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Excitation of microseisms

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[1] Excitation of microseisms is generally considered to be due to pressure change at ocean bottom, for which Longuet-Higgins derived his celebrated formula in 1950. Use of this formula is an approximation, however. Comparison with a more rigorous normal-mode formula shows that this conventional approach is acceptable for ocean depths less than 1 km but fails in deep oceans. On the other hand, there seems to be a multitude of evidence that source region for double-frequency microseim is near the coast and thus is generally in shallow water. An evidence from buoy data for nonlinearity in ocean waves is presented to support this view. If a source region is in shallow water, use of the Longuet-Higgins pressure formula at ocean bottom for the excitation of microseisms is justified, although one should pay attention to ocean depths very carefully. **Citation:** Tanimoto, T. (2007), Excitation of microseisms, *Geophys. Res. Lett.*, *34*, L05308, doi:10.1029/2006GL029046.

1. Introduction

[2] Microseisms are ubiquitous in seismograms. Their predominant frequency range varies from region to region but it can be bracketed within the frequency band between 0.05 and 0.4 Hz almost anywhere in the world. Because they are excited by ocean waves, microseisms show seasonal changes in amplitudes, reflecting the vigor of oceanic activities, and thus may be a good tool to monitor Earth's near-surface environment.

[3] Our view on the generation of microseisms has been dominated by the nonlinear mechanism, pointed out by *Longuet-Higgins* [1950] and expanded by *Hasselmann* [1963]. The main point of the mechanism is that, if a standing wave exists because of collision of ocean waves near the coast (due to reflection of waves) or near the eye of a low-pressure system, pressure change occurs at ocean bottom even though the original ocean waves do not have energy at ocean bottom. This mechanism predicts the main energy to be at twice the frequency of colliding ocean waves, which explains spectral behavior of seismic signals quite well. Excitation of microseisms is thus often viewed that, first, ocean waves collide and generate pressure at ocean bottom and second, this pressure change in turn generates seismic waves as though the solid Earth is being hit at ocean bottom.

[4] The aim of this paper is two-fold; first, we will point out that this conventional view of microseism excitation is an approximation which may require some modifications. Our discussion will be based on a more rigorous normal-mode excitation formula derived recently [*Tanimoto*, 2007]. Our results indicate that differences are large for deeper oceans but are not important if a source region is in shallow water of less than 0.5–1 km. Secondly, we will examine the question of excitation between deep vs. shallow oceans, using ocean-wave spectra data from buoys. We find that buoy spectral data tend to imply the excitation region to be in shallow oceans. This is consistent with recent results by several authors based on different evidence [e.g., *Bromirski et al.*, 1999, 2005; *Bromirski and Duennebier*, 2002; *Stehly et al.*, 2006; *Tanimoto et al.*, 2006]. It thus appears that the discrepancy between the Longuet-Higgins approach and the more rigorous normal-mode approach is not so important in practice, although one should carefully take a look at ocean depth in the source region.

2. Normal Mode Excitation

[5] Forcing by the Longuet-Higgins pressure term at ocean bottom is an approximation to the problem of microseism excitation. This should be easy to understand if one thought of excitation of Rayleigh waves, the dominant wave-type in microseisms, in the source region. Rayleigh-wave eigenfunctions have some amplitudes in the ocean, in addition to components in the solid Earth, and generally vary with depth. Therefore, not only the pressure at ocean bottom but also variations of eigenfunctions in the ocean could affect the efficiency of excitation.

[6] A normal-mode formula was derived recently [*Tanimoto*, 2007] by first deriving the body/surface force equivalent for the nonlinearly interacting ocean waves and then applying it to the excitation problem of normal modes (Rayleigh waves). It was shown that the approach by the Longuet-Higgins pressure formula at ocean bottom is correct if the vertical eigenfunction of Rayleigh waves is constant in the ocean. For oceans with depth larger than 3 km, such a condition occurs below 0.01 Hz. But if the ocean is shallow, constancy of U in the oceanic layer may persist up to a much higher frequency.

[7] Quantitative evaluation of this discrepancy is made in the following way; if the surface displacement of ocean waves is given by $\zeta = \int d\mathbf{k}' a(\mathbf{k}') \sin(\omega t - \mathbf{k}'\mathbf{x})$ where $\mathbf{k}'$ is its spatial wavenumber, the seismic displacement is given by

$$\begin{aligned}\mathbf{u}(t) = &-\frac{L_x L_y}{(2\pi)^2}\int_{-\infty}^{\infty} d\mathbf{k}\,\mathbf{R}(\mathbf{k})\int_{-\infty}^{t} d\tau \int_{-\infty}^{\infty} d\mathbf{k}' \\ &\times \frac{\sin\omega(t-\tau)}{\omega}\left\{\frac{\rho\omega'^2 U(0)}{2} + \Phi_Z(\mathbf{k}')\right\}\cos(2\omega'\tau) \\ &\times a(\mathbf{k}')a(-\mathbf{k}') \qquad (1)\end{aligned}$$

where a term related to horizontal forcing exists in the original derivation but is dropped here because it is not

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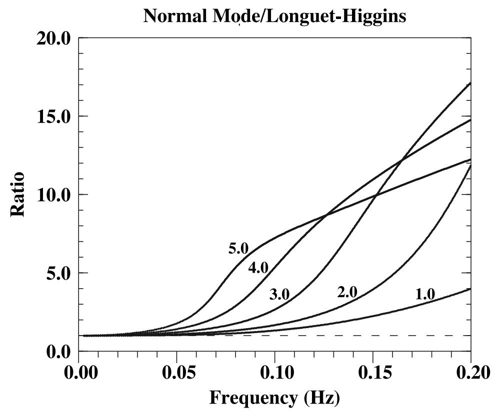

Figure 1. Ratio of excitation between the normal-mode approach and the Longuet-Higgins' approach. The latter applies the Longuet-Higgins pressure formula at ocean bottom. Discrepancies are significant if the source region is at a location with more than 1 km in ocean depth.

important in microseismic frequency range [*Tanimoto*, 2007]. In this formula, $L_x L_y$ is the source area, $\mathbf{R}(\mathbf{k})$ is a three-component Rayleigh wave eigenfunction which contains the vertical ($U(r)$) and horizontal ($V(r)$) eigenfunction. ρ is density of ocean, ω' and $\mathbf{k}'$ are the frequency and wavenumber of ocean waves and the integration over $\mathbf{k}'$ takes into account contributions from all wavelengths. The integral over τ exists because the source is continuous in time (thus the formula is in the form of convolution) and various wavenumbers (and thus frequencies) of ocean waves contribute through the integration over $\mathbf{k}'$. The last term $a(\mathbf{k}')$ $a(-\mathbf{k}')$ indicates that collision of ocean waves makes standing waves and dominant contributions.

[8] The excitation has two separate contributions as two terms in the braces in (1) indicate; one is the surface contribution, expressed by the term $\rho\omega'^2 U(0)/2$ where $U(0)$ is the surface value of the vertical eigenfunction. This arises from the kinematic boundary condition at ocean surface. The second term $\Phi_Z(\mathbf{k}')$ is an integral over the depth of the ocean from $-d$ (sea bottom) to 0:

$$\Phi_Z(\mathbf{k}') = \int_{-d}^{0} \rho U(z) \frac{k' \sinh\{2k'(z+d)\}}{\sinh^2(k'd)} dz \qquad (2)$$

The pressure change at ocean bottom is equivalent to replacing these two terms in the braces by $\rho\omega'^2 U(-d)$. They become equal if $U(r)$ is constant in the ocean [*Tanimoto*, 2007]. The degree of approximation with the use of the Longuet-Higgins pressure term is then measured by examining the ratio

$$\xi = \frac{\rho\omega'^2 U(0)/2 + \Phi_Z(\mathbf{k}')}{\rho\omega'^2 U(-d)} \qquad (3)$$

[9] Figure 1 shows this ratio (ξ) as a function of frequency. Five different cases of ocean depths, from 1 km to 5 km, are shown in Figure 1. They all converge at low frequency end, below about 0.05 Hz, as the thickness of ocean becomes a small fraction of the depth extent of eigenfunctions. The dominant frequency of microseisms is 0.15 Hz in Southern California (and higher up to about 0.3 Hz in other regions). Therefore, ocean depth in the source region may make significant differences between the two approaches. It shows that the deeper the ocean, the larger the discrepancies from the approach that uses the Longuet-Higgins pressure term at ocean bottom. At depths more than 3 km, the difference can reach a factor of ten.

3. Buoy Data and Fourier Spectra

[10] There is an increasing number of evidence that a source region of predominant (double-frequency) microseism is in shallow water. They include a seismic array study [*Friedrich et al.*, 1998], correlation study between ocean-wave amplitudes and seismic-wave amplitudes [*Bromirski et al.*, 1999, 2005], noise correlation study [*Stehly et al.*, 2006] and source direction study using Rayleigh wave characteristics in microseisms [*Tanimoto et al.*, 2006]. Source region of primary-frequency microseisms (at about 0.05–0.07 Hz) may not be so certain, however [e.g., *Stehly et al.*, 2006].

[11] We point out another piece of information that favors shallow ocean as a source of double-frequency microseisms. This evidence is from ocean-wave Fourier spectra (buoy data).

[12] First, we recall microseismic spectra in Southern California are remarkably similar among seismic stations [*Tanimoto et al.*, 2006]. Examples from three stations are shown in Figure 2 in which monthly averages of microseismic spectra are shown by different colors. They are Fourier spectral amplitudes of ground velocity records after removal of instrumental effects. Small peaks around 0.05–0.07 Hz are at the same frequency with ocean waves (swell) and large peaks between 0.1 and 0.2 Hz (the maximum at 0.15 Hz) are the double-frequency peaks. Despite the fact that some stations are more than 200 km away (Figure 3), similarity in spectra is surprising.

[13] The nonlinear normal-mode excitation theory implies that, when the double-frequency seismic peaks exist, there must also be double-frequency ocean waves because both are proportional to the term $a(\mathbf{k}')$ $a(-\mathbf{k}')$. Therefore, one of the necessary condition for a source region is to show the double-frequency ocean wave spectra.

[14] We have searched for such nonlinear effects in ocean-wave spectra from buoy data. Archived spectra data from the National Data Buoy Center (NDBC) and the Coastal Data Information Program (CDIP) were examined. Out of many buoy stations, six representative stations are shown in Figure 4. Instead of showing reported ocean-wave spectra directly, ocean wave spectra at each station were multiplied by angular frequency. This is because seismic data in Figure 2 are velocity spectra and they should be proportional to this multiplied product.

[15] The top two panels, A and B, show spectral characteristics of the outer Pacific ocean. A is located west of San Francisco, in a deep ocean at depth 4559.4 m. Locations of stations B-F are shown in Figure 3. B is at a relatively shallow ocean (384.1 m) but because of its location, the spectral characteristics is similar to those in the outer ocean. The most important characteristics is that both A and B show predominant single peaks without much hint of secondary, double-frequency peaks, especially in winter.

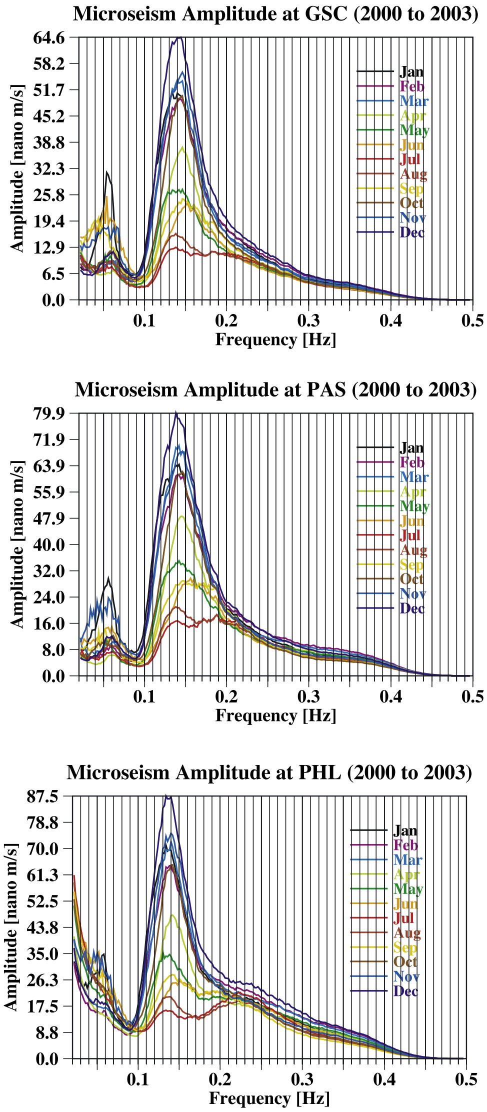

Figure 2. Microseism spectra at three separate location in Southern California. They are remarkably similar despite the fact that distance between PHL and PAS is more than 200 km. At the predominant frequency of ocean waves, about 0.05–0.07 Hz, only small peaks are seen. The predominant microseisms are at 0.15 Hz due to nonlinear frequency-doubling effects.

[16] On the other hand, spectra at C, D, E, and F clearly show effects of frequency doubling effects by the nonlinear mechanism. They are all close to the coast. In cases of D and E, secondary peaks are probably too broad to consider them to be at twice the frequency; they may contain successive nonlinear effects to make them broad as to reach 0.3 Hz. Since seismic spectra show rather sharp peaks between 0.1 and 0.2 Hz, these locations are probably not the source region but they do show clear nonlinear effects. In general, it is very hard to make precise estimate as to where the most likely sources of seismic excitation are due to variations of ocean wave spectra. In fact, there is a good chance that the source region is spatially extended, perhaps like a line source. However, spectra from outer-ocean buoys are almost entirely devoid of double-frequency peaks, suggesting that the source regions must be close to the Southern California coast. It is important to note that as long as a buoy is in the Southern California Bight, approximately the shallow water area in Figure 3, nonlinear behavior of ocean waves seem to exist, regardless of ocean depths. Coastal reflection and complex seafloor topography are probably helping to create standing ocean waves in this region.

[17] The evidence of these shallow sources implies that the discrepancies between the Longuet-Higgins approach and the normal-mode approach for microseism excitation may not be so important in practice. We do make a note, however, that the location C is at an ocean depth of 1856 m, although it is relatively close to the coast. Excitation at such a place requires the use of normal-mode formula in order to make correct estimates.

4. Discussion

[18] *Bromirski et al.* [2005] pointed out that there are two main frequency bands for the double-frequency microseisms, one peaks at about 0.15 Hz and the other peaks above 0.2 Hz. They termed the former LPDF (Long Period Double Frequency) and the latter SPDF (Short Period Double Frequency). Through the correlation of wind data and seismic signals at station H2O, they demonstrated that SPDF is related to winds (and wind-generated ocean waves) and similar waves can also be seen at stations in California. Because of their high frequency range, SPDF do not propagate long distance. Therefore, the observed SPDF in California are associated with winds and ocean wave activities near the coast.

[19] This paper and *Tanimoto* [2007] deal with LPDF, the predominant double-frequency microseisms that are generated by ocean swells with frequencies 0.05–0.07 Hz. In order to understand SPDF, one needs to monitor local wind sources and the generated ocean waves by them, although we believe the same theory should quantitatively explain seismic energy.

[20] One may ask why the spectra in Figure 2 do not show SPDF. We believe there are actually SPDF signals in one of the spectra, the one at PHL. Because the data are so dominated by the peaks at 0.14–0.15 Hz, it is hard to see but the spectra at PHL contain a secondary peak at about

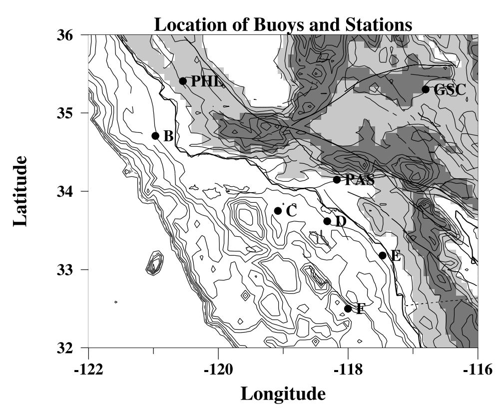

Figure 3. Location of buoys and three seismic stations in Figure 2.

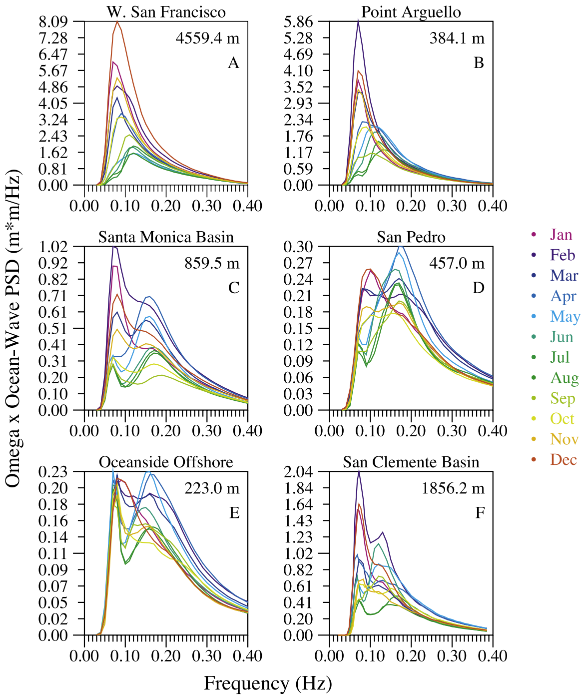

Figure 4. Ocean wave Fourier spectra at six locations. A and B show typical spectra in the open ocean and are devoid of frequency doubling effects that are required for microseism excitation. Stations C-F all show some nonlinear effects due to generation of standing ocean waves. Interaction between coastal reflection and incoming ocean waves probably create these nonlinear effects. Because of these nonlinear effects in ocean waves, source region of double-frequency microseisms must be somewhere near the coast, generally in shallow water.

0.23–0.25 Hz. The ratio of amplitudes between LPDF and SPDF is about 3–4 in this case and this is about the same ratio that can be read from spectra of *Bromirski et al.* [2005].

[21] The reason the secondary peaks do not exist in spectra for GSC and PAS is probably because of distance from the coast. PHL is very close to the coast (about 20 km) while GSC and PAS are much further from the coast. Considering the attenuation effects for higher frequency waves, the lack of SPDF at GSC and PAS is not surprising.

5. Conclusion

[22] A conventional view on the generation of microseisms is through the nonlinear mechanism, pointed out by *Longuet-Higgins* [1950], which generates pressure variation at ocean bottom. Treating the source only at the ocean bottom by this pressure effect is an approximation, however, since Rayleigh waves, the dominant wave type in microseisms, have displacement throughout the ocean and their amplitudes vary with depth. Comparison with a more rigorous normal-mode formula showed that this conventional approach fails in deep oceans. Therefore, under certain conditions, the approach by the use of the Longuet-Higgins' pressure formula at ocean bottom may be misleading. If the source regions are in shallow water, it is justified.

[23] New evidence of nonlinearity in ocean waves from buoy data was presented to support the view that the source regions for microseism excitation is in shallow water. Therefore, the conventional use of the Longuet-Higgins pressure for the excitation of microseisms seems to be justified after all, although one should pay attention to ocean depths in the source region very carefully.

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