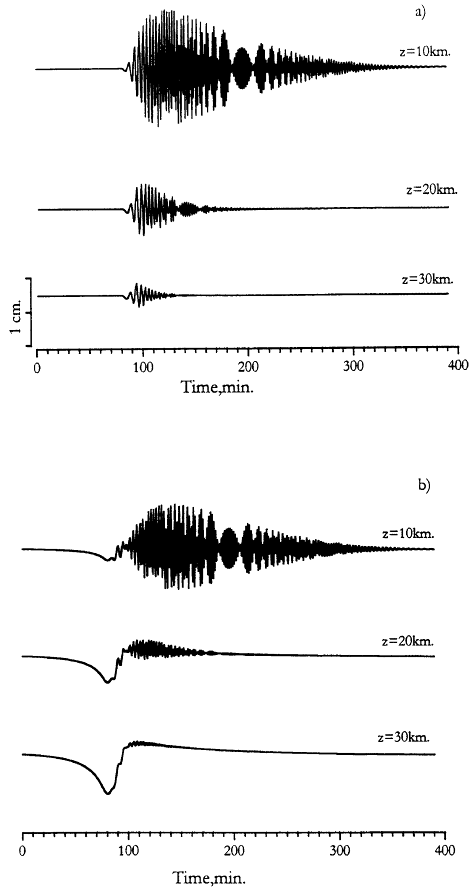

Fig. 5. A comparison of theoretical marigrams of gravity wave from the dip-slip source for model III a)- with non-gravitating half-space, b)- with gravitating half-space. The right numbers at the curves indicate the source depth in half-space.

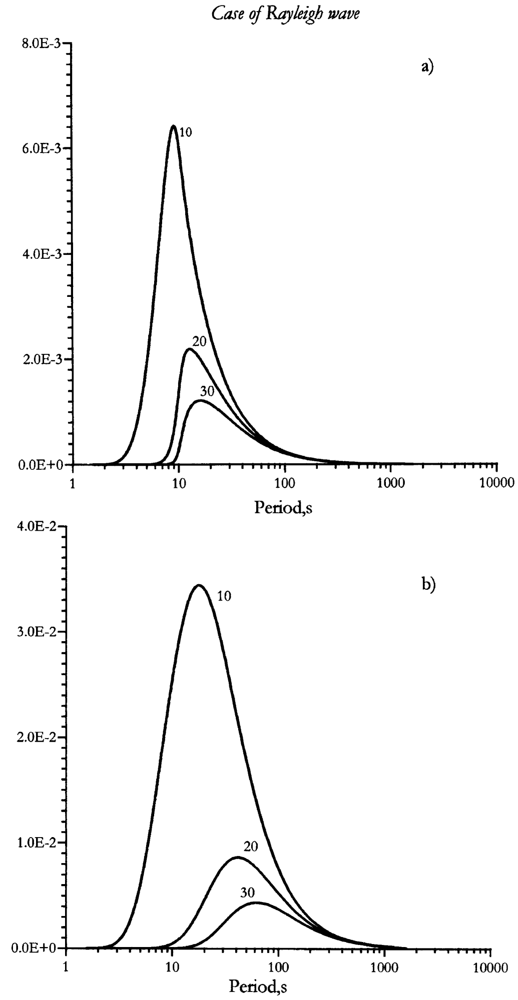

Fig. 6. The dependence of $|q(h, \omega, \varphi)|$ on period for the dip-slip source located a) in gravitating half space model III , b) in gravitating half-space model I. Results without gravitational force are identical. Numbers at the curves indicate the source depth in half-space.

5. Numerical Results

The parameters of our model are shown in Table 1 (Nafe and Drake, 1963).

We focused on the influence of gravitating sediments on excitation of gravity and Rayleigh waves. Although in real Earth thickness of ocean sediments seldom exceeds a few kilometers (Mooney *et al.*, 1998), for our numerical experiment we used structures with thick sedimentary mass like deep-sea trenches (Yoshii *et al.*, 1970; Westbrook *et al.*, 1973), and we vary the source location from shallow to deep. In spite of that we are using real paremeters of the ocean-Earth structure, proposed model is still a theoretical one, whose purpose is to explore some physical aspects of the role of sediments in the excitation of gravity and Rayleigh waves.

5.1 Case of gravity wave

As shown previously (Novikova, 1997), compressibility of the liquid leads to the appearance of two maxima (Figs. 1(a)–3(a)) in the spectrum of gravity waves, namely a high-frequency maximum at periods of 100–200 s and a low-frequency maximum around at 1000 s. For all liquid half-space models (Table 1) in the 10–30 km depth interval, the amplitude of the high-frequency maximum is larger than that of the low frequency maximum. When the half-space is composed of sediments excitation function of gravity wave is one order of magnitude larger compared with that in hard rock half-space (Figs. 1(a)–3(a)). Theoretical marigrams (vertical component) of gravity wave, calculated by integrating formula (16) at the epicentral distance of 1000 km, prove this as well (Figs. 4(a) and 5(a)). We assume that the seismic moment M_0 for the source model is 6×10^{21}Nm, corresponding roughly to an earthquake of magnitude 8.5.

By including the action of gravity in the half-space character of the excitation function spectrum changes. Third maximum appears (Figs. 1(b)–3(b)) and at the deepening of source its amplitude increases. Theoretical marigrams (Figs. 4(b) and 5(b)) show that including gravity in the half-space increases period of the gravity wave excited by deep sources by a factor of two, up to 10 minutes.

One possible explanation of this third maximum is that including gravity in the elastic medium results in the additional (gravity) root in the dispersion equation. This root corresponds to a gravity wave propagating at the liquid-elastic medium interface (Gusiakov, 1972) and cases the change in excitation spectrum.

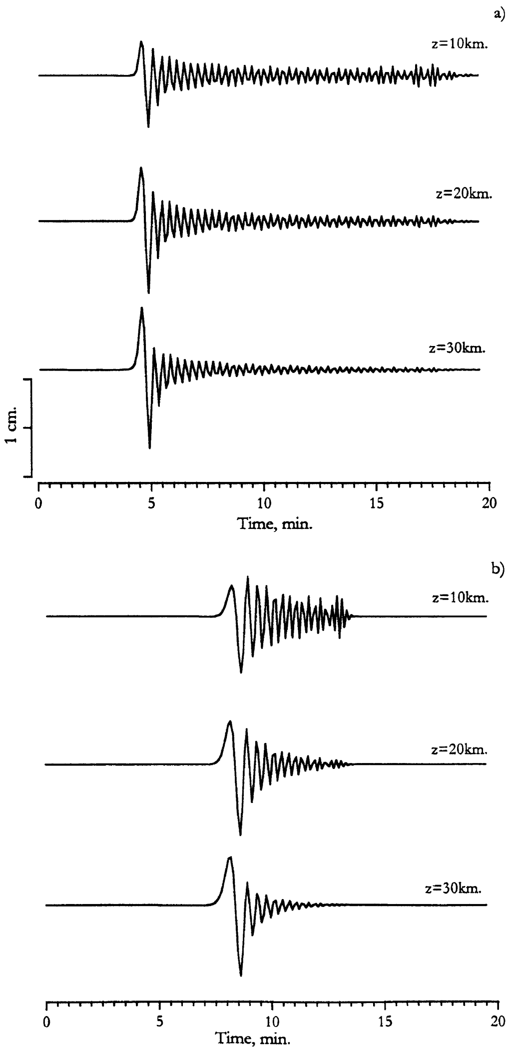

Fig. 7. A comparison of theoretical marigrams of Rayleigh wave from the dip-slip source a) for model III, b) for model II with gravitating half-space. Results without gravitational force are identical. The right numbers at the curves indicate the source depth in half-space.

5.2 Case of Rayleigh wave

Figures 6 and 7 demonstrate that low-rigidity media also amplify the excitation and propagation of Rayleigh waves.

At the deepening of source Rayleigh wave amplitude decreases, similar to the case of gravity wave. Including gravity in the half-space has no effect on the Raylegh wave (excitation function curves and theoretical marigrams with and without gravity coincide). It can be explained as follows. In case of Rayleigh wave the restoring force is elastic, not gravity. Gilbert (1967) showed that influence of gravity acting in an elastic medium, on a surface wave is important only if the medium is filled with soft sediments with wave velocities $b < 100$ m/s. In our model, however, s-wave velocities vary from 1.1 to 4.1 km/s, i.e. our model medium is rigid enough to suppress action of gravity on surface wave.

6. Conclusions

Excitation and propagation of gravity and Rayleigh waves are treated in the frame of model considering them as surface waves of the coupled type in a liquid layer overlying the gravitating half-space filled with sediments. Normal mode formalism was applied to calculate far-field displacement of surface waves generated by dip-slip point source located at different depths in the solid Earth underneath the ocean.

Using the liquid-solid model with different parameters of half-space we found that low-rigidity medium may be seen as an amplifier for both gravity and Rayleigh waves.

Including gravity in the half-space changes the character of excitation spectrum of gravity wave. The difference is more obvious for the gravity wave excited by a deep source in the weak media. In such conditions gravity wave becomes more long-period.

Proposed model is the theoretical one aimed to study the influence of gravitating sediments on the wave excitation. Consequently, for future investigations, it is assumed to use a more realistic structure with layered ocean crust. In particular, such model will be important for the study of Rayleigh waves that depend considerably on the layered structure (especially in the 1–10 seconds period range).

In further study of gravity waves, the results of the present work will be generalized for the case of "tsunami-earthquake" events, using a more realistic model of a finite size seismic source. Recent investigations (Houston, 1999; Bilek and Lay, 1999) suggested that some earthquakes rupture slowly because of the presence of shallow, unconsolidated sediments. The process of gravity wave excitation by a slow rupture on a shallow fault may be relevant, as slow rupture can generate large low-frequency gravity waves which take the form of tsunami waves at the ocean surface.

Acknowledgments. The authors would like to thank Dr. C. Lomnitz and an anonymous reviewer for suggestions that helped to improve an original version of the manuscript. This research was supported by a grant of Academia Sinica Institute of Earth Sciences, Taiwan.

References

Aki, K. and G. Richards, *Quantative Seismology*, v.1., 557 pp., W. H .Freeman, San Francisco, 1980.

Alexeev, A. and V. Gusiakov, Numerical modeling of tsunami and seismic surface wave generation by a submarine earthquake, in *Tsunami Research Symposium, Bull. R. Soc. N. Z.*, **15**, 243–251, 1976.

Bilek, S. and T. Lay, Rigidity variations with depth along interplate megathrust faults in subduction zones, *Nature*, **400**, 443–446, 1999.

Comer, R., The tsunami mode of a flat earth and its excitation by earthquake sources, *Geophys. J. R. astr. Soc.*, **77**, 1–27, 1984a.

Comer, R., Tsunami generation: a comparison of traditional and normal mode approaches, *Geophys. J. R. astr. Soc.*, **77**, 29–41, 1984b.

De, S. N. and P. R. Sengupta, Surface waves under the influence of gravity, *Gerlands Beitr. Geophys.*, **85**, 311–318, 1976.

Gilbert, F., Gravitationally perturbed elastic waves, *Bull. Seism. Soc. Am.*, **57**, 783–794, 1967.

Gusiakov, V., *Excitation of Tsunami and Oceanic Rayleigh Waves by Submarine Earthquake. Mathematical Problems in Geophysics*, pp. 250–267, Novosibirsk, 1972.

Houston, H., Slow ruptures, roaring tsunamis, *Nature*, **400**, 409, 1999.

Hwang, L.-S., H. L. Butler, and D. J. Divoky, Tsunami model: Generation and open-sea characteristics, *Bull. Seism. Soc. Am.*, **62**, 1579–1596, 1972.

Kanamori, H., Mechanism of tsunami earthquakes, *Phys. Earth Planet. Inter.*, **6**, 346–359, 1972.

Keilis-Borok, V., *Seismic Surface Waves in a Laterally Inhomogeneous Earth*, 293 pp., Kluwer Academic Publishers, 1989.

Landau, L. and E. Lifshitz, Fluid mechanics, in *Courses of Theoretical Physics*, v.6, 1980.

Lomnitz, C., Some observations of gravity waves in the 1960 Chile earthquake, *Bull. Seism. Soc. Am.*, **59**, 669–670, 1970.

Lomnitz, C., Mexico 1985: the case for gravity waves, *Geophys. J. Int.*,

102, 569–572, 1990.

Lomnitz, C., On the transition between Rayleigh waves and gravity waves, *Bull. Seism. Soc. Am.*, **81**, 273–275, 1991.

Matuzawa, T., On the possibility of the gravitational waves in soil and allied problems, *J. Inst. Astr. Geophys. Tokyo*, **3**, 161–174, 1925.

Mooney, W., G. Laske, and T. Masters, Crust 5.1: A global crustal model at $5^0 x 5^0$, *J. Geophys. Res.*, **103**, 727–747, 1998.

Nafe, J. and C. Drake, Physical properties of marine sediments, in *The Sea*, vol. 3, edited by M. H. Hill, pp. 794–815, Interscience Publishers, New York, 1963.

Novikova, T., Numerical modeling of the tsunami generation by seismic sources, Ph.D.thesis Earth Physics Department, Institute of Physics, St. Petersburg University, 98 pp., 1997.

Okal, E., Seismic parameters controlling far-field tsunami amplitudes: a review, *Natural Hazards*, **1**, 67–96, 1988.

Pod"yapol'sky, G. S., Excitation of a long gravitational wave in the ocean from a seismic source in the crust, *Izv. AN SSSR, Fizika Zemli*, **1**, 1968 (in Russian).

Pod"yapol'sky, G. S., Generation of the tsunami wave by the earthquake in Tsunamis in the Pacific Ocean, edited by W. M. Adams, pp. 19–32, East-west Center Press, Honolulu, 1970.

Satake, K., The mechanism of the 1983 Japan Sea earthquake as inferred from long-period Surface waves and tsunamis, *Phys. Earth Planet. Inter.*, **37**, 249–260, 1985.

Ward, S., Relationships of tsunami generation and an earthquake source, *J. Phys. Earth*, **28**, 441–474, 1980.

Ward, S., On tsunami nucleation: a point source, *J. Geophys. Res.*, **86**, 7895–7900, 1981.

Ward, S., On tsunami nucleation: an instantaneous modulated line source. *Phys. Earth Planet. Inter.*, **27**, 273–285, 1982.

Weidner, D., Rayleigh waves from mid ocean ridge earthquakes: source and path effects, Ph.D. thesis, Harvard College, 253 pp., 1967.

Westbrook, G. *et al.*, Lasser Antilles subduction zone in the vicinity of Barbados, *Nature Phys. Sci.*, **244**, 118–120, 1973.

Yamashita, T. and R. Sato, Generation of tsunami by a fault model, *J. Phys. Earth*, **22**, 415–440, 1974.

Yamashita, T. and R. Sato, Correlation of tsunami and sub-oceanic Rayleigh wave amplitudes. Possibility of the use of Rayleigh wave in tsunami warning system, *J. Phys. Earth*, **24**, 397–416, 1976.

Yoshii, Y. *et al.*, Crustal structure of Tosa deep-sea terrace and Nankani trough (in Japan), in *Island Arc and Ocean*, edited by M. Hoshino and H. Aoki, pp. 93–103, Tokai University Press, Tokyo, 1970.

T. Novikova (e-mail: tatyana@earth.sinica.edu.tw), K.-L. Wen, and B.-S. Huang