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Background Love and Rayleigh waves simultaneously generated at the Pacific Ocean floors

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[1] Earth's background free oscillations known as Earth's hum have been interpreted as the Earth response to vertical pressure loads due to atmospheric and/or oceanic disturbances. Such excitation mechanisms, however, can hardly excite Love waves. Here we show clear evidence of background Love waves from 0.01 to 0.1 Hz, based on the array analysis of tiltmeters in the Japanese islands. The observed kinetic energy of Love waves is as large as that of Rayleigh waves through the whole period of analysis. The predominant incident azimuths are common to the Love and Rayleigh waves, the strongest in directions along ocean-continent borders, next from deep ocean floors and the weakest from continents. These observations indicate that background Love and Rayleigh waves are largely generated by the same mechanisms other than vertical pressure loading. We suggest that the most likely excitation source is shear traction acting on a sea-bottom horizon due to linear topographic coupling of infragravity waves. **Citation:** Nishida, K., H. Kawakatsu, Y. Fukao, and K. Obara (2008), Background Love and Rayleigh waves simultaneously generated at the Pacific Ocean floors, *Geophys. Res. Lett.*, *35*, L16307, doi:10.1029/2008GL034753.

1. Introduction

[2] Background Rayleigh waves from 2 to 20 mHz known as background free oscillations have now been firmly established [*Nawa et al.*, 1998; *Suda et al.*, 1998; *Kobayashi and Nishida*, 1998]. Their root-mean-square amplitudes are on the order of 5 ngal (1 ngal = 10^{-11} ms^{-2}) with little frequency dependence. Cumulative effects of many small earthquakes are too small to explain the amplitudes [*Suda et al.*, 1998; *Kobayashi and Nishida*, 1998]. Statistical examination of the excited normal modes [*Nishida and Kobayashi*, 1999] indicates that these oscillations must be excited randomly and persistently by globally distributed sources. The intensities of these modes clearly show annual and semiannual variations with the largest peak in July and a secondary peak in January [*Nishida et al.*, 2000; *Tanimoto and Um*, 1999; *Ekström*, 2001]. The observed amplitudes of the modes that are coupled with the atmospheric free oscillations are anomalously large relative to the adjacent modes [*Nishida et al.*, 2000]. All of these observations suggest that atmospheric and/or oceanic disturbance may be the source for this phenomenon. Assuming that atmospheric pressure disturbance acting on the Earth's surface is a primary excitation source, some quantitative comparison has been made between the atmospheric pressure disturbance and Earth's background free oscillations [*Kobayashi and Nishida*, 1998; *Fukao et al.*, 2002].

[3] Shortly after the discovery, pressure changes at the ocean bottom due to oceanic infragravity waves were suggested to be the probable excitation sources [*Watada and Masters*, 2001]. *Rhie and Romanowicz* [2004] found that the excitation sources are dominated in the north Pacific ocean in winter of the northern hemisphere and in the southern hemisphere near the Antarctica in winter of the southern hemisphere. By comparing this result to the oceanic wave height data they concluded that the most probable excitation source is oceanic disturbance, substantiating the hypotheses of the excitation by ocean infragravity waves through their nonlinear interaction [*Tanimoto*, 2007; *Webb*, 2007]. However, the proposed mechanisms work efficiently only in shore regions, and hence may not be consistent with the observed spatial extent of the excitation sources [*Nishida and Fukao*, 2007]. At present there is little consensus about the excitation mechanism that can explain all of the observed features.

[4] On the other hand, the excitation mechanism of background Rayleigh waves from 0.05 to 0.2 Hz, known as microseisms, is established more firmly. Microseisms are identified at the primary and double frequencies: The primary microseisms at around 0.08 Hz have been interpreted as being caused by direct loading of ocean swell onto a sloping beach [*Haubrich et al.*, 1963]. The typical frequency of the secondary microseisms at about 0.15 Hz approximately doubles the typical frequency of ocean swells, indicating the generation of the former through nonlinear wave-wave interaction of the latter [*Longuet-Higgens*, 1950].

[5] The mechanisms of background Rayleigh waves so far proposed commonly assume random pressure disturbances, either in the atmosphere or oceans, which cannot generate Love waves if the Earth is spherically stratified. There are, however, several reports of background Love waves in the microseismic bands, where the energy ratio of Love to Rayleigh waves is much higher for the primary than for the secondary microseisms [*Friedrich et al.*, 1998]. Although these observations suggest some mechanisms other than vertical pressure forces to excite background surface waves, noisy horizontal records of long–period seismometers have prevented us from detecting Love waves below 0.05 Hz. The most recent data analysis at the quietest sites has revealed existence of background Love waves even in the free oscillation band from 3 to 7 mHz [*Kurrle and*

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Widmer-Schnidrig, 2008], yet their excitation mechanism still remains enigmatic. In order to understand why background Love waves are excited as efficiently as Rayleigh waves, we examine temporal and azimuthal changes of these waves, based on an array analysis of the dense Hi-net tiltmeter array over the Japanese islands [*Obara et al.*, 2005].

2. Data Analysis

[6] The Hi-net tiltmeter network, operated by the National Research Institute for Earth Science and Disaster Prevention, consists of 679 tiltmeters (Figure 1a) buried in deep boreholes of 100 m depth or more [*Okada et al.*, 2004]; it can be used as a network of horizontal long-period seismometers [*Tonegawa et al.*, 2006]. For each station, we removed glitches and divided all the records from June 2004 to December 2004 into 1024 s segments with an overlap of 512 s. Each segment was Fourier-transformed. In order to analyze the background wave-field, we discarded outliers such as earthquakes, and instrumental noise.

[7] With an assumption that signals at a station can be represented by a superposition of plane waves, we calculated two-dimensional frequency–slowness spectra in a window from 0.01 to 0.25 Hz [*Rost and Thomas*, 2002; *Nishida et al.*, 2005].

[8] The spectrum at a certain slowness vector is essentially the frequency-domain representation of the sum of all the array records with time delays predicted by that slowness vector. This sum would have a local maximum of amplitude somewhere in the slowness vector space, where signals of all the stations are in phase. The slowness vector that gives the local maximum has a greater uncertainty if it is oriented more perpendicular to the array direction. The two-dimensional frequency-slowness spectrum tends to be sharpened in the direction parallel to the array direction and broadened in its orthogonal direction, accordingly. This effect would bias the directional search for the incident Love and Rayleigh waves. For a quantitative estimation of the azimuthal variation of the amplitudes of the incident waves, we deconvolved their array response functions from the observed spectra using the Lucy-Richardson deconvolution algorithm [*Lucy*, 1974; *Bertero and Boccacci*, 2005]. Figure 1b shows a typical example of the deconvolution at 0.0125 Hz.

[9] Figure 1c shows the spectral values at a frequency of 0.0125Hz in the slowness vector domain averaged over every 60 days from June to December of 2004. The upper and lower three diagrams show the results for the transverse components (perpendicular to the propagation direction) and radial components (along the propagation direction), respectively. The transverse-component spectra show clearly the Love-wave propagation from all the directions although their excitation amplitudes vary with incident azimuth. In Figure 1c they are identified as the circle with a slowness of about 0.22 [s/km]. The radial-component spectra show clearly the Rayleigh wave propagation with a slowness of about 0.26 [s/km]. We can also identify shear-coupled PL waves trapped in the upper mantle and the crust with a slowness of about 0.13 [s/km] [*Oliver*, 1961]. To our surprise, the observed amplitudes of Love waves are about three times (in power spectral densities) as large as those of Rayleigh waves, and their azimuthal variation is similar to that for Rayleigh waves.

[10] Figure 2a shows the incident-azimuth variations of Love and Rayleigh-wave amplitudes at 0.0125 Hz as a function of time at an increment of 7 days. Figure 2a shows the strongest amplitudes in directions along the continental shelf. Waves traveling from the southwest along the Philippine-Ryukyu system are strong in a period from June to November, while strong waves are observed from the northeast along the Aleutian-Kurile trench-arc system in December. Throughout the period, waves traveling from the northwest (from the Eurasian continent) are very weak. We observe significant surface waves from deep seafloor regions (from the south to east). These temporal changes of the incident azimuths are consistent with the estimated spatial distribution of excitation sources of background Rayleigh waves [*Rhie and Romanowicz*, 2004; *Nishida and Fukao*, 2007]. The observed Love and Rayleigh waves from deep seafloors are also consistent with the observed large extent of excitation sources [*Nishida and Kobayashi*, 1999; *Nishida and Fukao*, 2007]. The azimuthal distribution of Love waves coincides with that of Rayleigh waves throughout the observation time. In Figure 2a we also plot the mean power spectral densities of Rayleigh waves from 0.08 to 0.09 Hz, corresponding to primary microseisms, and those from 0.12 to 0.13 Hz, corresponding to secondary microseisms. Their temporal variations coincide approximately with those of the mean amplitudes of Love and Rayleigh waves around 0.0125 Hz. Such coincidence suggests that low-frequency background Love and Rayleigh waves may have a common origin with the microseisms which are known to be strongly correlated with the activities of oceanic infragravity waves [*Okeke and Asor*, 2000; *Darbyshire and Okeke*, 1969].

[11] In order to discuss an excitation mechanism of background Love and Rayleigh waves, we estimated their total kinetic energies from 0.01 to 0.025 Hz. We first calculated the mean square amplitudes by integrating the spectral values within a narrow band around a constant-slowness circle for every 7 days segment. Here, we extrapolated the amplitudes at the surface to depths assuming that they can be represented by eigenfunctions of fundamental spheroidal modes for a spherically symmetric Earth structure [*Dziewonski and Anderson*, 1981]. We then estimated kinetic energy by integrating them in the whole Earth with an assumption of spatially homogeneous excitation of background surface waves. Figure 2b shows that the kinetic energy of Love waves is comparable to that of Rayleigh waves. The energy ratio of Love to Rayleigh waves is approximately 1 and depends little on time. We also plot the energy ratio against frequency in Figure 2c. The energy ratio of the Love to Rayleigh waves is around 1.2 at frequencies well below 0.1 Hz. The energy ratio suddenly decreases at 0.1 Hz and remains low at the higher frequencies. Such a sudden change of the energy ratio within the microseismic bands has also been reported in Europe [*Friedrich et al.*, 1998]. Microseisms above 0.1 Hz (secondary microseisms) have amplitudes an order of magnitude larger than those of primary microseisms and are known to be exited through nonlinear wave-wave interactions of ocean swells near the sea surface [*Longuet-Higgens*, 1950]. These interactions can be regarded as pressure

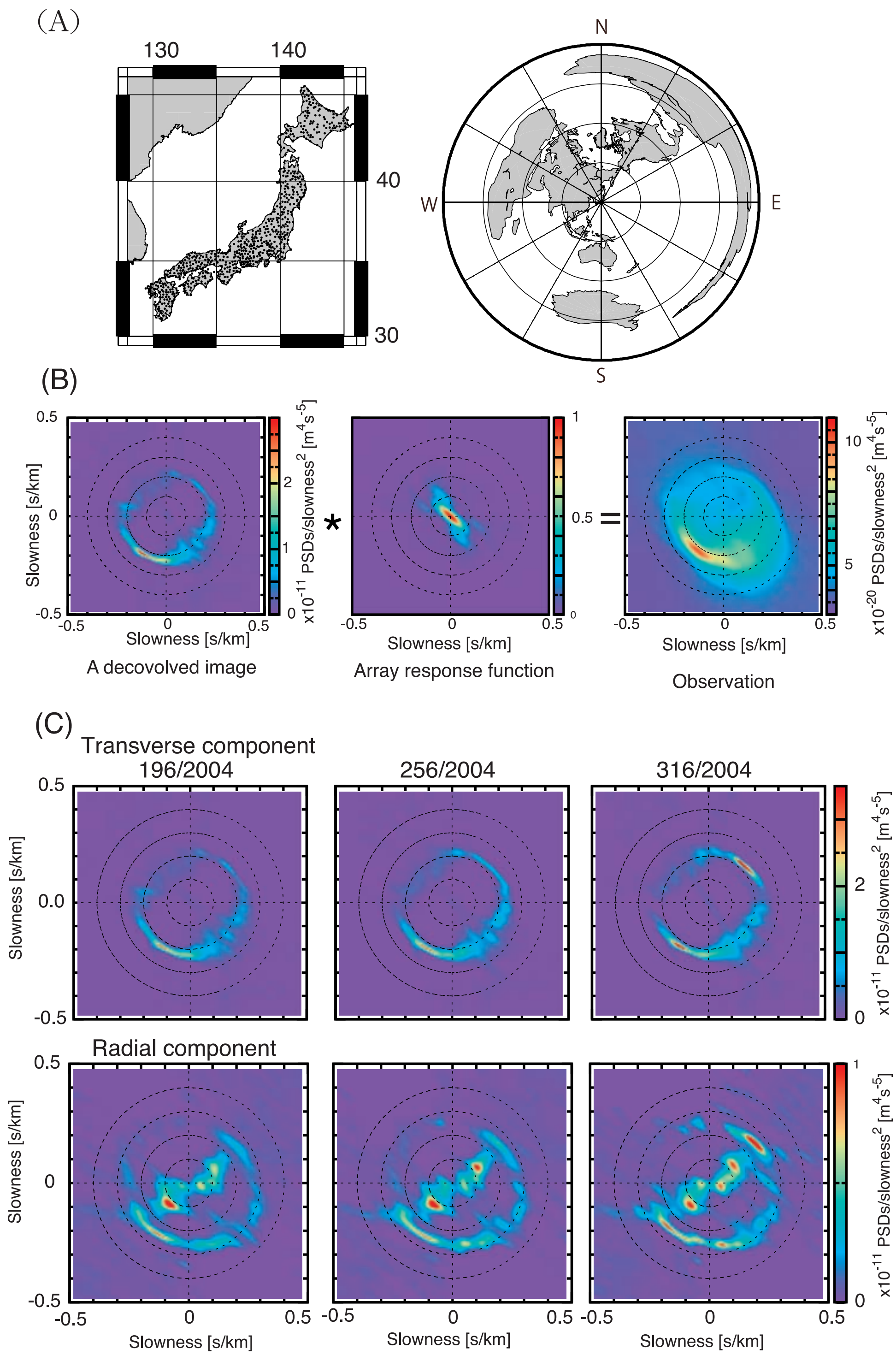

Figure 1. (a) Location map of 679-Hi-net tiltmeters and distribution of continents and oceans in the azimuthal projection from the center of the Hi-net array. (b) A resultant deconvolved image (radial components), calculated array response function at 0.0125 Hz for the Hi-net array and a typical example of observation of background Rayleigh waves at 0.0125 Hz in a time period from 166/2004 to 226/2004. (c) Frequency–slowness spectra at 0.0125 Hz, calculated for every 60 days from 166/2004-346/2004.

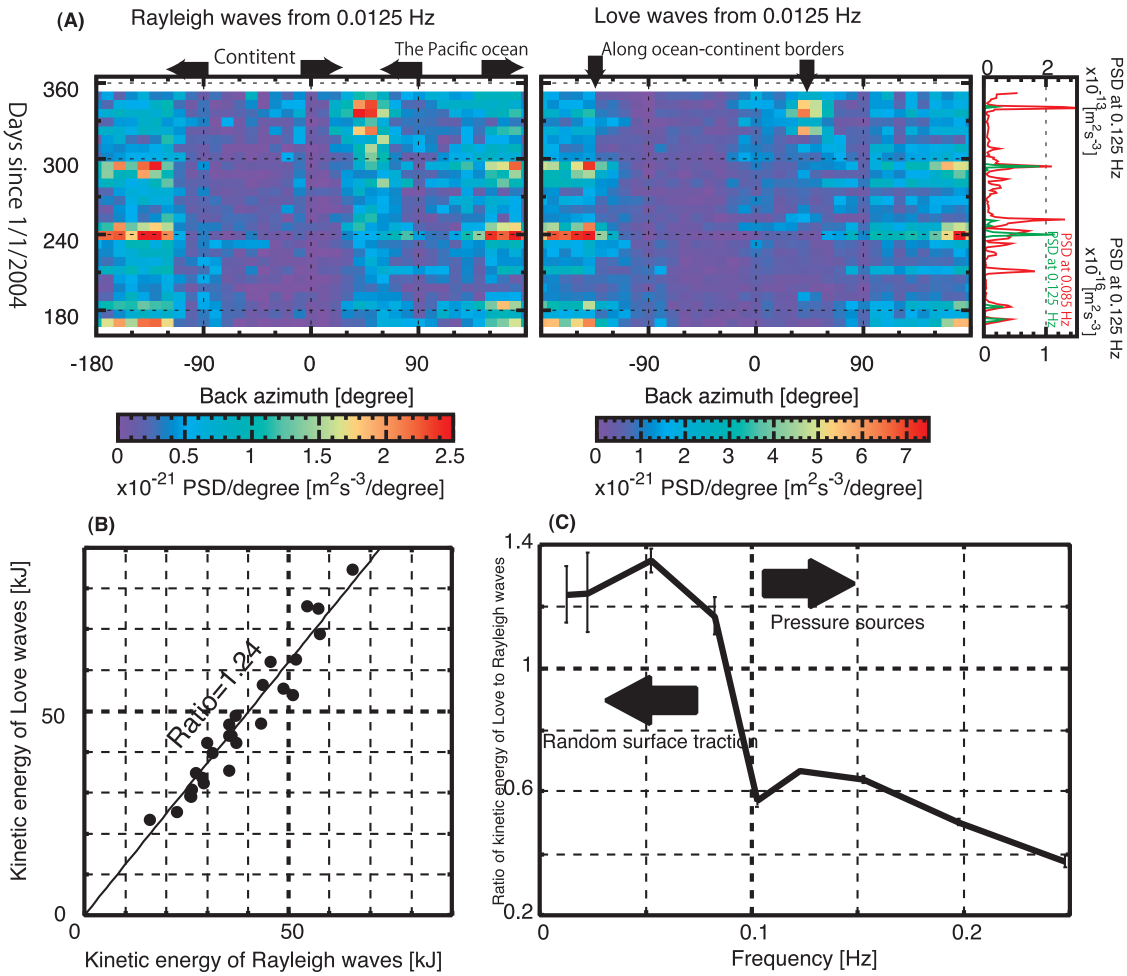

Figure 2. (a) Azimuthal variations of Love and Rayleigh-wave amplitudes at 0.0125 Hz as functions of time showing the similar azimuthal patterns. The right column indicates the temporal change of amplitudes of primary microseisms in red (mean power spectral densities from 0.08 to 0.09 Hz), and secondary microseisms in green (and those from 0.12 to 0.125 Hz) showing the activity pattern similar to those of Love and Rayleigh waves at 0.0125 Hz. (b) Relation between kinetic energies of Love and Rayleigh waves at 0.0135 Hz for every 7 days. The energy ratio of Love to Rayleigh waves is approximately constant, about 1.2. (c) Energy ratio of Love to Rayleigh waves as a function of frequency (from 0.01 to 0.25 Hz). Below 0.1 Hz the ratio is about 1. There is a sudden decrease at 0.1 Hz above which the energy ratio is about 0.5. This sudden change in energy ratio indicates a change of excitation mechanism across 0.1 Hz.

sources near the sea surface so that they hardly excite Love waves in principle. This pressure mechanism is consistent with the observed Love to Rayleigh wave energy ratio above 0.1 Hz.

3. Topographic Coupling on the Seafloor

[12] We have shown that the kinetic energies of background Love and Rayleigh waves below 0.1 Hz are almost equal. For a spherically symmetric Earth, any pressure sources either in the atmosphere or oceans cannot excite Love waves to first order. Primary microseisms from 0.05 to 0.1 Hz have been interpreted as Rayleigh waves directly generated by pressure load of infragravity waves acting on a sloping coast [*Darbyshire and Okeke*, 1969]. Such a model relying on vertical pressure load cannot be applied straightforwardly to generation of Love waves. Very weak signal of Love waves from the continent, significant signal from the Pacific abyssal floors and very strong signal from the ocean-continent borders consistently imply the excitation of Love waves at the ocean bottom most probably by the topographic coupling with ocean infragravity waves.

[13] Below 0.03 Hz, nonlinear interaction of oceanic infragravity waves has been proposed as the excitation source of background Rayleigh waves [*Tanimoto*, 2007; *Webb*, 2007]. If this mechanism is a dominant one, the typical frequency of background Rayleigh waves at about 10 mHz [*Peterson*, 1993] must approximately double the typical frequency of ocean infragravity waves through nonlinear wave-wave interactions [*Longuet-Higgins and Stewart*, 1962]. The observed typical frequency of oceanic infragravity waves is also about 10 mHz [*Webb*, 1998], however. Coincidence of the typical frequency suggests that their dominant mechanism is not the nonlinear interactions but linear coupling. Coincidence of kinetic energy of background Love and Rayleigh waves shows that their

likely excitation source is shear traction acting on a sea-bottom horizon.

[14] If infragravity waves are low enough in frequency ($\leq$0.03 Hz) and long enough in wavelength ($\geq$10 km), they cause pressure fluctuations on the seafloor. The resultant traction acting on a slope of the seafloor topography (e.g., seamounts with height of a few hundreds of meter and width of few tens of kilometer) has a horizontal component which acts as shear traction on the presumed bottom horizon. This shear traction fluctuates in time with the same frequency as the infragravity wave but can fluctuate in space with a wavelength long enough more than 100 km) to generate Love and Rayleigh waves simultaneously by topographic coupling where the dominant wavelengths of the infragravity wave and seafloor topography almost coincide with each other. This topographic coupling mechanism should work not only in shallow seas but also in abyssal basins. A stronger topographic coupling can be expected in shallow seas because of larger wave amplitudes and more rugged topography.

[15] On the other hand, the topographic-coupling mechanism alone is unable to explain the observed coupling of background free oscillations between the atmosphere and the solid Earth [*Nishida et al.*, 2000]. Nonlinear wave-wave interactions of infragravity waves may generate low-frequency acoustic waves but at a level much lower than the observed amplitudes [*Webb*, 2007]. The observation of acoustic/seismic coupling clearly indicates that there is some resonant mechanism between atmospheric acoustic free oscillations generated by lowermost atmospheric turbulences and Earth's free oscillations generated by oceanic infragravity waves through topographic coupling. The phenomena of background free oscillations should be understood as those in a single system of the atmosphere, ocean and solid Earth.

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