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Title: OCEAN WAVE-GENERATED MICROSEISMS AT THE OREGON COAST

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Sea wave activity off the coast at Newport, Oregon, generates small, regular ground oscillations, called microseisms. These are of magnitudes in the order of microns (10^{-4} cm), and their periods are consistently close to half those of the local sea waves. The 2:1-frequency ratio indicates that the microseisms are caused by second-order pressure variations exerted by standing water waves on the sea bed, as is proposed in a theory given by Longuet-Higgins (1950). Such pressure variations do not decay with depth in water, and their amplitudes are proportional to the product squared, of the sea-wave heights and frequencies.

Standing waves require the presence of trains of opposite-traveling waves, and, typically in coastal regions, they arise from the interaction of incoming waves with waves reflected seaward by the shore.

Analysis of three sets of simultaneously obtained sea-wave and microseisms records from the Oregon Coast (near Newport) shows microseisms of rms amplitudes from about 2 to 11 microns to be associated with rms sea-wave heights from about 0.4 to 4.5 meters. Simple numerical calculations in accord with Longuet-Higgins' (1950) theory show that such microseisms may be quantitatively accounted for in terms of generation by a hypothetical system of standing waves, created along the entire Oregon coast by reflection at the shore of incoming waves of the same heights and periods as those observed. Assuming small reflection coefficients in the order of 0.01 to 0.1, enough standing wave activity should result in a narrow (400 meters wide) region along the coast to produce ground oscillations of the amplitudes observed at the measuring site.

There exists a roughly linear relationship between the rms amplitudes of the microseisms and the squared product of the local sea-wave heights and frequencies. Microseisms measured at the Oregon coast may thus be conveniently used in place of direct sea-wave measuring methods for determining the significant heights and periods of the waves off the shore.

Ocean Wave-Generated Microseisms at the Oregon Coast

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1.0 INTRODUCTION

1.1 General Definition of Microseisms

Microseisms are minute, continuous, more or less regular oscillations of the ground, which may be detected with sensitive instruments. Magnitudes are typically in the order of a few to a couple of microns (10^{-6} meters), and their periods, though ranging from as low as a fraction of a second to several minutes, tend to predominate in the range from about 2 to 10 seconds. Such ground noise is present more or less everywhere on the earth's surface. It is not caused by earthquakes or explosions, but arises largely from continuous forces in the atmosphere and oceans, including pressure variations in the atmosphere and in water due to ocean wave activity.

The most commonly observed types of microseisms (about 4 to 10 seconds) tend to be greater in winter than summer, are typically associated with storms and cyclones over the oceans, high sea states or the passage of cold fronts across continental margins.

A good deal of seismic noise is also characteristically present in volcanically active and geothermal regions, which is due largely to thermal (convective) processes below the earth's surface. This noise occurs both in the form of frequent, small-magnitude earthquakes (microearthquakes) as well as continuous oscillations or trembling (microseisms). In addition, high-frequency ground vibrations (cultural noise) may usually be detected near cities or other communities due to traffic, industrial activities, etc., but, being due to artificial causes, these are not considered by many as microseisms.

Microseisms consist mostly of a mixture of the two types of seismic surface (Rayleigh and Love) waves, and possibly channeled waves similar to the so-called Lg and Rg modes. Body (P and S) seismic waves constitute in general only a minor portion of microseisms. Surface seismic waves, which, in contrast to body waves, are characterized by propagation confined to and along the surface and other boundaries within the earth, attenuate less rapidly with distance of travel than do body waves, and they tend to form the more conspicuous portions of seismograms and of seismic noise in general.

Microseisms contribute to the background noise in various seismic and non-seismic field measurements, and are usually a nuisance in geophysical work.

1.2 Description and Classification

The character of microseisms varies greatly depending on the source, time and place of observation. High-frequency and usually irregular ground noise with variable amplitudes and periods includes volcanic tremors, cultural noise and noise due to air turbulence, wind, rain, water flow, and sometimes shore surf and short-period chop. Fairly regular oscillations of periods from about 4 to 10 seconds are typically associated with the action of ocean waves. These latter are among the most common and conspicuous types of microseisms, occurring worldwide and tend to be particularly prominent in coastal regions. Longer-period microseisms, also commonly observed, result from low-frequency swell, surf and sometimes also natural oscillations of polar ice covers. Microseisms with very long periods (i. e. minutes) have been attributed to effects such as temperature changes (freezing) of the ground.

Microseisms of periods several seconds or more are observed to be typically able to propagate over considerable distances across the continents. They cross hundreds and thousands of miles with very little loss in energy, as long as there are no major discontinuities in their paths, such as mountain ranges or continental margins. Sudden, temporary increases in microseism activity, often called "microseismic storms," are also usually detected simultaneously over widely separated regions of the continents. The high microseismic activity, which usually lasts for a few days, may in general be associated with a storm or cyclone over the ocean, and the intensity decreases again when the storm moves into a continental region.

On the other hand, microseisms seem to attenuate rapidly in oceanic path, being essentially lost beyond detection over distances exceeding a few hundred miles. In general, the shorter periods are lost more rapidly than the longer periods, and the predominant periods of microseisms tend to increase with the distance of propagation. High-frequency ground noise due, for example, to community activities, traffic, etc., is detected only within a few miles or so of a city.

Microseisms in the period range from 2 to 10 seconds have, for convenience, been classified by Iver (1963) as "storm microseisms," as opposed to "short-period" and "long-period" microseisms outside of this range. Storm microseisms, which incidentally includes those observed at the Oregon Coast, are those which have received the most attention, have been the most studied, and many discussions in the literature implicitly refer to seismic noise in that category.

1.3 Early and Current Theories on Microseisms

Along with an extensive observational work which has been carried out on microseisms and their varyingly consistent relations to meteorological phenomena, sea states and other conditions, there were efforts made to explain how they were generated. While there was generaly agreement that the energy source of most microseisms was largely the atmosphere, the physical mechanism whereby the various possible natural forces (pressures from sea-waves, atmospheric turbulence, etc.) excited oscillations of the earth's surface was, until relatively recently (about 1950), not at all well understood. Several theories were advanced, favoring either atmospheric pressures or sea waves as primary forcing functions, but most failed to account for all the observations, or were inadequate from theoretical points of view.

One of the earliest suggestions given as to the cause of microseisms was that they resulted from the impact of ocean waves breaking against steep shores (Wiechert, 1907). This theory was for a long time supported by many authorities (including Gutenberg, Bath and others). Significant correlations had been noticed by Wiechert (1904), Gutenberg (1912) and others between sea states and microseism intensities in Europe. There seemed moreover to be no problem in accounting for their energy, since apparently only a small fraction (10^{-7}, 10^{-6} or 10^{-3}, according to Byerly (1942), Bath (1949) and Gutenberg (1951), respectively) of the overall amount of energy dissipated in shore surf was required to be transmitted to the ground to give ground oscillations of the magnitudes observed. This hypothesis was never firmly established, owing to the lack of data on surf heights. Neither was it entirely abandoned, however, and it is presently considered only as a minor source of some microseisms at best.

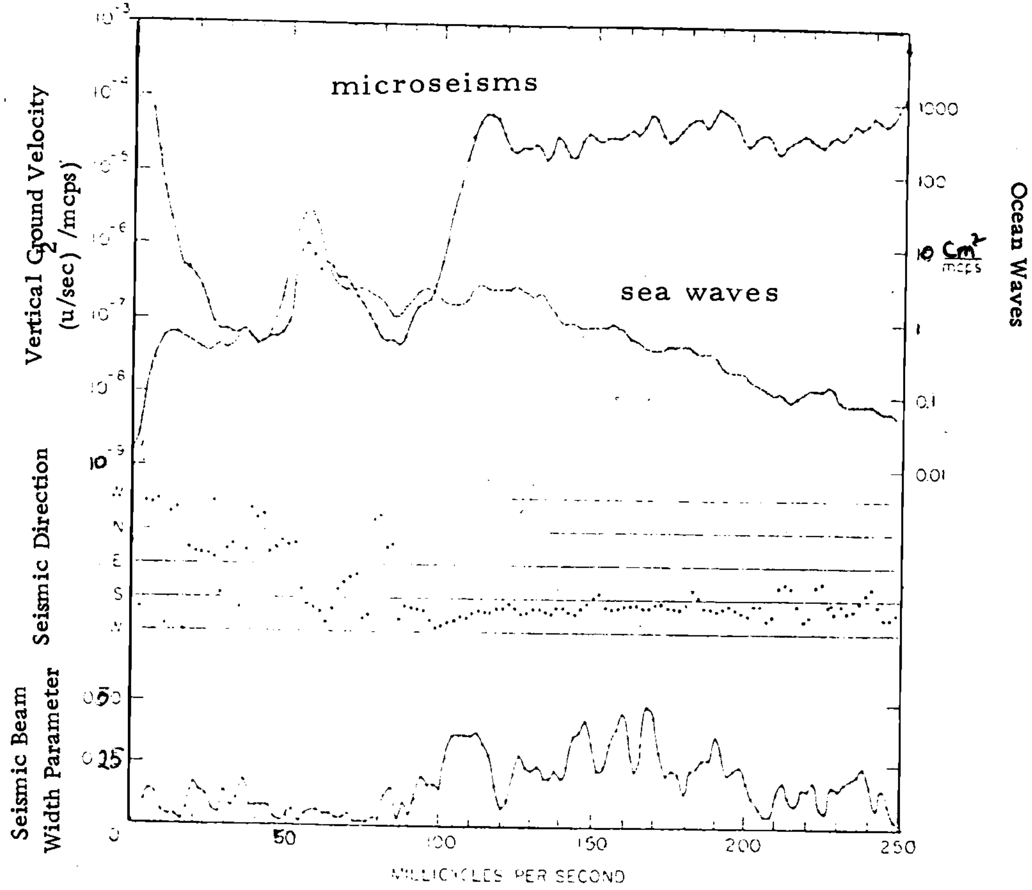
Banerji (1935) attempted to account for them in terms of pressures on the deep-ocean bottom due to gravity waves at the ocean surface, but this mechanism was criticized as inadequate by Gutenberg (1931) and Whipple and Lee (1935) on account of the rapid exponential decline of the (first-order) pressure variations with depth in water. Later, Scholte (1943) tried to revive the theory by considering the compressibility of water. He developed an expression (equation 4.0-14) for the propagation of (vertical) oscillations along the ocean bottom due to excitation by a periodic point force applied at the sea surface.

Others, including Gherzi (1924) and Bradford (1936), thought that pressure variations in the atmosphere caused microseisms by "barometric pumping" or "kneading" of the surface while a pressure change moved horizontally over it.

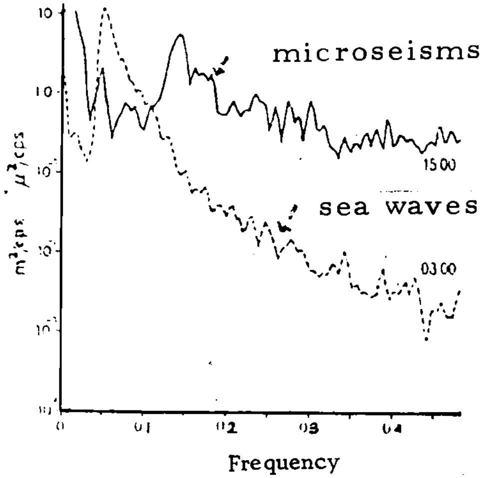
It had soon become apparent, as more field observations were made, that wave breaking at sea shores could not be the only source of microseisms. Certain observations frequently made were difficult to explain in terms of Wiechert's surf theory. For example, that microseism intensities often appeared to increase hours before the arrival of high sea-waves at shores, which were supposed to generate them. Frequency-correlations furthermore often tended to indicate that the periods of the microseisms were considerably shorter, or more nearly about one-half the values of the predominant sea-wave periods (Bernard, 1941; Deacon, 1947; Darbyshire, 1950). It was suggested by Bernard (1941) that some kind of sea wave interaction (possibly "standing waves") might be involved in the generation of microseisms, but he gave apparently no explanation for the observed 2:1-frequency ratio between the sea waves and the microseisms. The currently accepted explanation for most seismic noise of the storm microseism type was developed shortly after, following theoretical work by Miche (1944), by Longuet-Higgins and Ursell (1948) and Longuet-Higgins (1950). It is a well-known theory now, which proposes that microseisms are caused by second-order pressure variations in water, associated with standing waves on the ocean surface. These are non-linear pressure variations produced by the interactions of opposing waves of equal periods. Their periods are half those of the waves themselves. Their amplitudes are proportional to the squared product of their heights and frequencies and they do not decay with depth in water.

The second-order pressure variation appears in a previously overlooked, second-order term in the solution for the wave velocity potential for the case of a standing wave (Miche, 1944). Longuet-Higgins (1950) also derived the second-order pressures directly from the Bernoulli equation, and he showed that it would be of sufficient magnitude to cause microseisms of the amplitudes observed. The standing wave condition requires the interaction of waves traveling in opposite directions and having the same periods and lengths. Such interactions are likely to be found under many ordinary conditions at sea, including, for example, cyclonic depressions over oceans and near continents, where waves approaching the shores interact with waves reflected from the coasts. This theory is reviewed in section 3.0.

Standing ocean-wave activity has been supported repeatedly as a cause of microseisms by observations, including those of Bernard (1941), Deacon (1947), Darbyshire (1950), Dinger (1954), Dinger and Fisher (1955), Haubrich et al. (1963), Darbyshire and Okeke (1968) and many others. Examples of comparative sea-wave and microseism spectra from two different coastal stations are shown in Figure 1.3-1. In each of these cases, a half-period microseismic energy peak is seen to correspond to the predominant periods of the sea waves. Coastal wave reflection is probably a condition responsible for most or all of the microseismic activity observed in coastal regions in general. It is not known what fraction of the incoming wave energy is actually reflected by most sea-shores, since reflection would be difficult to observe directly, but reflection coefficients are usually taken to be in the order of a few percent for most average beaches. Reflection depends on shore slope and topography and on the lengths and heights of the sea waves. The formation of standing-wave systems, and hence



From Haubrich et al , La Jolla, Calif. (1963)



From Darbyshire and Okeke, at Anglesey, Ireland (1963)

Figure 1.3-1. Spectra of simultaneously measured sea-wave and microseisms by previous workers in other coastal regions microseismic activity, should also depend on the direction of approach of the incoming waves to the shore.

Microseisms are also produced on the bottom of deep oceans (Bradner et al., 1965; Prentiss and Ewing, 1963), although they are unlikely to be detected far from the sources, because of the relatively rapid attenuation in the oceanic crust. The generation process there (by standing waves) would, moreover, by subject to resonance effects (see section 3.0), because of the great water depths. Resonance involves several-fold amplifications of the non-linear pressures at certain, discrete depths, depending on the excitation frequencies. Longuet-Higgins' theory therefore suggests that the source areas of most of the microseismic energy are the open oceans.

It is not certain, however, how far compression-waves generated by the non-linear wave interactions at the ocean surface may travel in the oceans themselves before exciting seismic oscillations on the sea bed, including continental shelves and shelf slopes, where microseisms may enter the continental crust.

In addition to generation by standing waves on the ocean surface, as proposed by Longuet-Higgins (1950), and wave breaking on beaches, as was first suggested by Wiechert (1907), microseisms may sometimes also be produced by progressing wave systems. According to Hasselmann (1963), ocean swell moving through shoaling water on a sloping beach may cause small, local ground oscillations having the same periods as the predominant sea waves. Such "primary frequency" oscillations are usually much smaller, and they are often observed in coastal areas along with the typical "double frequency" microseisms generated by standing waves off the shore (e.g. Haubrich et al., 1963; Darbyshire and Okeke, 1968; Hinde and Hatley, 1965; Bossolasco et al., 1973, and many others). The increase in the heights of each of a train of incoming waves, which results from the reduction in group velocity associated with the shoaling, generates, in effect, an oscillating pressure field along the shore. This pressure field has the same frequency as the sea waves, and a magnitude depending on the shore slope and heights, periods and direction approach of the waves to the shore. The resulting microseisms, when detected, are nearly always much smaller than those generated by standing waves, and their periods are usually relatively long (tens of seconds). According to Hasselmann (1963), the process is effective only for fairly low-frequency swell.

Other theories of microseisms may be mentioned. Press and Ewing (1948) attempted to explain the typically observed storm microseisms in terms of the character of seismic surface wave transmission in oceanic regions. The oceanic regions were regarded as two-layered liquid-solid, acoustic systems, with the sea surface representing an antinode, receiving the input a broad spectrum of arbitrary generating forces. These were assumed to be sea-wave or atmospheric pressure variations associated with storms or cyclones over distant areas of the oceans, far from the points of observation. They developed dispersion relations (seismic wave velocities as functions of frequency and ocean depth; Figure 1.3-2) for the first two normal (Rayleigh wave) modes propagating along the liquid-solid interface on an idealized, semi-infinite, two-layered system composed of a layer of water over a uniform, solid half-space. The physical properties of the solid were chosen as equivalent to those of the average oceanic crust, so that the system resembled an actual oceanic region. Based on theoretical group velocity curves (group velocities as functions of non-dimensional frequency, FH/β; see Figure 1.3-2) for idealized ocean-earth systems, actual characteristic depths (H) of about 4 km of the oceans, and shear wave velocities (β) of about 2.5 km/sec for the oceanic crust, they gave an explanation for the relative predominance seismic noise periods within the typical range of 2 to 10 seconds.

It has for a long time been a known fact in seismology that those wave components in a continuous seismic wave spectrum which travel with the lowest group velocities tend to show up in seismic recordings with relatively large amplitudes (Airy phases), particularly when the observation is made far from the source of the waves. Being the slowest traveling of all the waves in the spectrum, they arrive relatively late, and their predominance is due to additive interaction, which is greatest between wave components represented by points in

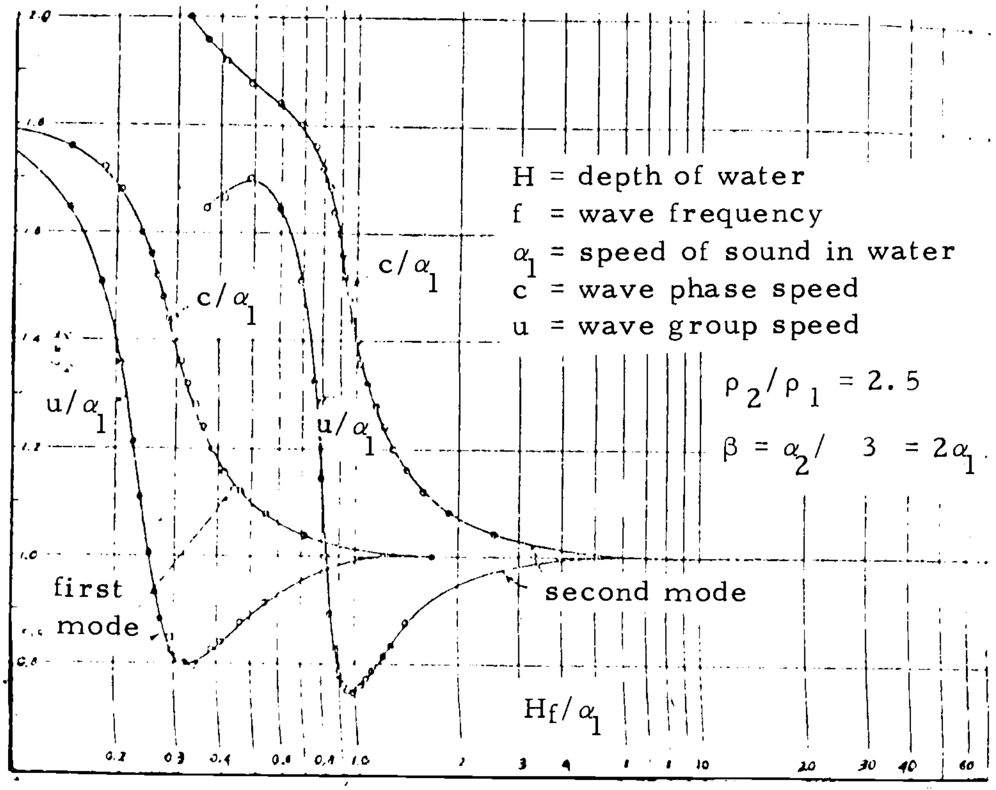


Figure 1.3-2. Phase and group velocity curves for normal mode propagation in a system consisting of a liquid layer overlying an infinitely thick solid. (From Press and Ewing, 1948)

the neighborhood of those of the minimum group velocities, of a dispersion curve (e.g. Figure 1.3-2). These wave components are of nearly the same wavelength and travel for long distances without separation. Their amplitudes decrease as the inverse one-third power of the distance traveled, compared with the inverse square-root decrease which characterizes seismic surface waves in general.

Rayleigh waves, which are ideally non-dispersive in uniform, homogeneous, half-space media, become dispersive if the medium is layered, and the layer thicknesses are in the order of some small, but significant (i.e. tens of percent) fraction of the wavelength. Typical thicknesses of the earth's oceanic and continental crust are illustrated by the cross section shown in Figure 1.3-3 of the North American continental margin. They are of orders causing significant dispersion of Rayleigh waves in the microseisms frequency range.

The "normal modes" for the below system are simply those coupled sets of waves in the liquid and the solid, which are associated with undamped energy transmission along the horizontal boundaries without loss to the adjacent media. They consist of those compressions in the liquid which undergo total reflection at the upper and lower boundaries and additive interaction with each other and with the (Rayleigh) seismic surface waves at the liquid-solid interface. Relations between velocities and wave numbers or frequencies are obtained for each normal mode by solving sets of simultaneous linear algebraic

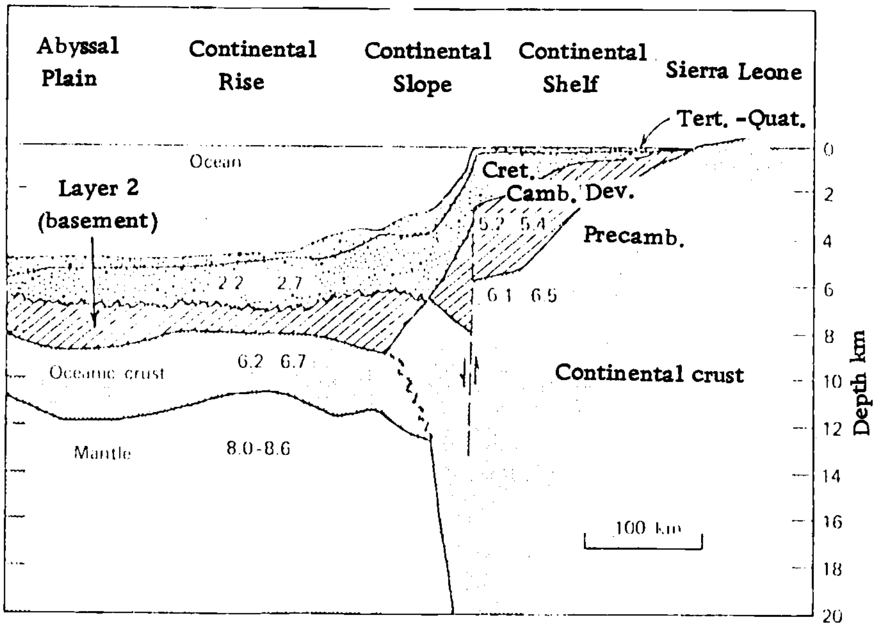


Figure 1. 3-3. Seismic cross sections through continental margins and transitional zones, showing P-wave velocities.

equations, which arise with the applications of seismic wave solutions to be presented in the discussion of seismic waves in the following section. The (dispersion) relations between the wave velocities and frequencies are continuous, because of the horizontally infinite dimensions of the transmitting media.

Hasselmann (1963) analyzed theoretically wave motions induced in generalized, many-layered half-spaces by artibrary, homogenous, stationary force fields applied at the free surfaces of the systems. He developed relations between the statistical properties of the forcing functions and the resulting wave motions in the half-space systems. He computed transfer functions and dispersion curves (Figure 1.3-4)

$$\tilde{T}_{s_v}^{R} = \omega\cdot\chi_{s_v}(\alpha_2/\beta_2)\rho_2^{-2}\cdot\beta_2^{-5}$$
$$\tilde{T}_{s_h}^{R} = \omega\cdot\chi_{s_h}(\alpha_2/\beta_2)\rho_2^{-2}\cdot\beta_2^{-6}$$
$$\tilde{T}_{e}^{R} = \omega^2\cdot\chi_{e}(\alpha_2/\beta_2)\rho_2^{-1}\cdot\beta_2^{-4}$$

$$\chi_{s_v}(\sqrt{3}) = 0.460$$
$$\chi_{s_h}(\sqrt{3}) = 0.214$$
$$\chi_{e}(\sqrt{3}) = 1.36$$

c = phase velocity; v = group velocity

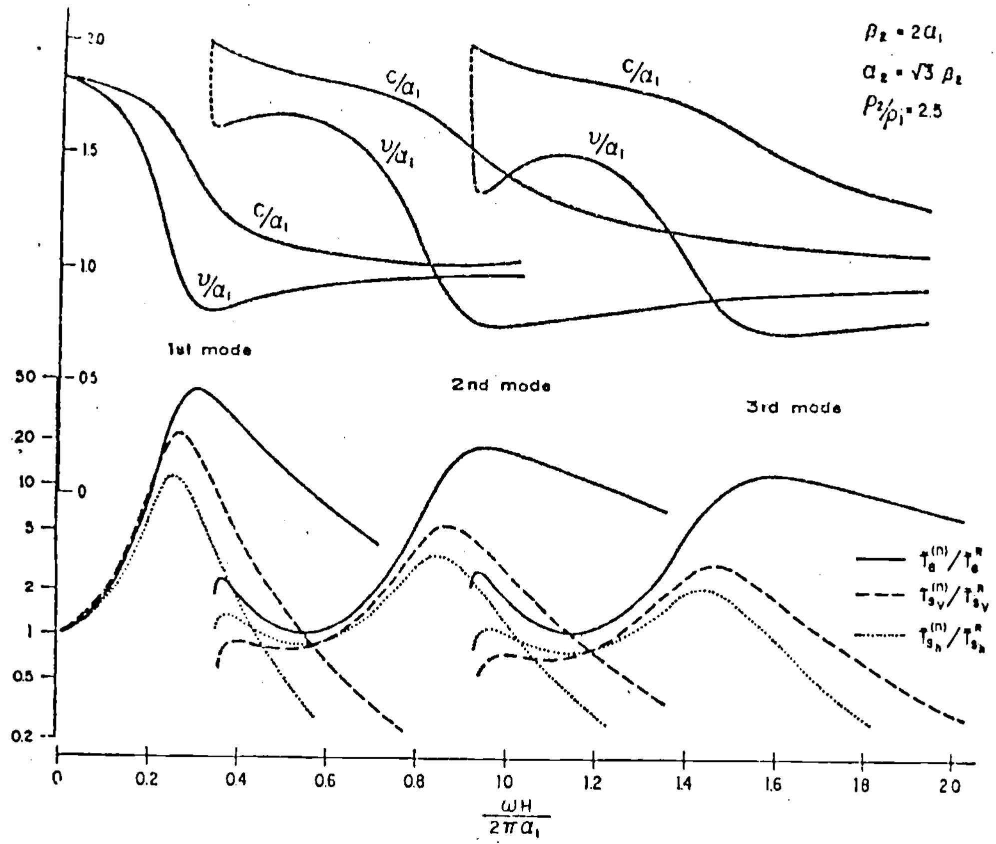


Figure 1.3-4. Top: Dispersion curves for the first three modes of a fluid layer over an elastic half-space. Bottom: Ratios of the transfer functions $T^{(n)}$ for the two-layer system to the transfer functions $T^{(R)}$ for the Rayleigh wave of the elastic medium alone. The layer parameters correspond approximately to the media water and granite (from Hasselmann, 1963).

for waves (the first three normal modes) theoretically propagating waves along the surface and boundaries of an ideal, two-layered half-space, composed of a layer of water overlying a uniform granite half-space. He discussed three different types of microseism generating mechanisms. One involves generalized, non-linear sea-wave interactions of essentially the nature discussed by Longuet-Higgins (1950), another is the above-mentioned amplitude modulation of progressing waves on sloping beaches, and the third (considered least important) is excitation by pressure fluctuations and turbulence in the atmosphere.

Difficulties remaining with the above theories, and, incidentally also with Longuet-Higgins' theory regarding origins in the open oceans of microseisms observed on the continents, arises from the apparent rapid attenuation of short-period seismic surface waves and microseisms in oceanic paths. Apparently there is a virtual absence in oceanic records of all first-mode Rayleigh waves in the period range from 1 to 12 seconds, although these waves are quite abundant on the continents. The theory of Press and Ewing requires considerable distance of travel of microseisms for the development of transmission peaks in the spectra. Seismic noise generated by sea-wave activity in distant areas of the oceans is therefore unlikely to be detected at any land stations, and it may be concluded that the spectral distributions of most microseisms are in general a reflection of spectral energies of more or less local generating forces (ocean waves). Distances between areas of generation and observation of microseisms may be considerable if the travel paths are within the continental crust. The continental crust extends from the interior of the continents to the shelf edges and slopes.

Possible excitation of the earth's surface by atmospheric forces directly has been discussed by many (including Sezawa and Kanai, 1933; Haskell, 1951; Ewing et al., 1957) and others (Gutenberg, 1958). On the whole, this role of the atmosphere is not as well established as that of ocean waves, and it is considered by some (including Hasselmann [1963]) as relatively unimportant.

1.4 Previous Work on Microseisms

Research on microseisms has continued since the beginning of the nineteenth century, when it was first noticed in the noise of various non-seismic (e.g. gravity) measurements. One of the earliest observations probably ever made of microseisms was in 1870 when Bertelli noticed during some years continuous motions of a pendulum he installed, which lasted for hours and days and of which he observed a correlation with disturbed air pressures. Seismic noise received special attention as better and more refined instruments came into use. Microseisms were identified as the different types of seismic waves (P, S, Rayleigh, Love waves), and their characteristics (amplitudes, periods, velocities, energy spectra, etc.) were studied in relation to the various meteorological and other conditions. Following a large amount of essentially empirical work, contributions were given within the last couple of decades from theoretical points of view, through which microseisms, and in particular their mechanism of origin, became fairly well understood.

Possibilities of predicting weather conditions by tracking and locating storm centers over the oceans with microseisms prompted a large amount of data collection during World War II. Microseisms observations had occasionally shown promise as additional tools in weather forecasting, but the overall success of such methods turned out to be limited and were eventually abandoned.

At the same time interest in microseisms continued to grow, both from a purely theoretical point of view, as well as other possibilities of practical application. For example, microseisms might be used in conjunction with information about storm location for inferring something about the structure and properties of the earth's crust under different parts of the oceans (Press and Ewing, 1948). There have been a number of studies (e.g. Oliver et al., 1955; Oliver and Gorman, 1961) carried out on the relations between oceanic crustal structure and dispersion of short-period surface seismic waves in oceanic paths. Microseisms might possibly also be used in locating geothermal areas (Clacy, 1968).

Several investigations on the relations between microseisms and sea-wave activity and atmospheric conditions led to the theories given by Longuet-Higgins (1950), Hasselmann (1963) and others, which were followed by numerous observations reported as either supporting or contradicting them.

Literature on microseisms is quite extensive, as the subject has remained a challenge attracting workers from various diverse fields, including geophysics, oceanography and meteorology. There are, however, a few general review articles to which the interested reader may be referred, and they include one by Gutenberg (1958), one by Iyer (1963) and a couple of shorter ones which appear in the reference list. A 600-reference bibliography was prepared by Gutenberg and Andrews (1952).

Recently, microseisms observed at the Oregon coast have been used for determining significant heights and periods of ocean waves off the shore. There is a good correlation between the heights of the sea waves and the amplitudes of the microseisms, and their periods are always close to half the predominant periods of the sea waves. Measurements were made by Zopf (1972), assisted by Clayton Creech of the Marine Science Center at Newport, Oregon, simultaneously of ground noise at Newport and sea wave heights at an offshore site nearby. These data were gathered in order to ascertain and evaluate the possibility of using microseisms for measuring the characteristics (heights and periods) of the offshore sea waves. Microseisms were used with apparent success for sea-state prediction verification by Enfield (1974) in lieu of the traditional methods of direct measurements with sea-wave sensors or visual observations. Visual observation of waves depended on visibility and were usually too infrequent and sporadic. In addition, problems were experienced with malfunctioning of a wave sensor, which had been installed near Newport, Oregon, in the fall of 1971. However, a brief period of correct functioning was obtained in the early fall of 1972.

The seismometer, on the other hand, performed continuously and reliably, providing continuous strip-chart recordings of the ground motion. The instrument is a commercial type (Teledyne-Geotech, model SL210) seismometer, designed for geophysical surveys. It was installed in the Marine Science Center building in May, 1971, on a concrete floor. It was felt that seismic methods for routine sea-wave measurements would be more convenient and feasible, less expensive, and thus preferred whenever essentially qualitative information on the sea state was desired.

Preliminary investigations of the frequency relations between the local sea waves and microseisms by Zopf (1972) showed that the microseisms at the Oregon coast were undoubtedly generated by standing sea waves off the shore. Zopf (1972) infers that the standing-wave activity off the Oregon coast arises from the interaction between incoming waves and their reflections at the shore. The seismic signal generated by the sea waves is very prominent in the seismic records, compared with all other frequencies, and the peaks of these signals occur at frequencies consistently close to twice those of the peaks in the corresponding sea wave records. For this reason the microseismic records were found useful for estimating both the predominant periods and approximate significant heights of sea waves off the shore. (Zopf (1972) suggested that the 50-foot sand-filled surface layer of the area might damp out high frequency noise.)

Based on measurements and on the assumption of a linear relation between the amplitudes of the microseisms and the apparent force per unit area, as inferred on basis of local measurements to be exerted by the standing waves on the sea bed, at the measuring site, Zopf (1972) derived the following empirical relation between the significant sea wave heights $H_{1/3}$, in feet), and the rms amplitude ($\overline{a}_w$) and period (T_s, in seconds) of the seismic oscillations:

$$\delta_{\%} = 32 \ H^2_{1/3} \ / \ T^3_s \qquad (2.4\text{-}1)$$

where $\delta_{\%}$ represents the seismometer deflection, given in percent of the full-scale on the chart. Equation 1.4-1 was obtained by a regression analysis of the above parameters ($H_{1/3}$, $\overline{\delta}_{\%}$ and T_s), which were determined from each of a series of one year's records. The procedures are described in detail by Zopf (1972).

Significant sea-wave heights, obtained using equation 1.4-1, were plotted against visually observed and wave sensor determined heights by Zopf (1972) and are shown in Figure 1.4-1. The points are seen to fall essentially on a 1:1-line, with some scatter. Although Figure 1.4-1 (b) was based on only one week's pressure-sensor data, equation 1.4-1 was considered (by Zopf, 1972; and Enfield, 1974) to provide a good estimate of the sea-wave heights and periods in terms of the microseismic measurements, at least for the purpose of wave prediction verification.

1.5 Purpose and Scope of This Thesis

The purpose of the present work is to demonstrate the causal relationship between the microseisms and the local sea waves at Newport, Oregon, implied by the theory of Longuet-Higgins (1950). It will be shown that adequate standing-wave energy for the generation of observed microseismic activity is likely to form near the shore from the interaction of incoming with coastally reflected waves. A set of simultaneously measured sea waves from pressure-transducer recordings and microseisms, obtained at Newport, Oregon, are spectrally analyzed and presented. The observed microseismic activity is quantitatively accounted for in terms of the sea-wave observations, the probable amount of wave reflection and the consequent standing-wave distribution along the shore. Conditions for the presence of

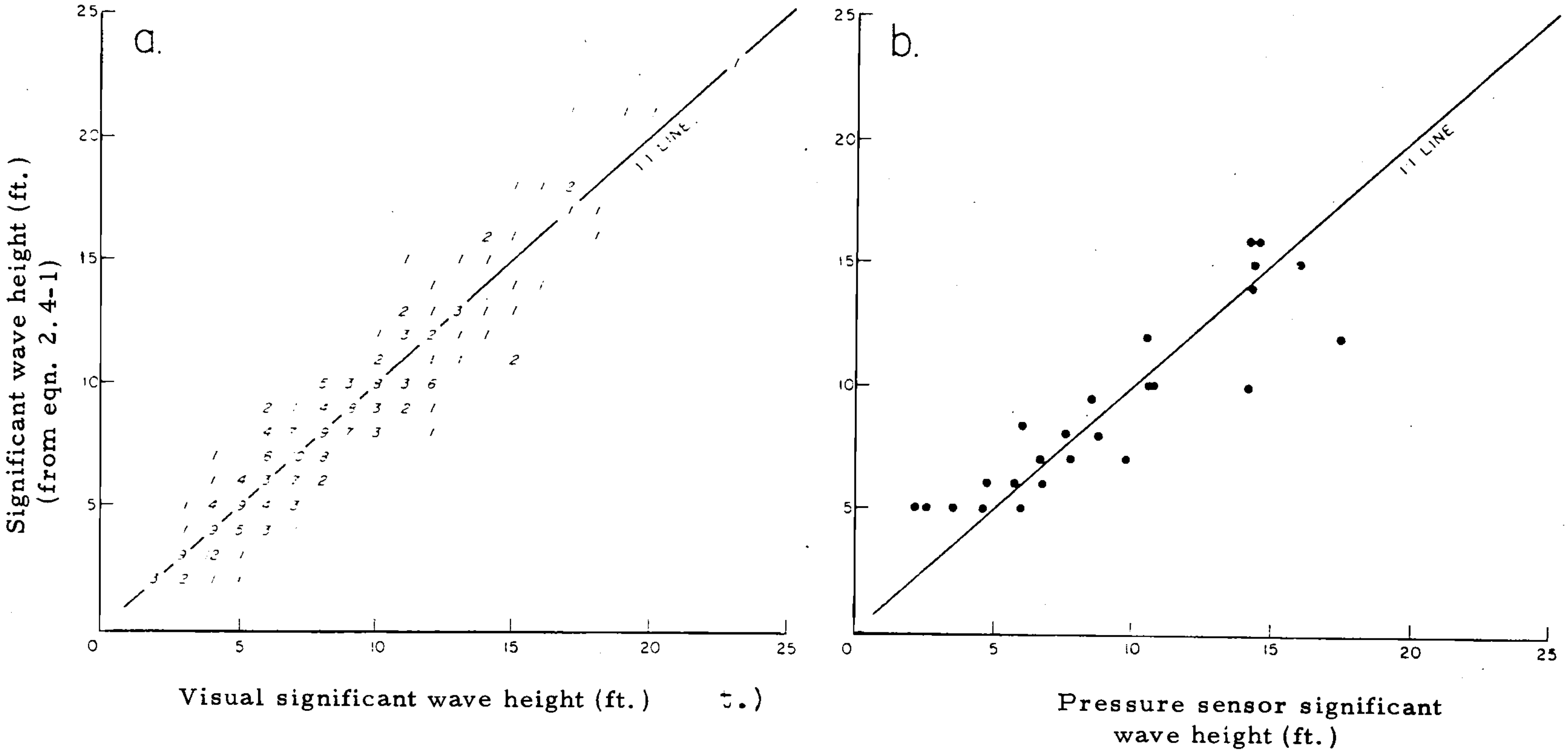


Figure 1.4-1 Scatter diagrams of seismometer-inferred heights vs. visually observed heights (a) and pressure-sensor (significant) heights (b) (from Zopf, 1972).

standing-wave systems, and consequent microseism activity, will be discussed, particularly in regard to the utility of microseisms for measuring sea waves. The nature and generation of microseisms by ocean waves, as best understood at the present time, will be reviewed. Ground motion at the measuring site, due to other effects or generating mechanisms, will be evaluated.

2.0 SEISMIC WAVES

A brief review is given here of the general nature of seismic waves, particularly the seismic surface waves, as an additional illustration of the physical nature of microseisms in general. Any textbook in seismology (see Bullen, 1968) may be consulted for a more thorough presentation on the subject.

The two basic types of seismic waves encountered in the field are the bodily waves and the surface waves. Body waves travel inside the medium (the earth) and they are of two kinds; namely the P (primary) waves, which are compression waves, and the S (secondary) waves, which are shear or rotational waves. Particle motion associated with P-waves is parallel to the direction of propagation, while for S-waves it is at right angles to it.

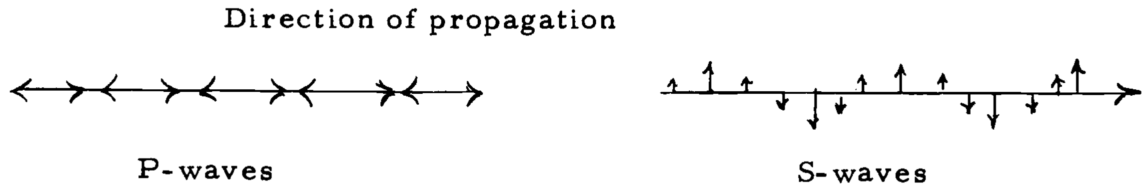


Figure 2.0-1. The two types of body wave motions.

P-waves are the fastest traveling of all seismic waves, and are thus the first to register on seismic recordings, where they usually appear as relatively inconspicuous first deflections. The phase velocity of P-waves is typically about 4 to 7 kilometers per second in most regions of the earth's crust. S-waves may be either horizontally or vertically polarized, in which case they are generally termed SH- and SV-waves, respectively. Phase velocities of S-waves are a little more than half those of P-waves. Body waves are able to penetrate deep into the earth's interior, whereas surface seismic waves are restricted to regions at and relatively near the earth's surface.

Two types of seismic surface waves are recognized in seismology, and they are called Rayleigh and Love waves. The Rayleigh wave propagates along the free surface of a solid half-space in a similar way as a gravity water-wave. It is characterized by particle motion confined within the vertical plane, which is parallel to the direction of propagation. The motion is elliptical and retrograde, with amplitudes diminishing exponentially with depth from the surface. This type of a wave was discovered theoretically by Lord Rayleigh (1880). It may be considered as a combination of a P-wave and an SV-wave. It propagates non-dispersively (phase and group velocities not functions of wavelength of period) in a homogeneous, isotropic medium, and has a phase velocity about 92% that of the corresponding S-wave.

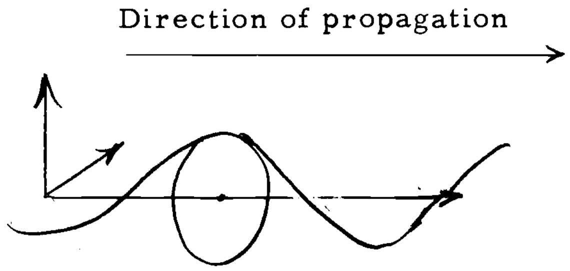


Figure 2.0-2. The Rayleigh surface wave.

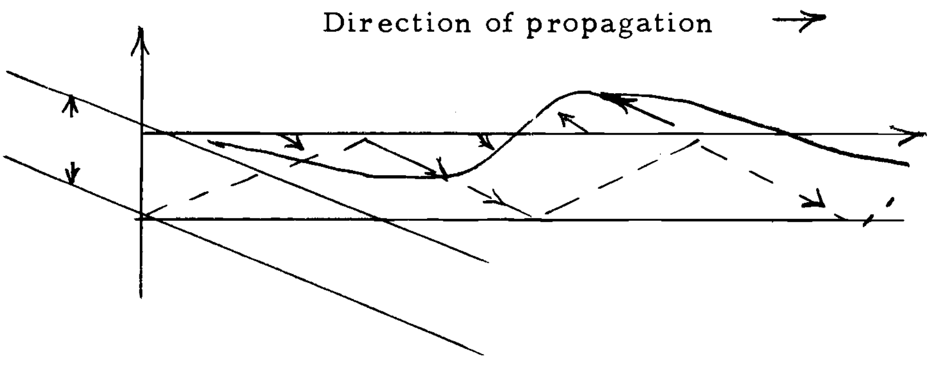


Figure 2.0-3. The Love surface wave.

Love waves are encountered where there is at least one shallow surface layer present. Love waves are the horizontal surface wave motions which occur along the earth's surface when there is a multiply reflected SH-wave component, channeled along the shallow surface layer between its upper and lower boundaries. The particle motion is in the horizontal plane and perpendicular to the direction of propagation. Unlike Rayleigh waves, Love waves are dispersive, and the propagation velocities depend on the wavelength of the wave and the thicknesses and properties of the surface and lower media. A condition for the existence of Love waves is that the S-wave velocity in the shallow surface layer should be less than that of the medium below. The Love-wave phase velocity will lie somewhere in between. This type of a surface wave was discovered by Love (1911) as corresponding to certain phases that were prominent in early land seismograms, when instruments used happened to be measuring mainly motion in the horizontal plane.

These different types of seismic waves represent specific cases of general solutions to the following pair of wave equations, which are applied to elastic wave propagation in seismology:

$$\frac{\partial^2 \phi}{\partial t^2} = \alpha^2 \nabla^2 \phi , \qquad \frac{\partial^2 \psi}{\partial t^2} = \beta^2 \nabla^2 \psi \qquad (2.0\text{-}1)$$

where α and β are the wave velocities, assumed constant; ∇^2 is the Laplacian operator, $\nabla^2 = \partial^2/\partial x^2 + \partial^2/\partial y^2 + \partial^2/\partial z^2$; and ϕ and ψ are scalar and vector potentials, which are defined such that the displacement $\vec{u} = (u_x, u_y, u_z)$ may be written:

$$\vec{u} = \nabla \phi + \nabla \times \psi \qquad (2.0\text{-}2)$$

The most general form of equations of motion at some point, x, y, z, in a medium is usually written as follows:

$$\rho \frac{\partial^2}{\partial t^2} u_x = \partial P_{xx}/\partial x + \partial P_{yx}/\partial y + \partial P_{zx}/\partial z + \rho X_x$$

$$\rho \frac{\partial^2}{\partial t^2} u_y = \partial P_{xy}/\partial x + \partial P_{yy}/\partial y + \partial P_{zy}/\partial z + \rho X_y$$

$$\rho \frac{\partial^2}{\partial t^2} u_z = \partial P_{xz}/\partial x + \partial P_{yz}/\partial y + \partial P_{zz}/\partial z + \rho X_z \qquad (2.0\text{-}3)$$

where P_{xx}, P_{xy}, etc. are components of stress at that point (see Figure 2.0-4), X_x, X_y and X_z are the vector components of a generalized long-range, or body force, such as gravity; and is the mass density, which is usually taken as constant. For a perfectly elastic solid, obeying Hooke's law, linear relations are assumed to hold between the components of stress, P_{xx}, P_{xy}, . . . and strain, e_{xx}, e_{xy}, . . . (see Figure 2.0-4) in such a way each stress component is given as a linear sum of all of the strain components, each multiplied by the appropriate elastic moduli. In mathematical work on seismic wave propagation, however, the media are assumed to be homogeneous and isotropic, and therefore, because of the symmetries

$$P_{ij} = -P\,\delta_{ij} + C_{ijkl}\,e_{kl}$$

$$e_{kl} = 1/2(\partial u_k/\partial x_l + \partial u_l/\partial x_k)$$

$$P = -1/3\,P_{ii}$$

i = direction perpendicular to plane

j = direction in which P_{ij} is acting

C_{ijkl} = constants

In a homogeneous, isotropic medium:

$$P_{ij} = -P\,\delta_{ij} + \lambda \frac{\partial u_k}{\partial x_k}\,\delta_{ij} + 2\mu\,e_{ij}$$

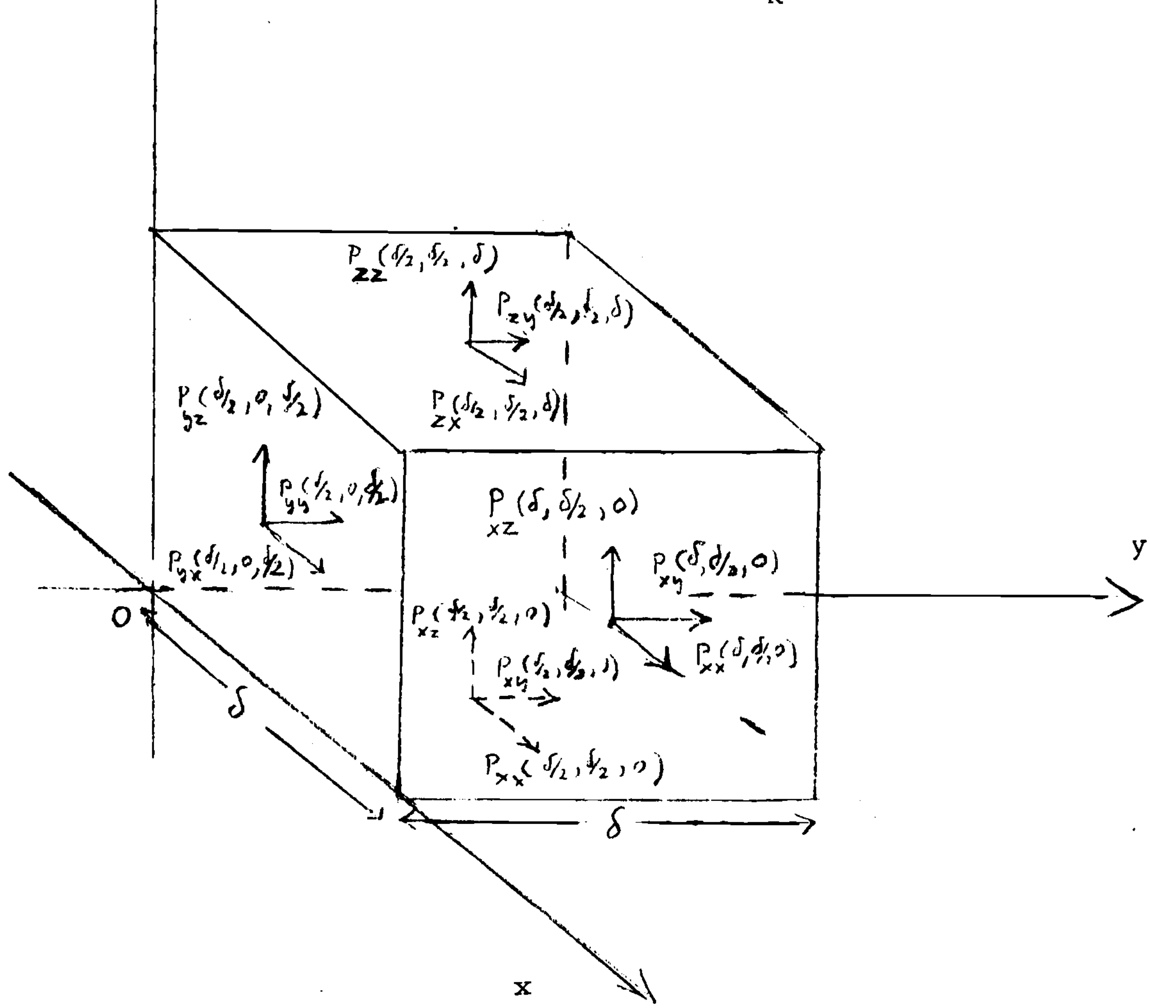


Figure 2.0-4. Components of stress on the faces of an infintesimal cubic volume.

involved between stress and strain components, only two elastic constants, Lame's constants, λ and μ, are required to describe relations between stress and strain completely. The equation 2.0-3 takes thus takes the much simpler form:

$$\rho \frac{\partial^2}{\partial t^2} u = (\lambda + \mu) \nabla (\nabla \cdot u) + \mu \nabla^2 u \qquad (2.0\text{-}4)$$

where the body forces have, as is usual in seismic work, been omitted, since they are relatively insignificant compared to the elastic forces in rocks. Equations 2.0-4 will be satisfied when and as defined by 2.0-2 are solutions of equations 2.0-1. Equation 2.0-1 has all simple plane-wave solutions of the following form:

$$u = (A_x, A_y, A_z) \exp\left[i \frac{2\pi}{\lambda} (\gamma_1 x, \gamma_2 y, \gamma_3 z \pm ct) \right] \qquad (2.0\text{-}5)$$

where A_x, A_y and A_z are constants of integration for each component of u, whose values depend on boundary conditions. The wavelength is $\lambda = 2\pi/k$, c is the phase velocity, and the direction cosines, γ_1, γ_2 and γ_3 obey the relation $\gamma_1^2 + \gamma_2^2 + \gamma_3^2 = 1$. Equations of the type 2.0-5, which represents a basic form of seismic wave solutions in rectangular coordinate systems, may be decomposed into three independent parts. One corresponds to a compression or P-wave component, traveling which a phase velocity given by $c = \alpha \equiv \sqrt{(\lambda + 2\mu)/\rho}$, and the other two represent SH- and SV-waves, whose phase velocities are equal to $c = \beta \equiv \sqrt{\mu/\rho}$.

For many solids, and for rocks of the earth in particular, the constants λ and μ are not appreciably different, and may for practical purposes be taken as equal (Poisson's hypothesis). This gives a ratio between the phase velocities, α and β, of $\alpha / \beta = \sqrt{3}$.

Expressions describing propagation of wave potentials (ϕ , ψ) or vector displacements (u) along horizontal boundaries between separate media may usually be written as products of functions of the vertical coordinate (z) and a propagating-wave factor (expik(x-ct) as follows:

$$\phi = f(z) \exp(ik[x-ct])$$

$$\psi = g(z) \exp(ik[x-ct]) \qquad (2.0\text{-}5)$$

where the functions, f and g, are of the type:

$$f(z) = A \exp(-ik \sqrt{c^2/\alpha^2 - 1}\, z)$$

$$g(z) = B \exp(-ik \sqrt{c^2/\beta^2 - 1}\, z) \qquad (2.0\text{-}6)$$

and A and B are appropriate constants. The x-axis has been chosen here along the direction of propagation. The arguments in the parentheses of equations are real, and the negative sign gives exponential decay of the amplitudes with depth (z), a necessary condition in the limit $z \longrightarrow \infty$ for all semi-infinite half-space systems. Equations of the form 2.0-5 and 2.0-6 characterize Rayleigh-wave solutions for semi-infinite, homogeneous, isotropic, half-spaces, bounded only by the free surface (z = 0). A generalized equation for Rayleigh waves may be written in a two-dimensional representation ($u_y \equiv 0$)

as follows:

$$u_x = a[\exp(.85kz) - .58\exp(.39\,kz)]\sin[k(x-ct)]$$

$$u_z = a[-.85\exp(.85\,kz) + 1.47\exp(.39\,kz)]\cos[k(x-ct)] \qquad (2.0\text{-}7)$$

where a is a constant, and c and k are the phase velocity and wave number of the wave, and c being equal to 0.92 β .

In Love-wave solutions an additional expression is required for u_y, because of the transverse horizontal wave motions characteristic of Love waves. In the similar expression for u_y,

$$u_y = h(z)\exp[ik(y-ct)] \qquad (2.0\text{-}8)$$

the function h has the more complete form:

$$h(z) = C\exp(-ik\sqrt{c^2/\beta^2-1}z) + D\exp(ik\sqrt{c^2/\beta^2-1}z) \qquad (2.0\text{-}9)$$

C and D being constants, depending on boundary conditions. As Love waves also require at least two adjacent media, there must moreover be at least two of each of the above solutions; one for each of the solid media with the appropriate physical parameters. All the seismic wave solutions are, of course, subject to appropriate boundary conditions; namely vanishing stresses at the free surfaces, convergence of solutions (for ϕ, ψ, $\vec{u}$, etc.) to zero at infinite depth (z), and continuity across interfaces. Love propagation is dispersive, as already stated, and the velocities depend on wave numbers or frequencies of the waves, and on the thicknesses and physical properties of the solids. Although

Love waves are common constituents of microseisms in general, they will not be discussed further here, since the Oregon coastal observations to be presented in this thesis involve measurements only of the vertical component of ground motion and would therefore not contain any contribution from Love waves.

3.0. GENERATION OF MICROSEISMS BY STANDING WATER-WAVES

As was shown by Miche (1944), there is, in the second order mathematical approximation to the standing wave, a pressure term which does not attenuate with depth in water. It is proportional in magnitude to the square of the product of the wave height and frequency and oscillates with a frequency twice that of the original wave. This pressure variation, averaged over a wavelength, λ, of the wave in an irrotational, incompressional fluid, is given at any depth, z, in two dimensions by Longuet-Higgins as follows:

$$P_2 = \frac{1}{\lambda}\int_x^{x+\lambda} [P-P_a-g\rho z]\, dx = \rho/\lambda\int_x^{x+\lambda} [\frac{\partial^2}{\partial t^2}(\frac{1}{2}\eta^2) - w^2(z)]\, dx$$

$$\approx \frac{\rho}{\lambda}\int_x^{x+\lambda} \frac{\partial^2}{\partial t^2}(\frac{1}{2}\eta^2)dx = 2\rho a_1 a_2 w^2 \cos(2wt) \qquad (3.0\text{-}1)$$

The free surface elevation, η, may be written as a sum of two opposing trains of progressing waves:

$$\eta = a_1 \cos(kx - wt) + a_2 \cos(kx + wt) \qquad (3.0\text{-}2)$$

having equal wave numbers, k, and frequencies, w. The quantity, P_2, represents the mean of the difference between the total pressure, P, and the atmospheric pressure, P_a (assumed constant), and the weight, g z, of the water column. The vertical particle velocity, w(z), at the depth z is omitted in the last term of equation 3.1-1, as it becomes negligible for large z. Equation 3.0-1 represents a non-linear pressure (P_2) which arises from the second time-rate of change of the potential energy of the water column, averaged over an area of the sea surface, and it is non-zero only when there is standing-wave action on the ocean surface. It was derived by Longuet-Higgins (1950) directly from the equations of motion for an incompressible, irrotational fluid, and it applies to water depths ranging from half a wave-length to the order of kilometers, which characterize the deep-ocean. At depths exdeeding half a wavelength the second-order pressures (P_2), given by equation 3.0-1, predominate overwhelmingly over the first-order pressure, whose magnitudes vanish exponentially with the water depth. The corresponding result of equation 3.0-1 vanishes, on the other hand, when applied to progressing-wave systems.

If the water depths are considerable, such as those typical of deep oceans, the compressibility of water, which was neglected by equation 3.0-1, must be taken into account. The pressure variations (P_2) will then be propagated through the ocean as compression waves with a phase speed (c) equal to that of sound in water, and the expression for P_2 is dependent on the vertical coordinate, z, and the total water depth, h, as follows:

$$\tilde{P}_2 = \frac{\rho}{2} \frac{a_1 a_2 w^2}{\cos(2w \frac{z}{c})} \cos\left(\frac{2w(z-h)}{c}\right) \cos(2wt) \qquad (3.0\text{-}3)$$

Cases where $\cos(2wz/c) = 0$ correspond to a resonance situation in which there is theoretically an infinite-fold increase in the pressure (P_2), and this occurs at depths for which $2wz/c = \pi(n+1/2)$, $n = 0, 1, 2, \ldots$. The pressure is not infinite, however, since energy is propagated away from the source region in all directions in the form of a coupled system of compression waves in the water and elastic (seismic) waves along the ocean bottom, and the pressure amplification amounts to no more than five- or six-fold increase at most. In actual cases there would also be energy losses due to viscous damping, which is being neglected in the mathematical analysis.

Depths in the ocean at which resonance should occur may be shown, as in Figure 3.0-1, as a function of the period of (standing) gravity waves on the ocean surface. The solid lines show the resonance-peak depths for the first two modes ($n = 0$, $n = 1$), and the dashed curve is a sketch of a hypothetical sea-wave spectrum, whose period of maximum energy is at about 10 seconds, a value rather typical for gravity waves in general. According to Figure 3.0-1, the predominant, 10-second components in the wave spectrum should cause resonance at depths of about 1.5 and 5.5 kilometers for the first and second modes, respectively. In alternative terms, resonance should be experienced at the typical, deep-ocean depth of 4 kilometers from waves having periods of about 7 and 16.7 seconds.

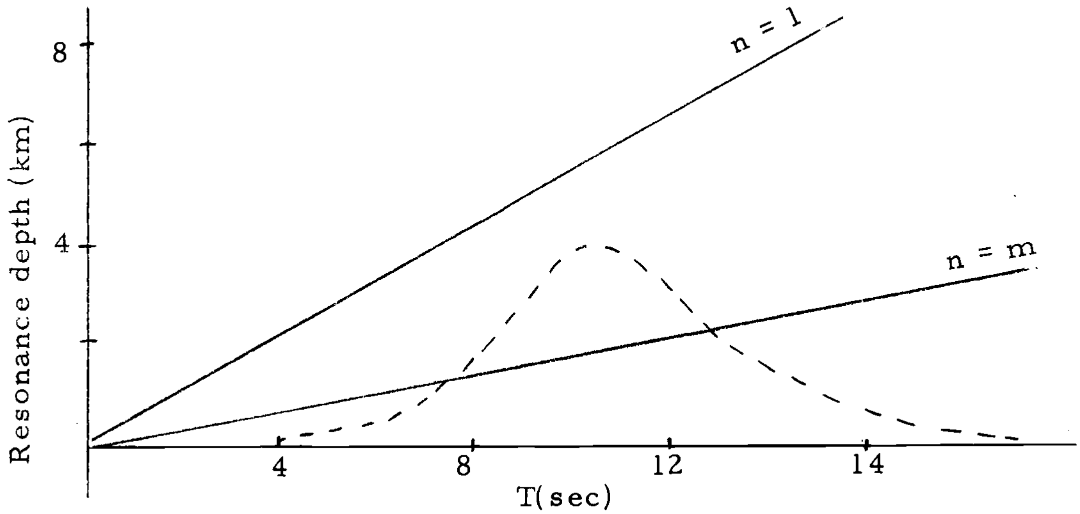


Figure 3.0-1. Resonance depth in the ocean as a function of the sea-wave period for the first two modes. Dashed curve is a qualitative, hypothetical sea-wave spectrum.

Longuet-Higgins explained the presence of the non-linear pressure variations intuitively by pointing out that the potential energy of the average water column changes periodically for an area containing a standing water-wave system. Considering the simple case of a "long-crested" standing-wave group (for which the sea surface elevation may be described in two dimensions by equation 3.0-2), the center of gravity of the water column is higher at the times of maximum elevation or displacement of the wave profile than at instants of time, when the water surface is in its equilibrium position. It oscillates between these two levels with a period half that of the wave itself. There must consequently be forces exerted on the bottom of the water column to balance the accelerations. In contrast to the standing-wave case, the center of gravity for a train of progressing waves remains unchanged in its vertical position, and there are hence no such forces exerted on the sea bed associated with progressing waves.

Equations 3.0-1 and 3.0-3 may be obtained as approximations from the following, more complete, second-order solution for the wave potential, given by Longuet-Higgins for the standing wave:

$$\phi_2 = -\frac{3w}{8}\frac{\cosh 2k(z-h)}{\sinh^4(kh)}\left[a_1^2 \sin 2(kx-wt) - a_2^2 \sin 2(kx+wt)\right]$$

$$-\frac{w}{8}\frac{\cosh(3kh)\cos(2w(z-h)/c)}{\sinh 2kh \cosh kh \cos(\frac{2wh}{c})} 2a_1 a_2 \sin(2wt)$$

$$+\frac{(a_1^2 + a_2^2)\, w^2 t}{4 \sinh^2(kh)} \qquad (3.0\text{-}4)$$

where

$$a_1 = wb_1/2k^2 \sinh(kh),$$

$$-a_2 = wb_2/2k^2 \sinh(kh),$$

and b_1 and b_2 are constants which appear in a first-order solution for the potential. Equation 3.0-4 represents a solution contained in the second term in the series:

$$\phi = \epsilon\,\phi_1 + \epsilon^2\,\phi_2 + \ldots \qquad (\epsilon \text{ is a small parameter}), \qquad (3.0\text{-}5)$$

which was found by a method of successive approximation to the following differential equation:

$$\frac{\partial^2 \phi}{\partial t^2} - c^2 \nabla^2 \phi - g \frac{\partial \phi}{2z} - \frac{\partial}{\partial t}(\frac{1}{2} u^2) - u \cdot (\frac{1}{2} u^2) = 0 \qquad (3.0\text{-}6)$$

in which the compressibility of water was taken into account. The solutions, which are subject to appropriate boundary conditions, are based on similar series expansion as 3.0-5 of the particle-velocity in the water, the free surface elevation, pressure and density, and as well as of ϕ, and by Taylor series expansion of the free surface conditions about the reference plane, $z = 0$.

Longuet-Higgins gave formulas for computing the pressure forces on the sea bed associated with generalized wave conditions on an ocean surface. An actual sea surface always contains mixed spectrum of waves of varying heights and periods. The second-order pressure variations discussed above result only from the interaction of opposite-traveling wave components, which themselves are distributed over a range of amplitudes and frequencies.

The instantaneous free surface elevation, η, at any point (x, y) of the sea surface may be represented as an integral of a continuous spectrum of wave components over a range of frequencies and wave numbers as follows:

$$\eta(x, y) = R \int\int_{-\infty}^{\infty} A(u, v) \exp i(ukx + vky + wt) du dv$$

$$= \int\int_{-\infty}^{\infty} [A(u, v) \exp(iwt) + A^*(-u, -v) \exp(-iwt)] \cdot \exp[i(ukx + vky)] du dv \qquad (3.0\text{-}7)$$

where A and A* are complex, and complex conjugate, amplitudes (functions of the non-dimensional wave number variables, u and v), of ideally long-crested wave components, which travel in a direction parallel to the line given by ux + vy = 0. Their wavelengths, λ , are given by $\lambda = 2\pi / \sqrt{u^2+v^2}\, k$, where k represents a wave number which is characteristic for the given wave spectrum. The angular frequency w, k, u, v and the water depth, h, are related by the familiar equation $w^2 = g\sqrt{u^2+v^2}\, k \tan h \sqrt{u^2+v^2}\, kh$. The amplitude of the wave components are evaluated from the area integral of the sea surface elevation:

$$\frac{1}{2}[A(u, v) + A^*[-u, -v)] = (\frac{k}{2\pi})^2 \iint_{\Lambda} \eta(x, y)\exp[-i(ukx+vky)]dxdy \tag{3.0-8}$$

where Λ denotes the given, finite, area of the sea surface representing the generating region. According to Longuet-Higgins, the integral of the squared sea surface elevation over the domain, Λ, may be written:

$$\frac{1}{2}\iint_{\Lambda} \eta^2 dxdy \equiv \frac{1}{2}\iint_{-\infty}^{\infty} \eta'^2 dxdy = \frac{1}{2}(\frac{2\pi}{k})^2 \iint_{-\infty}^{\infty} \frac{1}{2}(A'e^{iwt} + A'^*e^{-iwt})^2 dudv \tag{3.0-9}$$

where the primed variables A' and A'* and η' were introduced (by Longuet-Higgins), for convenience, to represent those equivalent variables, which would give the same result in equations 3.0-9 for the integration of the first term performed over the entire sea surface. The variable, η', is defined such that it is equal to η inside Λ, but zero outside, and the corresponding primed amplitudes, A' and A'*, were described as "blurred" components, which contained the influence from adjacent components in the uv-plane. Differentiation of equations 3.0-9 (last term) twice with respect to time then yields the total, second-order pressure force (assuming no resonance effects) due to the sea-wave activity in Λ:

$$F_{\Lambda} = \iint_{\Lambda} \tilde{P}_2 = -4\rho w^2 \, R\left[\left(\frac{\pi}{k}\right)^2 \iint_{\Omega} A'A'_{-} \exp(2iwt)\,du\,dv\right] \qquad (3.0\text{-}10)$$

For the general case of two independent, interacting sea-wave groups Longuet-Higgins reduced equation 3.0-10 to the following expression:

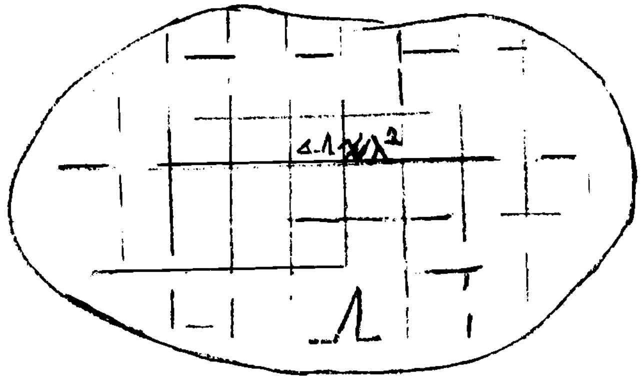
$$F_{\Lambda} = 4\pi\rho\, \bar{a}_1 \bar{a}_2 w^2 \left(\Lambda\Omega_{12}/\Omega_1\Omega_2\right)^{1/2} \exp(2iwt) \qquad (3.0\text{-}11)$$

where a_1, a_2, Ω_1 and Ω_2 are the mean amplitudes and wave number regions characterizing the two wave groups, w is the mean angular frequency of the resulting standing waves and Ω_{12} is the overlap of either of the wave number regions with the mirror image of the other. Equation 3.0-11 was derived using the fact that the amplitude of the total, overall force exerted by a collection of a large number (N) of independent, identical, oscillating point forces is proportional to the square root of their number (Longuet-Higgins, 1950). Longuet-Higgins divided the sea surface area, Λ, into a number of smaller, equal-sized, square areas, $\Delta\Lambda$ (see Figure 3.0-2), of dimensions similar to the characteristic wavelength of the waves in Λ. These area subdivisions were regarded as independent point oscillators; the (standing) wave motions in each were assumed in phase but randomly related to the phases of the wave motions in any of the other squares, and the whole region Λ was characterized by the same mean wave amplitudes ($\bar{a}$) and frequency (w). The effective non-linear force over the whole region Λ should then be approximately:

$$F_{\Lambda} = F_{\Delta\Lambda}\sqrt{N} = 2\rho\,\bar{a}^2\,w^2\,\Delta\Lambda\sqrt{\Lambda/\Delta\Lambda}\,\exp(2iwt) \qquad (3.0\text{-}12)$$

The wave number regions (Ω_1, Ω_2) in the uv-plane were also thought of as divided into a number(s) (Ω_1, Ω_2) of independent unit squares, containing wave components uncorrelated in phase. Mean values of $\bar{A}$ and $\bar{a}$ for a given wave group are, according to Longuet-Higgins (1950), related by $\bar{A} = \bar{a}k/\sqrt{\Omega}$. Equations 3.0-10 and 3.0-11 are practically equivalent for cases of $|\Omega_1| = |\Omega_2| = |\Omega_{12}|$ and $\Delta\Lambda = \lambda = 2\pi/k$.

Given an oscillating force F_{Λ}, at the sea surface, vertical oscillations (δ) of the ocean bottom at a distance r away will then be, according to Longuet-Higgins:



$$|F_{\Delta\Lambda}| - P_2\,\Delta\Lambda = 2\rho\,\bar{a}_w^2\,w'\,\Delta\Lambda$$

$$|F_\Lambda| = F_{\Delta\Lambda}\sqrt{N} = 2\rho\,\bar{a}_w^2\,w^2\,\Delta\Lambda\sqrt{\Lambda/\Delta\Lambda}$$

$$\Delta\Lambda \sim \lambda^2 = (2\pi/k)^2$$

Figure 3.0-2. An hypothetical generating region, Λ, containing standing waves, and divided into equal, smaller squares of area $\Delta\Lambda \approx \lambda^2$, where λ is the characteristic length of the waves.

$$\delta = \left| F_{\Lambda} \right| \frac{\sqrt{2\,w}}{\rho_2 \sqrt{2\pi\, r}\ \beta^{5/2}} \left(\sum_n C_n^2\right)^{1/2} \exp(ziwt) \qquad (3.0\text{-}13)$$

where ρ_2 is the density of the sea bed, β' is the S-wave phase velocity. The constants, C_n, represent contributions to the amplitudes of each of the resulting normal modes, n, of the seismic wave-motions along the ocean bottom, whose values depend on the depth of the ocean as well as on the physical properties of the sea bed. The distance, r, is assumed large, compared with the dimensions of the generating region, so that the latter may be regarded as an harmonic point forcing function, acting at the sea surface and generating a system of compression and elastic waves in the ocean and ocean bottom, which propagate radially in all directions from the source region. Equation 3.0-13 was developed by Longuet-Higgins by using a contour integration method from Sommerfeld (1909) and Jeffreys (1926) from the following integral expression, given by Scholte (1943):

$$\delta(\sigma, r)\exp(i\sigma t) = -\frac{1}{2\pi} \int_0^{\infty} \frac{J_0(\xi r)\,\xi\, d\xi}{\rho_2\, \sigma^2\, G(\xi)} \exp(i\sigma t) \qquad (3.0\text{-}14)$$

Equation 3.0-14 expresses the vertical motion of the sea bed due to a unit periodic point force, $\exp i\sigma t$, acting at the sea surface. Here σ denotes the generalized frequency of excitation and wave motion, J_0 is a Bessel function of the first kind of zero order, and the function, $G(\xi)$, is defined by:

$$G(\xi) = (\beta/\sigma)^4 \left[(2\xi^2 - \sigma^2/\beta^2)(\xi^2 - \sigma^2/\alpha_2^2)^{-1/2} - 4\xi^2(\xi^2 - \sigma^2/\beta^2)^{1/2} \right] \cdot \cosh(\xi^2 - \sigma^2/\alpha_1^2)^{1/2} h + (\rho_1/\rho_2)(\xi^2 - \sigma^2/\alpha_1^2)^{1/2} \sinh(\xi^2 - \sigma^2/\alpha_1^2)^{1/2} h \qquad (3.0\text{-}15)$$

where ξ is a variable of integration in wave number space. In equation 3.0-15, as throughout this thesis, parameters subscripted "1" refer to properties of the uppermost medium in a layered half-space; in this case of the water, and "2" to those of the (solid) medium below that, i.e., of the ocean bottom, etc. The constants C_n are found, using the formula:

$$C_n = (-1)^n (\beta/\sigma)^{5/2} \xi_r^{1/2} /[dG(\xi_{11})/d\xi_n] \qquad (3.0\text{-}16)$$

where ξ_1, ξ_2, . . . are the (finite) number of positive "zeroes" of $G(\xi)$. Values of C_n for the first four modes, as computed by Longuet-Higgins, are shown in Figure 3.0-3 as functions of the non-dimensional frequency, $\sigma h/\beta$, for the theoretical, two-layered ocean-earth model, having the physical parameters as indicated. The peaks on the curves are associated with resonance between the forcing function and the response of the ocean bottom, which will occur for certain depths in a deep ocean. The only relevant value of C_n used in the present study is $C_1 = 0.24$ for shallow water (see Figure 3.0-3).

The peaks on the first two curves (for C_1 and C_2) in Figure 3.0-3 lie at abcissa values of about $\sigma h/\beta = 0.8$ and $\sigma h/\beta = 2.7$, which

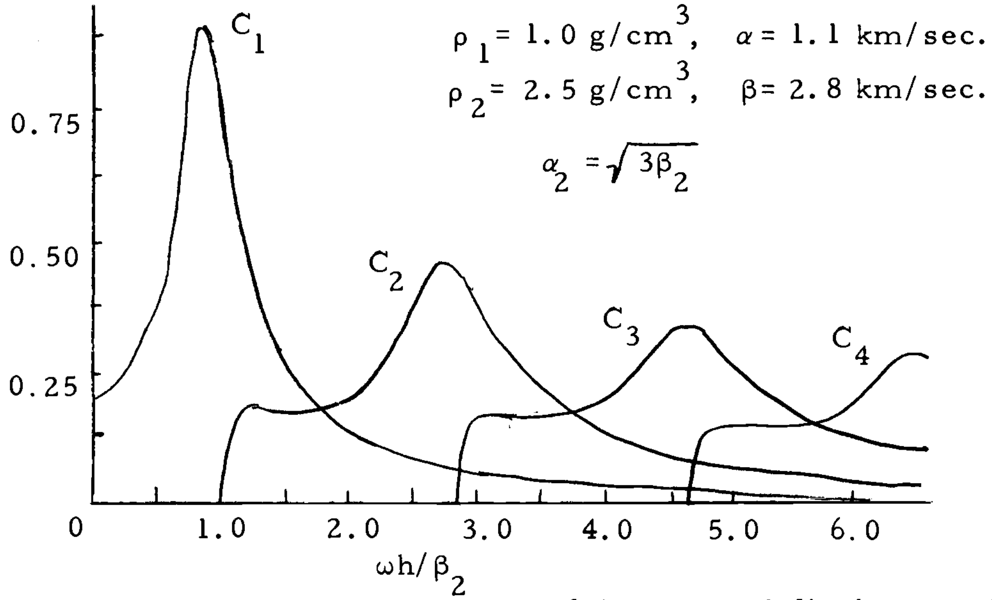


Figure 3.0-3. The amplitude of the vertical displacement of the sea bed as a function of depth (from Longuet-Higgins, 1950).

correspond for a typical open-ocean depth of 4 kilometers (and the same shear wave velocity value $\beta = 2.8$ kilometers/second) to resonance periods of about 10 seconds and 3.3 seconds. These periods are seen to be close to limits of the period range (3 to 9 seconds) obtained from the points of minimum group velocities on the dispersion curves (Figure 1.3-2) for the ocean-earth model studied by Press and Ewing (1948).

As an example to illustrate the orders of magnitude characterizing microseisms generated by standing water waves in general, one may consider an area of, say, one square kilometer of the surface of a relatively shallow ocean (so that the possibility of resonance can be disregarded), on which there is a narrow spectrum of unit-amplitude standing-wave motion. If the period of the waves is assumed to be 10 seconds, which is a value rather typical for most sea states, then dividing the region, Λ, into smaller squares, $\Delta\Lambda$, having sides equal to one wavelength of the waves, one obtains the number of $N = \Lambda/\Delta\Lambda = 45$ independent point oscillators. Then, by equating 3.0-13 and Figure 3.0-3, the following vertical ground oscillation amplitude ($\bar{\delta}$) is found for a distance away of, say, 10 kilometers.

Given:

$$\Lambda = 1\ km^2 = 10^6\ m^2$$

$$\Delta\Lambda = \lambda^2 = (150)^2 = 2.25 \times 10^4\ m^2$$

$$a_w = 1\ m$$

$$w = 2\pi/10$$

$$\rho_1 = 1\ gm/cm^3$$

$$\rho_2 = 2.5\ gm/cm^3$$

$$\beta = 2.8\ km/sec$$

$$r = 10\ km = 10^4\ m$$

$$C_1 = 0.24$$

The forcing function (magnitude) will be:

$$|F_\Lambda| = 2\rho_1 a_w^2 w^2 \Delta\Lambda\sqrt{\Lambda/\Delta\Lambda} = 8.9 \times 10^2\ nt/m^2$$

and the amplitude of the microseisms:

$$\delta = \frac{\sqrt{2w}\ C_1}{\rho_2 \beta^{5/2}\sqrt{2\pi r}} F_\Lambda m = \frac{3.3 \times 10^{-13}}{\sqrt{r}} |F_\Lambda| m$$

$$= \frac{3.3 \times 10^{-13} |F_\Lambda|}{\sqrt{r}} \times 10^6 \text{ microns} = 0.3 \text{ microns} \qquad (3.0\text{-}17)$$

Longuet-Higgins compared two other hypothetical cases of microseisms generation, one involving sea-wave interaction in mid-ocean, and the other was a case of standing-wave systems created near continents by the reflection of waves from the shores. He concluded that only standing-wave action in the open oceans would be effective enough to generate microseismic activity of the magnitude which are typically observed. Coastal reflection coefficients would generally be too small for most beaches to account for microseisms, except perhaps at stations near shores, even though the areas of the generating regions may be quite considerable.

4.0 MICROSEISMS AT THE OREGON COAST

4.1 Measurements and Instrumentation

Local sea wave heights and vertical ground motion were simultaneously measured (by Mr. Clayton Creech) during October 25 to 30, 1972, at the coastal sites shown in Figure 4.1-1. A total of six continuous ten-minute recordings were obtained of both sea waves and microseisms, of which only three of the sets, owing to limited funds available for computer processing, were analyzed and presented here. The measurements were made at the following dates and times, with the "observation numbers" being used here, as in the following, to identify each pair of sea wave and microseism records:

Observation no.	Hours of day	Month	Day	Year
1	1550-1600	Oct.	25	1972
2	1250-1300	"	26	"
3	1640-1650	"	30	"

An ocean wave pressure transducer, a Bendix model A-2, was used for measuring the pressures in the water, which fluctuated with the wavelike motions of the sea surface directly above. Installed in about 13 meters of water, the instrument gave continuous recordings of the water pressure fluctuations, from which the corresponding sea wave heights were mathematically determined. The wave heights were,

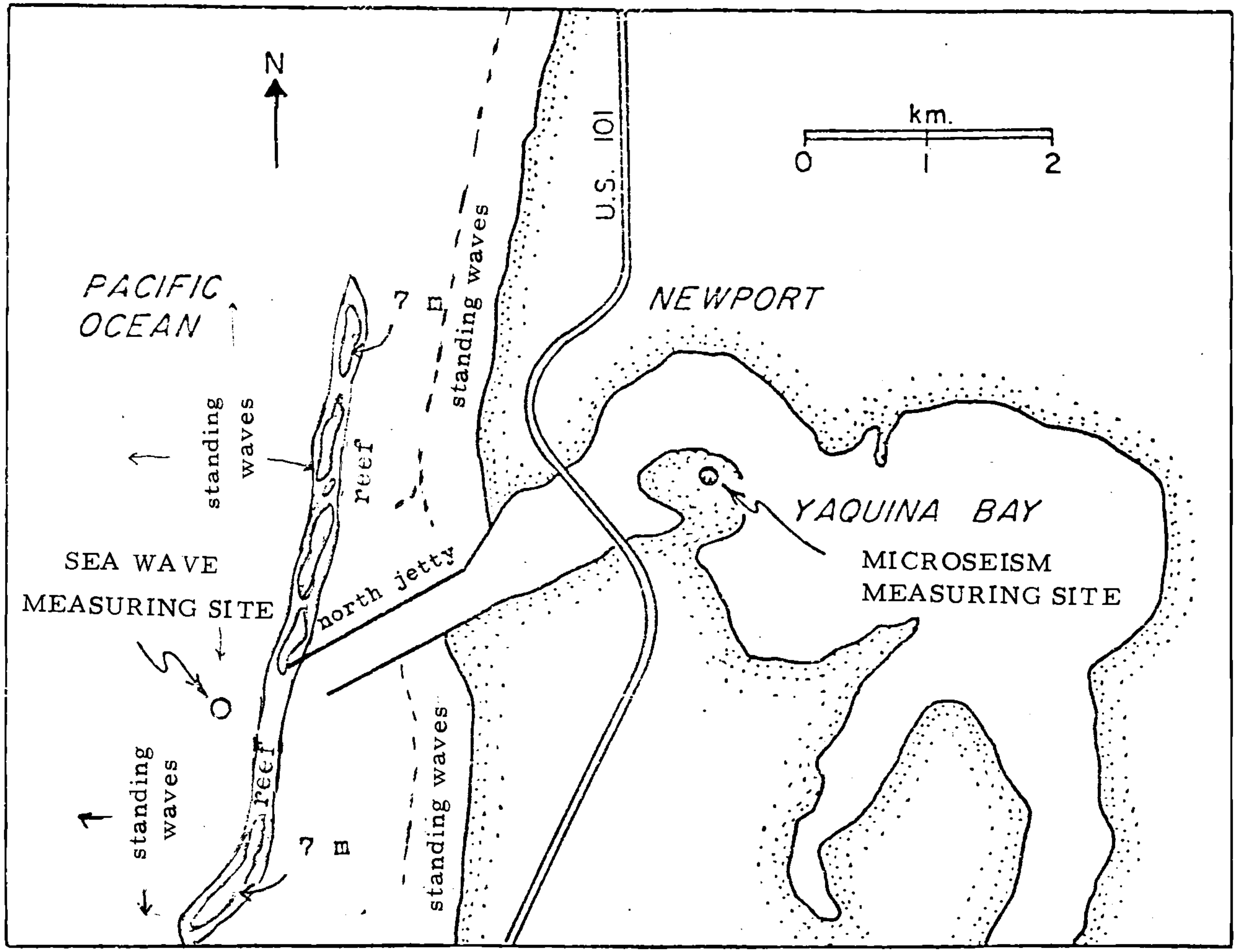


Figure 4.1-1 Site location of microseismic and sea-wave measurements (from Zopf, 1972)

according to linear wave theory, taken as proportional to the (amplitude of the) fluctuating weight of the water column, plus a small (13%) correction which was added to include the exponential decay of the water-wave pressure with depth, as estimated to occur for waves of the typical 10-second periods at that depth (13 meters).

Microseismic measurements were made with a portable, long-period, vertical-velocity (Teledyne-Geotech, model SL-220) seismometer. The instrument is a sensitive electromagnetic moving-coil device, which converts periodic, vertical ground motion (velocity) into electrical output. It has an adjustable resonance frequency, which was set at 18 seconds. Estimates were made of the root-mean-square (rms) amplitudes of the displacement of the earth's surface directly from the raw seismometer records, using an approximate gain factor of 1600, which, according to Zopf (1972), applies for the instrument over the limited (short-period) frequency range of interest. (The predominant periods in all three seismic records are about 5 seconds, and the energy distributions are about the same.) This gain factor is obtained from the following approximation formula given by Zopf (personal communication) for the predominantly observed period of 5 seconds:

$$\text{gain} \equiv \frac{C_0}{|\delta|} \approx 2000 \qquad (4.1\text{-}1)$$

where C_0 is the deflection (rms) on the seismic chart, corresponding to the ground displacement amplitude ($|\delta|$, rms) in the same units.

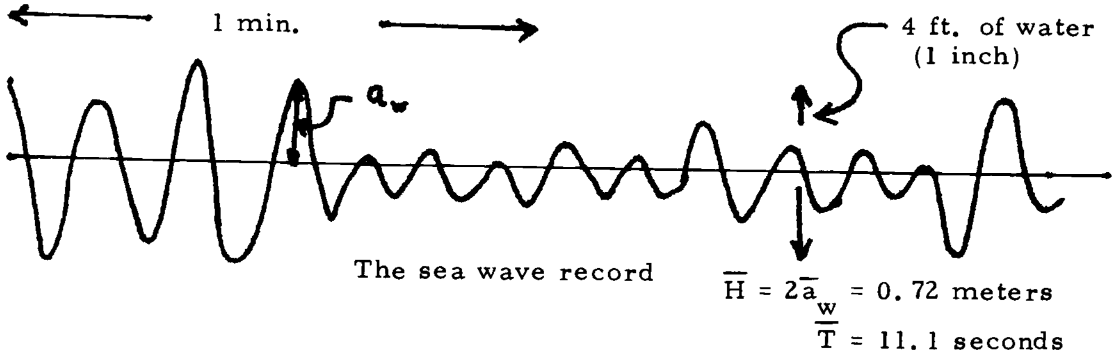
Segments of the sea wave and microseismic records are illustrated in Figure 4.1-2, and pertinent parameters to be used in the following defined there.

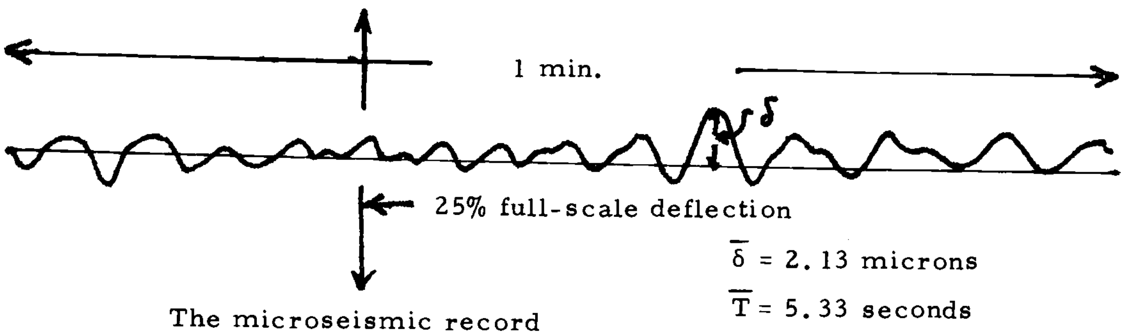
4.2 Data Analysis

The recordings were digitized, and calculations were made of power spectra and rms values of the sea and ground wave amplitudes, using the FORTRAN subroutine *SPECT2 of the Oregon State University computer library, and the conversion formula 4.1-1. On the pressure chart, a one-inch deflection corresponds to a four-meter column of water. The rms wave amplitudes, $\bar{a}_w$, were computed from the first-order pressures, $\bar{P}_1$, according to:

$$\bar{a}_w \simeq (\bar{P}_1/\delta g) \cosh kh \simeq 1.15 \times 10^{-4}\, \bar{P}_1 \qquad (4.1\text{-}2)$$

where $\bar{P}_1$ was determined from the chart deflections (rms), and k was, for convenience, taken to be the (constant) wave number of 10-second waves, and h is the depth (13 meters) of the water. The (rms) ground wave amplitudes, $\bar{\delta}$, were plotted against the square of the ratio of the corresponding (rms) sea wave amplitudes, $\bar{a}_w$, to the predominant sea-wave periods, $\bar{T}_w$. The spectra of the (inferred) forcing functions, i.e., of the non-linear pressure variations ($\bar{P}_2$) due to standing waves,





$H_w = 2a_w$ sea wave height (m)

δ = seismic wave amplitude (μ)

T_w = sea wave period (sec)

T_s = seismic wave period (sec)

$\overline{H}_w$, $\overline{a}_w$, $\overline{\delta}$ = rms values

$\overline{T}_w$, $\overline{T}_s$ = predominant (spectral) peak) periods

Figure 4.1-2. Segments of the sea wave and microseism recordings (obs. no. 1).

which were taken as proportional to $(\bar{a}_w/\bar{T}_w)^2$, were roughly estimated from the measured sea-wave spectra, and empirical transfer functions relating the microseisms to the non-linear sea-wave pressures were determined therefrom.

For comparison with values of microseism amplitudes derived from the data, theoretical estimates were also made of the ground oscillations, which would occur at the measuring site due to the effects of a hypothetical standing-wave system, assumed located in a narrow, shallow-water region along the entire Oregon coastline. The width of this region was arbitrarily taken as 400 meters for reasons which will be discussed further in the following section. The standing-wave system was regarded as composed uniformly of 10-second incoming waves, interacting with reflected waves of identical period and of amplitudes determined by reflection coefficients (R) assumed for the shore. Calculations were performed, using the following series expression:

$$\delta = 1.5 \times 10^{-2} a_w^2 R \beta^{-5/2} \sum_{n=1}^{120} (1+6.25 n^2)^{-1/4} \qquad (4.1\text{-}3)$$

and the following range of parameter values:

R = the reflection coefficient = 0.01 to 0.1

β = the shear wave phase velocity of the local sea bed = 2.0 to 3.0 km/sec

n = 1, 3, . . ., 120 = the total number of terms in the series 4.1-3 (each representing a 5-kilometer coastal segment of a total length of 1200 kilometers of the coastline)

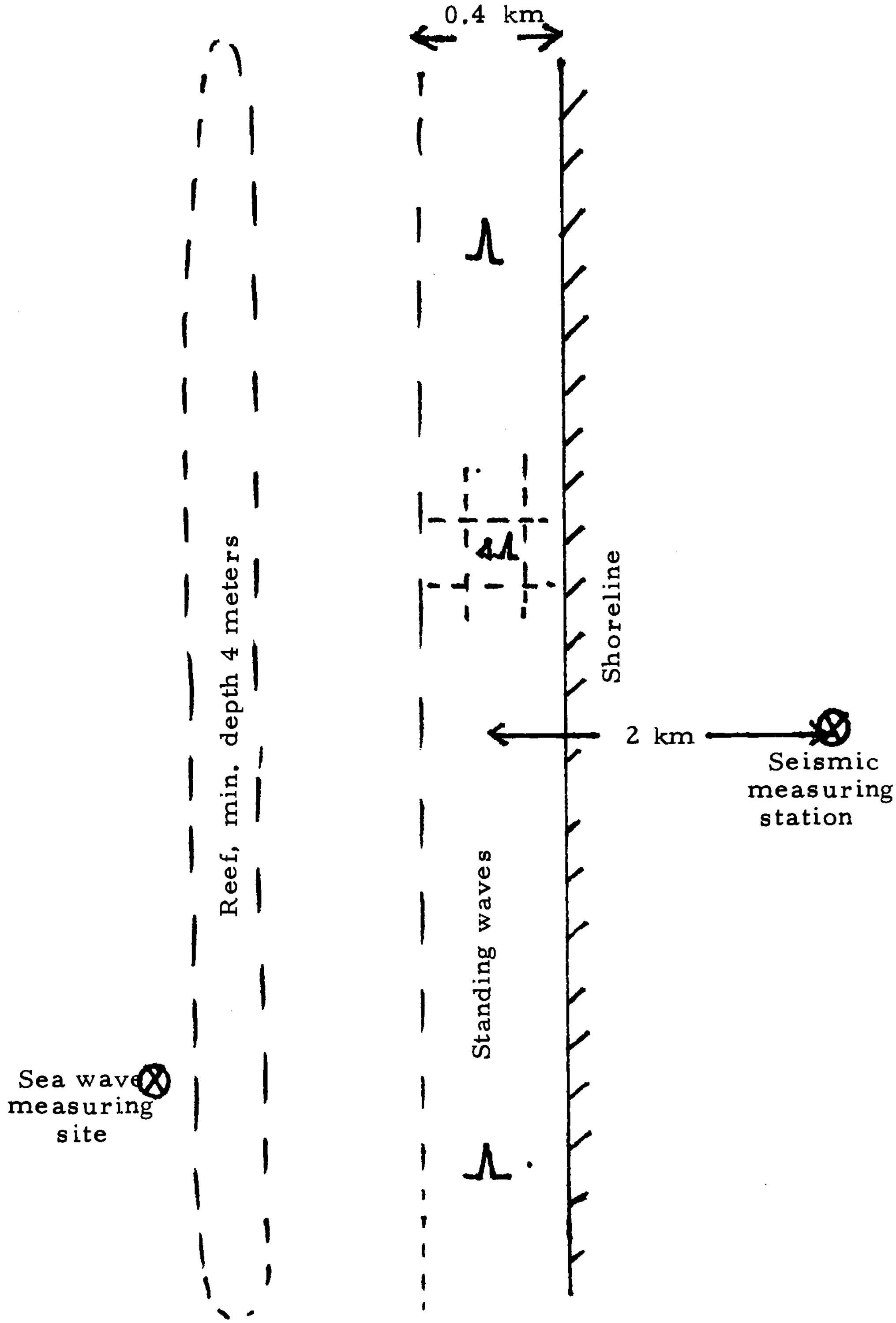


Figure 4.2-1. Schematic illustration of the probable (hypothetical) long-shore region of standing wave activity in relation to shore and measuring sites.

Equation 4.1-3 was derived directly from equations 3.1-1, 3.1-11 and 3.1-13 and represents an (approximate) expression for the amplitude of the microseisms, which should be produced by a hypothetical, 10-second standing-wave system, acting in a 1200-kilometer long and 0.4 kilometer wide coastal region, centered at the site of the measurements. A ground density value of about 2.0 gm/cm^3 was assumed for the sea bed.

The increase in incoming-wave height, which should occur due to shoaling on the beach, was accounted for, and it was estimated, on the basis of the refraction factor graph in Figure 4.2-2, to be about 1.3 times the deep-water wave height. The results for the hypothetical microseism amplitudes were then graphed as functions of the (hypothetical) incoming-wave amplitudes ($\bar{a}_w$) for different values of R, β_2 and $\Delta\Lambda$.

The relative contribution, if any, to the microseismic activity at Newport, Oregon, by the shoaling-water, wave-modulation mechanism, proposed by Hasselmann (1963), or by wave breaking on the beach, was investigated. These mechanisms produce ground oscillations of the same frequency as the sea waves, which may be identified as energy-peaks in the microseism spectra.

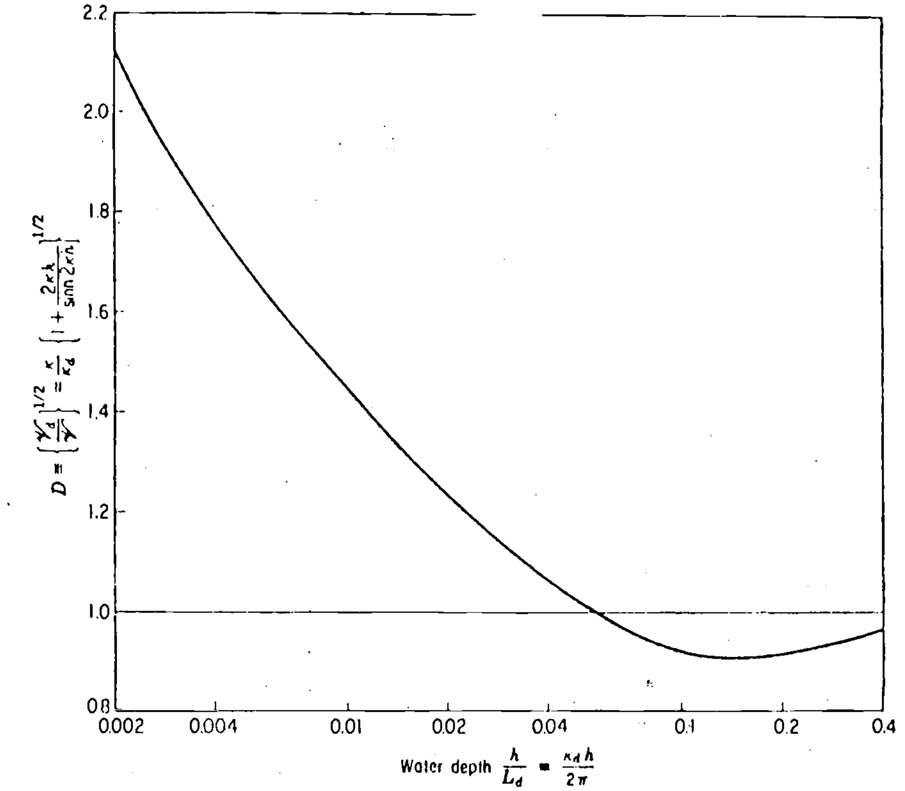


Figure 4.2-3. The refraction factor D as a function of water depth (from Kinsman, 1965).

4.3 Results and Discussion

The power spectra of the sea waves and microseisms, as computed from the data, are shown in Figures 4.3-1 to 4.3-3, with the (rms) amplitudes and predominant periods indicated for each. The microseisms are found to be in the order of several microns, ranging from about 2 to 11 microns of vertical oscillation, and they are associated with sea wave heights (a_w, rms values) from 0.4 to 4.5 meters. The power spectra are all distinctively peaked about definite predominant frequencies; or about 0.1 cps for the sea waves and about

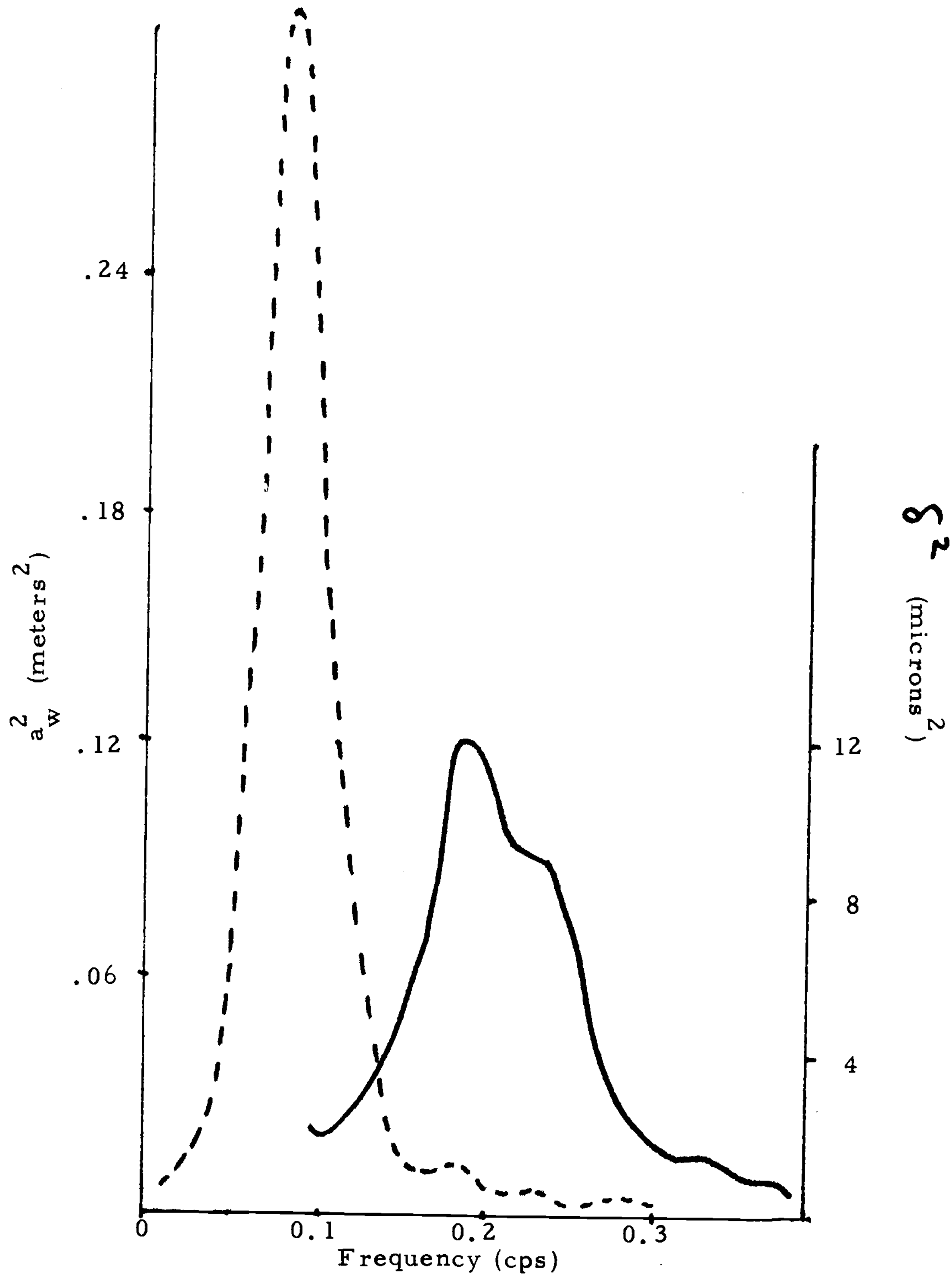


Figure 4.3-1. Power spectra as computed from sea wave (a) and microseismic (b) records (obs. no. 1).

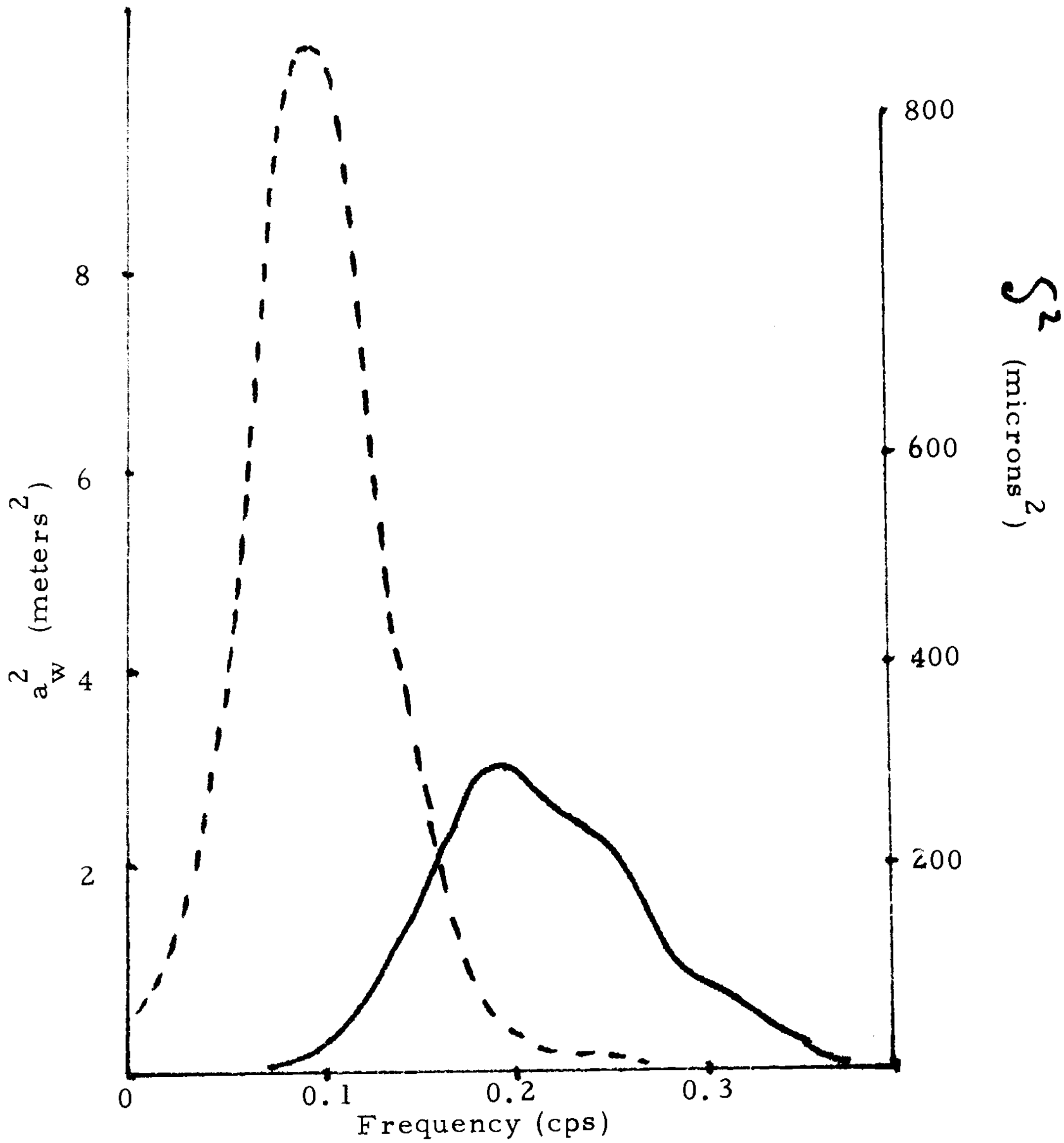


Figure 4.3-2. Power spectra as computed from the sea wave (a) and microseismic (b) records (obs. no. 2).

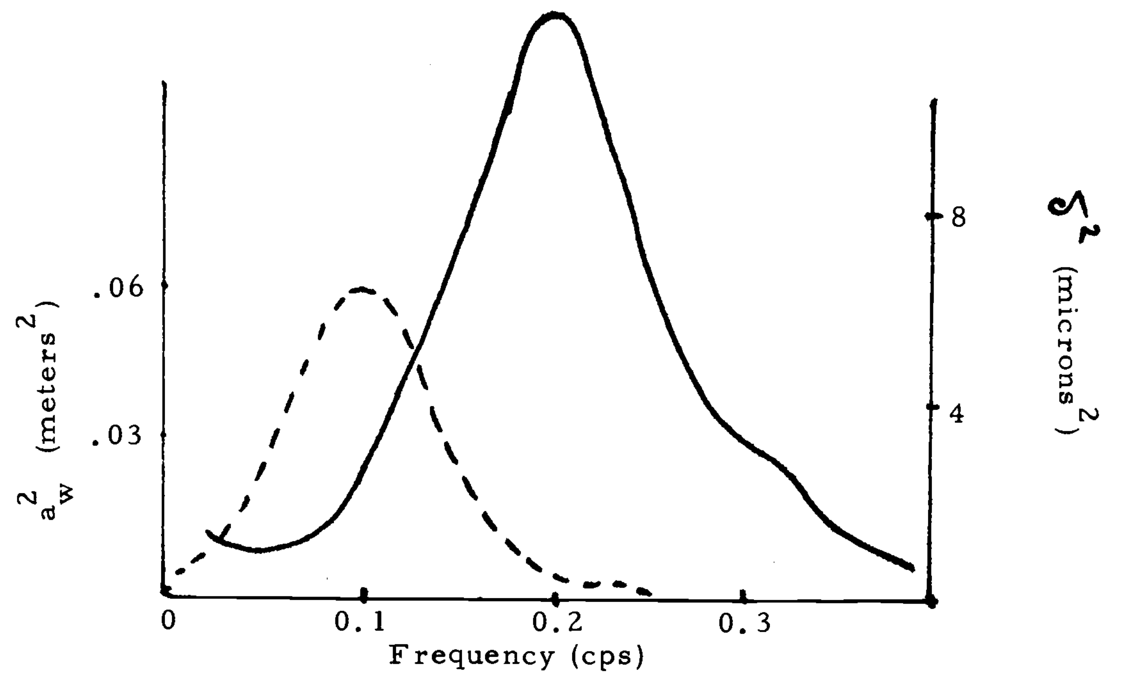


Figure 4.3-3. Power spectra as computed from the sea wave (a) and microseismic (b) records (obs. no. 3).

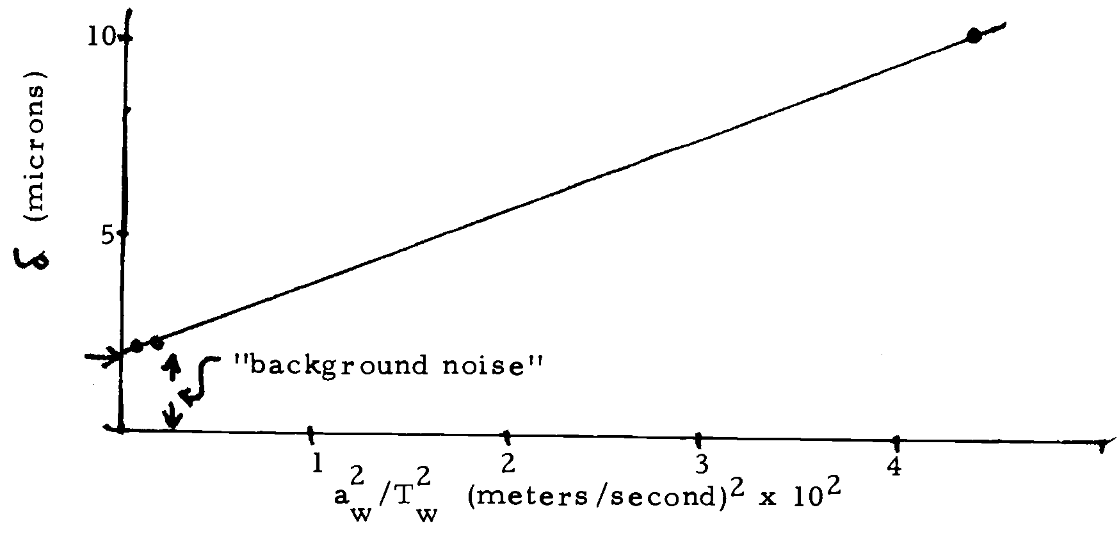


Figure 4.3-4. Rms vertical ground displacement vs. the ratio of the rms sea-wave amplitudes to predominant period.

0.2 cps for the microseisms, which correspond to periods of 10 and 5 seconds, respectively. A ratio close to 2:1 was maintained between the microseisms and the sea-wave peak frequencies in all of the observations, confirming that the generation of the microseisms occurs by the standing-wave mechanism proposed by Longuet-Higgins (1950).

The three points obtained in the plot, shown in Figure 4.3-4, of the (rms) microseism amplitudes versus the squared ratios of the corresponding (rms) sea wave amplitudes to periods (a_w^2 / T_w^2) lie practically on a straight line, which appears to intersect a "background noise" level on the vertical axis of about 2 microns. Although a sample of three points is obviously too small to be conclusive, this result is in accord with earlier observations, made by Zopf (1972) (see Figure 1.4-1), which were briefly discussed in section 1.4.

The estimated values of ground-motion amplitudes, which should theoretically be generated by the hypothetical, long-shore standing-wave system described above, are shown in Figures 4.3-5 to 4.3-7. Each of the curves are computations based on different values of the reflection coefficients (R), while the shear wave velocity β is the same in each figure. The observed (i.e., determined from data) microseism (rms) amplitudes are represented by the circled dots. There appears to be good agreement between the computed and the measured amplitude values, as indicated by most of the results shown

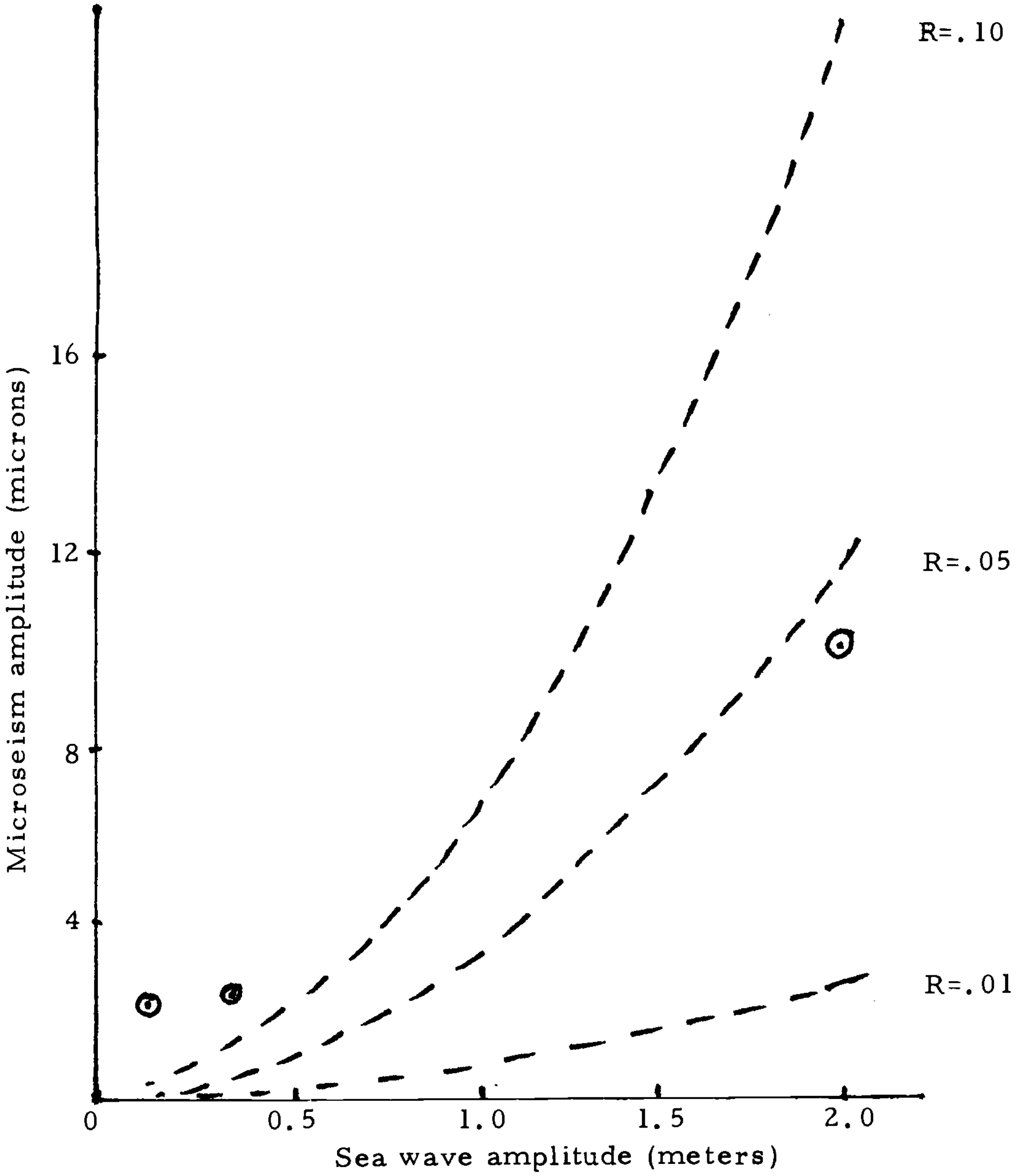


Figure 4.3-5. Theoretical (curves) and observed ground displacement as function of incoming sea wave amplitude, and different values of R.
($\sqrt{\Delta A}$ = one 10-sec. wavelenvth, β_2 = 2.0 km/sec.)

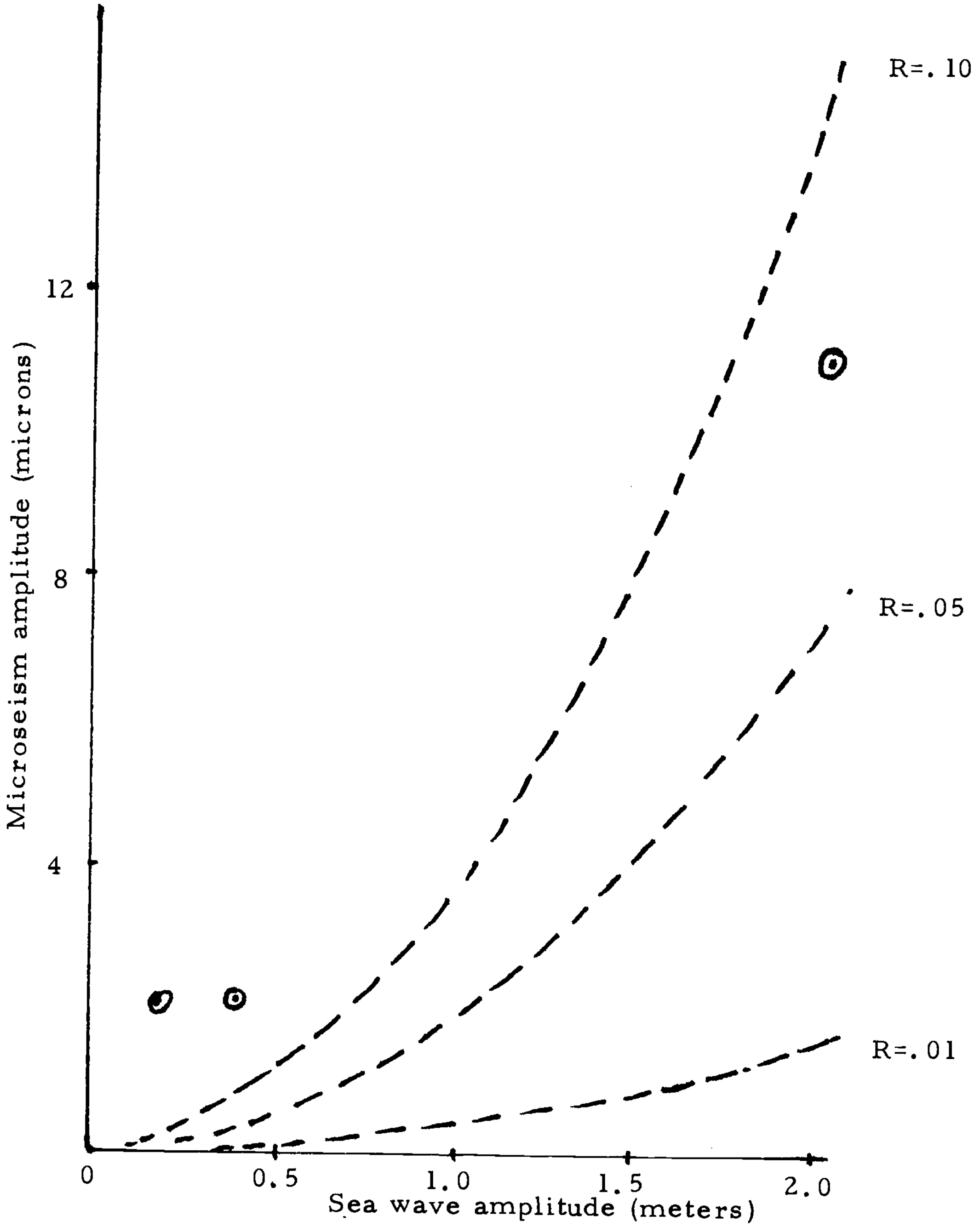


Figure 4.3-6. Theoretical (curves) and observed ground displacement as function of incoming sea wave amplitude and different values of R. ($\sqrt{\Delta A}$ = one 10-sec. wavelength, β_2 = 2.5 km/sec.)

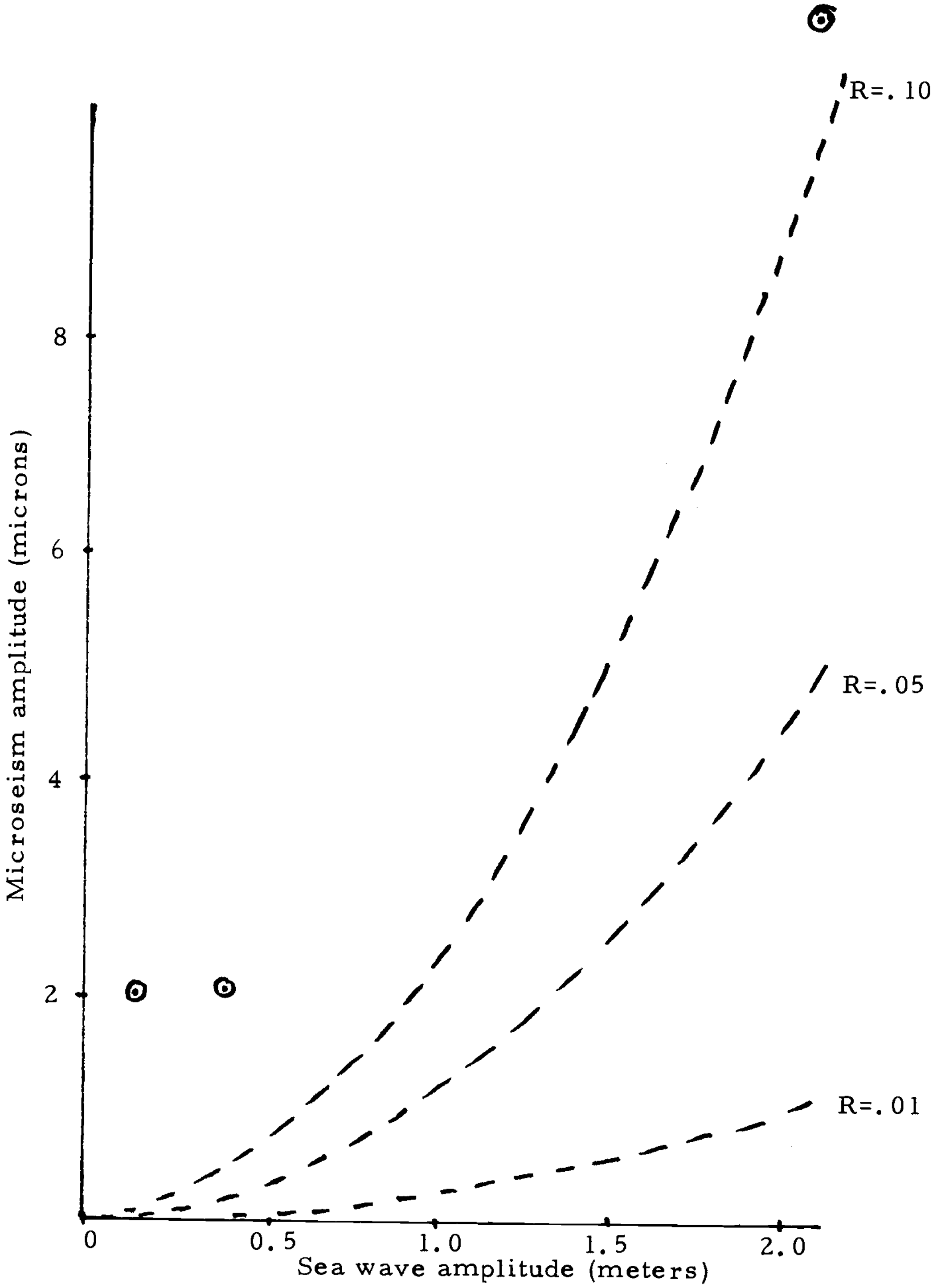


Figure 4.3-7. Theoretical (curves) and observed ground (⊙) displacement as function of incoming sea wave amplitude and different values of R.
($\sqrt{\Delta A}$ = one 10-sec. wavelength, β_2 = 3.0 km/sec.)

in Figures 4.3-5 to 4.3-7. This shows that the Oregon coastal microseismic activity may easily be accounted for in terms of generation by standing water-waves off the shore, which are inferred to be formed by the interaction between incoming and reflected sea waves, when small reflection coefficients are assumed for the waves at the coast.

The range of values assumed for the reflection coefficients (0.01 to 0.1) in the calculations above was considered reasonable for the Oregon coast, which may be regarded as fairly typical of most beaches. (A value of 0.05 was cited by Longuet-Higgins, 1950, in a numerical example of a hypothetical case of microseism generation by wave reflection at a continental shore, as a value representative for most beaches.) The reflection coefficient has to be inferred or assumed, since the actual amount of wave reflection at the shore is unknown and would be difficult to measure. The Oregon coast, however, has a slope of about 0.01, according to the east-west depth profile shown in Figure 4.3-8, and laboratory experiments at Oregon State University had indicated, for comparison, that as much as 10% reflection of water waves takes place off a 1:12-slope tank wall (Nath, personal communication).

The narrow width of 400 meters of the hypothetical standing-wave region (standing-wave activity outside of this limit was neglected) was chosen, for convenience, to allow for the influence

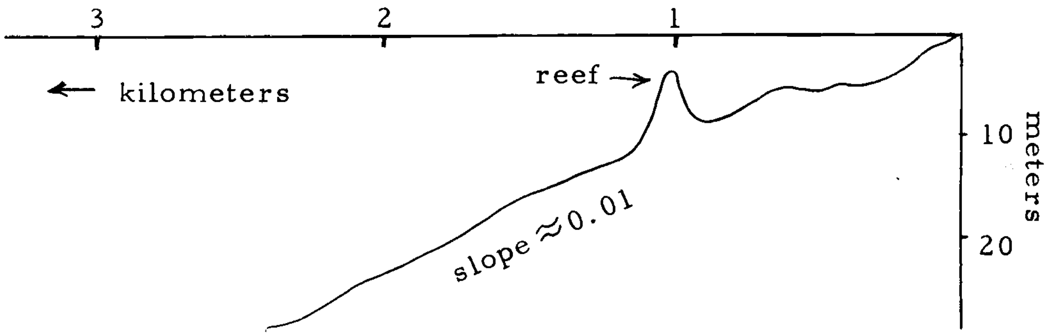


Figure 4. 3-8. Offshore depth profile (B-W) through Newport, Oregon.

which the deep-water, incoming-wave directions should have on the formation of standing wave systems and the microseismic activity. Waves may approach the shore from several directions simultaneously, and it is likely that some mean angle of approach at any given time would be oblique rather than perpendicular to the shore. Winds in this region tend, moreover, to be mainly from the southwest during that time of the year (fall), according to recent Oregon coastal (CUEA) observations (1971). It is considered likely that most of the standing-wave activity would be in the shallow-water regions close to the shore, where the wave crests, owing to refraction, are most nearly perpendicular to the shoreline. Wave refractions would also tend to increase the in-phase motion of the waves in the long-shore direction.

Spectra of the non-linear pressure variations, which were constructed from the sea-wave spectra based on the measurements, and the corresponding microseismic spectra and transfer functions are shown in Figures 4.3-9 to 4.3-16. The non-linear pressure spectra (dashed curves (b)) are qualitatively similar to the microseism spectra, as expected when assuming that they represent the generating forces of the microseisms. The transfer functions show generally broad peaks over the frequency (2 cps) of the maximum excitation, which is essentially an indication of the spectral energy distribution of the forcing functions. However, the transfer functions shown are only very approximate, because of the crudeness of the second-order pressure estimates, as well as the apparent presence of "background noise" in the microseisms (see Figures 4.3-4). They become therefore meaningless in the ranges involving very small sea-wave heights, as evidenced by the disagreement in the magnitudes of the transfer functions, as may be seen in Figures 4.3-9 to 4.3-16. The transfer-function estimates shown in Figure 4.3-14 correspond to the record pair showing the smallest observed sea-wave heights (obs. 3), and it appears to be several times greater than the transfer functions shown in Figures 4.3-10 and 4.3-12, which were derived from the other two record pairs (obs. 1 and 2), even when the "background noise" had been largely eliminated from the microseismic spectra (transfer functions, curves (b)). These results would suggest that at least

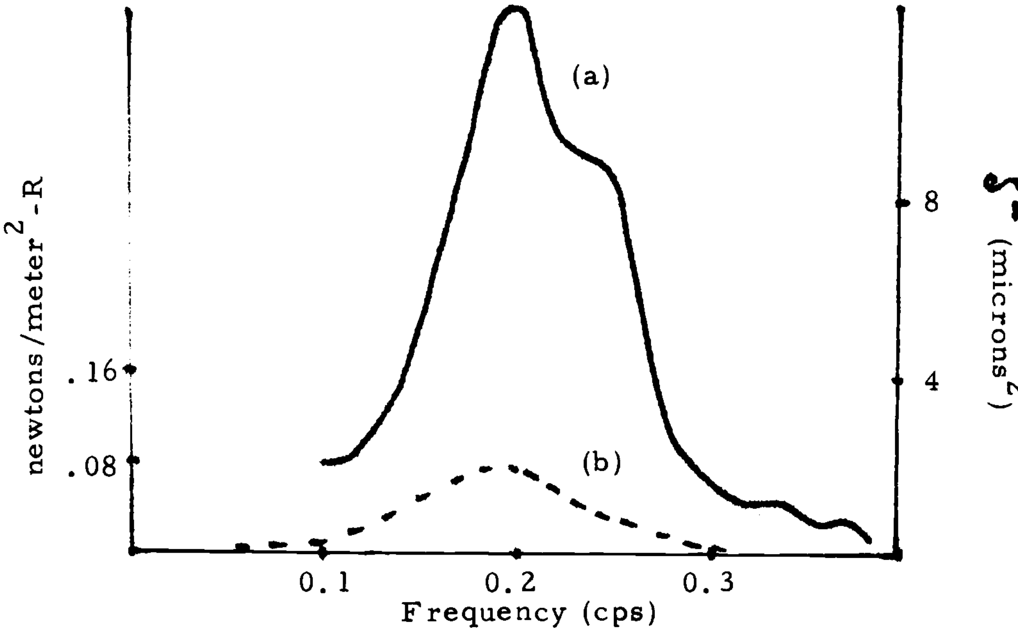


Figure 4.3-9. Power spectra of microseisms (a) and of second-order pressure variation (b), as inferred from the sea wave spectrum of obs. no. 1.

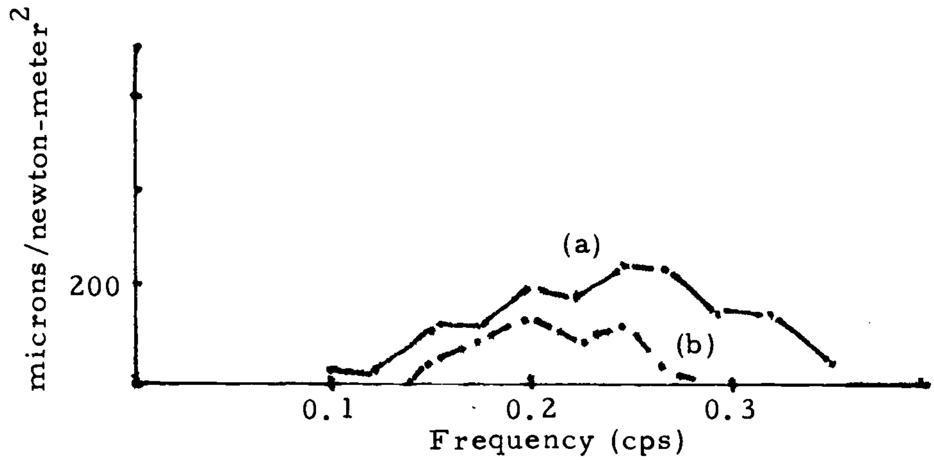


Figure 4.3-10. Transfer functions for the second-order sea wave pressure spe microseisms, as inferred from the spectra of obs. no. 1. Curve (a): including "background noise"; (b): without "background noise."

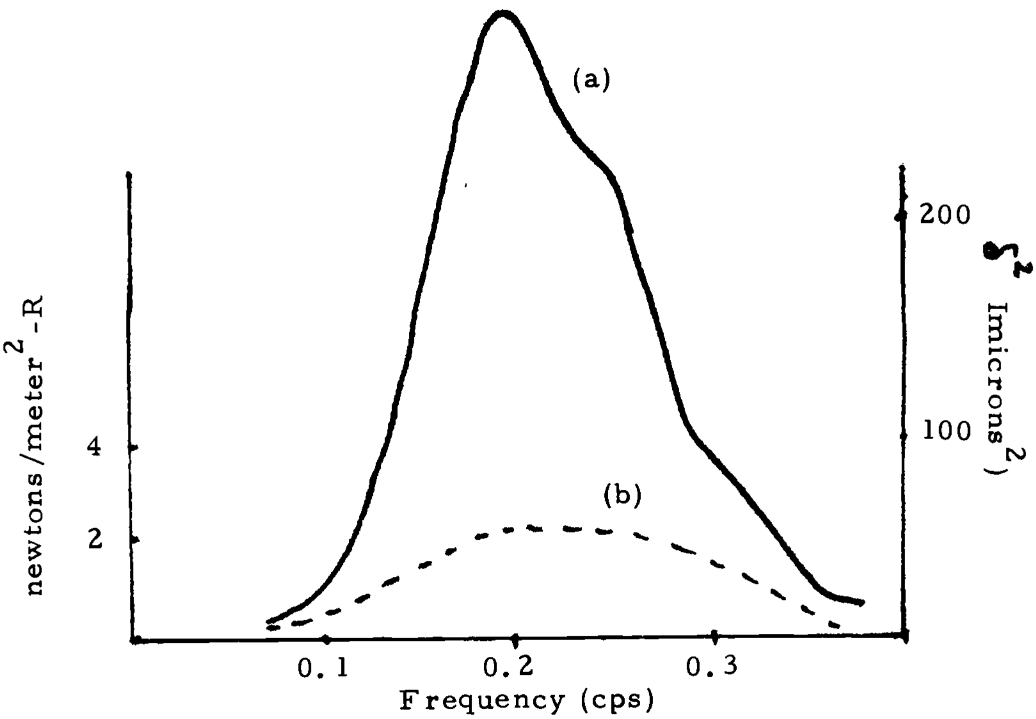


Figure 4.3-11. Power spectra of microseisms (a) and of second-order pressure variations (b), as inferred from the sea wave spectrum of obs. no. 2.

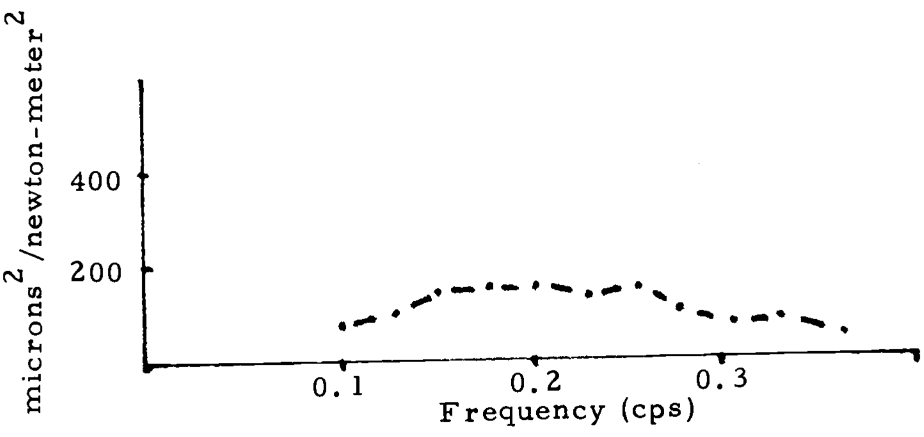


Figure 4.3-12. Transfer functions for the second-order sea wave pressures and microseisms, as inferred from the spectra of obs. no. 2.

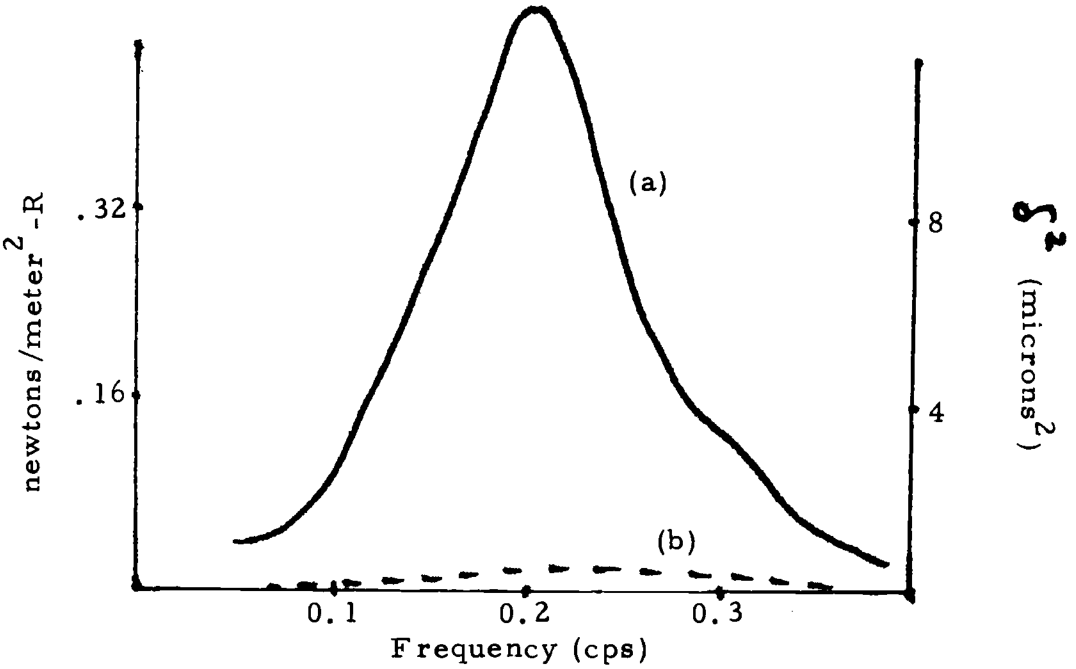


Figure 4.3-13. Power spectra of microseisms (a) and of second-order pressure variations (b), as inferred from the sea wave spectrum of obs. no. 3.

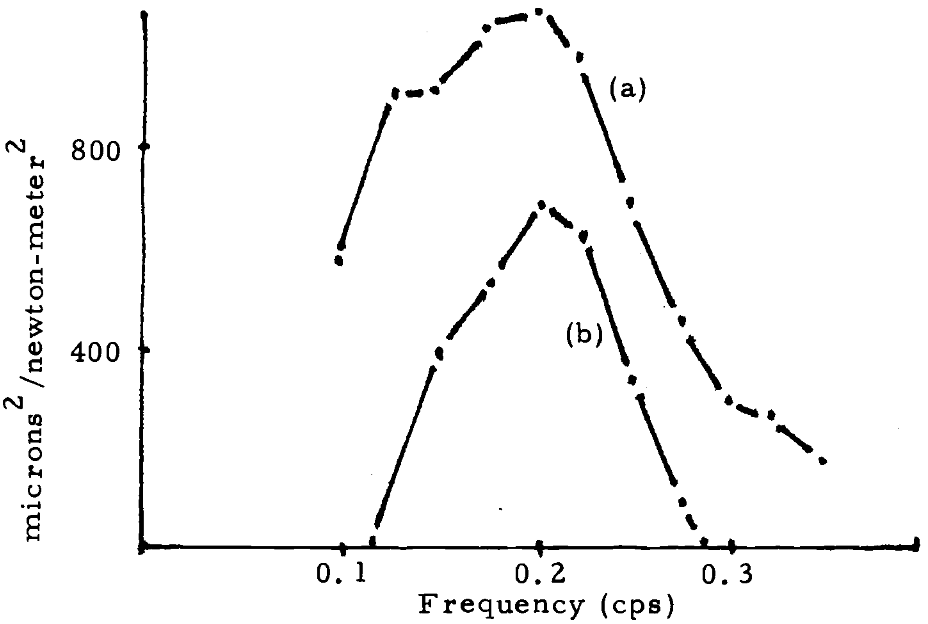


Figure 4.3-14. Transfer functions for the second-order sea wave pressures and microseisms, as inferred from the spectra of obs. no. 3. Curve (a): including "background noise"; (b) without "background noise."

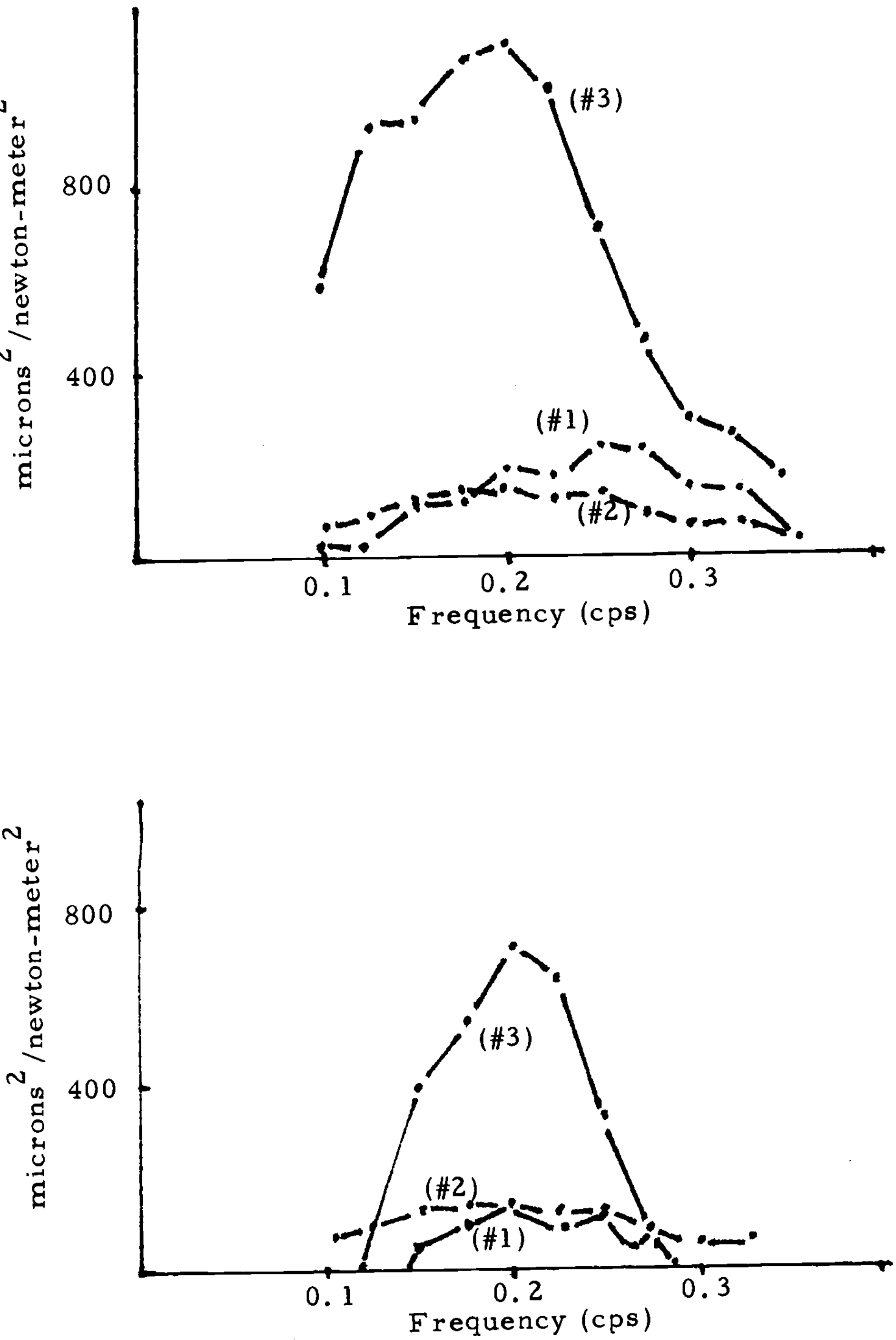


Figure 4.3-16. Transfer functions, based on all three record pairs (obs. no. 1-3), corrected for "background noise."

some of the microseismic activity is caused by sources or sea waves not directly represented by those measured at the present coastal site off Newport, Oregon, and possibly also that there is an influence on the microseismic activity by variations in offshore wave-direction, which is not related to the heights and periods of the waves observed. The submerged reef (shown previously in both Figures 4.1-1 and 4.1-8) lying parallel to the shoreline about one kilometer off the shore, and the north jetty (Figure 4.1-1), may possibly act as water-wave reflectors in addition to the shore proper, and thereby contribute to some extent to the microseismic activity in the area. The reef is approximately four kilometers long, and the minimum water depth over it is about four meters. The jetties extend southwest into the sea to a distance of about half a kilometer from the shore. The reflectivities of the reef and the jetty are unknown; water-wave reflection by these two structures is undoubtedly considerably greater than that of the beach itself, but, because of their relatively small dimensions, as compared with several hundreds of kilometers of the coastline, their contribution to the microseisms is, in spite of their nearby location, unlikely to be significant. Assuming shore reflections in the orders considered (0.01 to 0.1), the Oregon coast as a whole should overwhelm the few kilometers coastal segment near Newport, regardless of the reflectivities there. In fact, similar calculations as in example 3.1-14, section 3.1, seem to indicate that microseisms, due to reflection off the reef (even for an unrealistic maximum reflection coefficient of unity there), would be smaller by a couple of orders of magnitude than those observed. This could, however, be better determined, of course, by making similar measurements at other points along the coast.

A very small microseismic component, characterized by periods equal to those of the sea waves, was detected in only one spectrum; the one shown in Figure 4.3-17. The microseismic and sea wave spectra in Figure 4.3-17 were computed from the same set of observations (No. 2) as the spectra in Figure 4.3-2, only with a higher resolution. The sea wave spectrum shows the presence of at least two simultaneous swell groups, both apparently causing small ground oscillations at the same periods of 10 and 15 seconds, in addition to the more dominant oscillations characterized by half those periods. These very minor microseismic components, which appear to be smaller than the "double-frequency" components by factors of about 10 to 100, are assumed caused by the shallow-water, wave-modulation mechanism, proposed by Hasselmann (1963), and possibly also by waves breaking on the beach. The former (Hasselmann's) mechanism is strongly favorable to the lower-frequency waves, and its presence seems to be suggested by the relative prominence of the energy peaks in the microseismic spectrum in Figure 4.3-17.

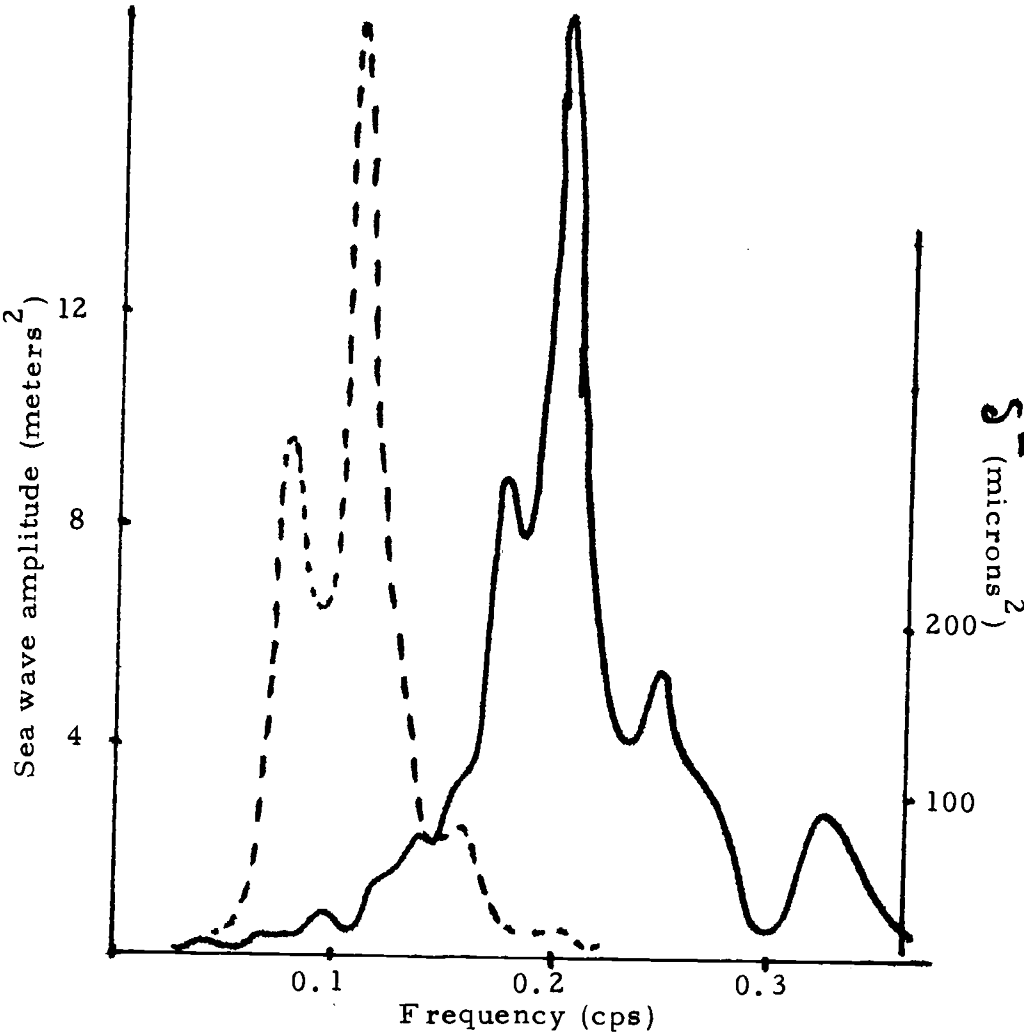


Figure 4.3-17. Power spectra for observations no. 2. Solid curve: microseisms; dashed curve: sea waves (higher resolution)

4.4 Conclusions

Results of spectral analysis of simultaneous sea wave and microseisms observations at the Oregon coast (Newport, Oregon) show that the local sea waves are associated with small ground oscillation whose mean or predominant periods are very nearly half those of the sea waves. This confirms that the microseisms are generated by standing water-waves off the shore, according to the theory of Longuet-Higgins (1950). The standing-wave activity is inferred to occur as a result of interaction between incoming waves and waves reflected from the shore, and the microseismic activity may easily be accounted for by assuming small coastal reflection coefficients in the order of one to ten parts per hundred. It is believed that the standing-wave action takes place mainly in the shallow-water regions along the coast. There seems to be relatively little seismic noise present due to other sources or mechanisms of generation.

Observations which have been made to date show that there may be, and probably is, a very approximate linear relationship between the magnitude of the seismic oscillations and the force per unit area exerted by the second-order pressure variations associated with standing waves off the coast, although the data are not conclusive. The magnitude of the second-order pressure variations is directly proportional to the squared product of the wave heights and frequencies, which may be measured with a wave sensor. Microseisms may thus for practical purposes be employed for determining significant heights and predominant periods of sea waves off the shore.

4.5 Comments on Coastal Reflection, Microseism Generation and Seismic Measurement of the Sea State

The basic purpose of investigating the relationship between the microseisms and local sea waves was the possibility of using seismic recordings for determining the offshore sea state. The unreliability of the conventional methods for measuring sea waves had prompted the use of a seissmometer, which was used with apparent success during the last few years at Oregon State University for wave prediction verification (see Enfield, 1974). The amplitudes and periods of the local sea waves showed a strong relationship with those of the microseisms, although this relationship has not yet been thoroughly investigated quantitatively, owing to lack of sufficient data. In order to evaluate the usefulness of the seismic method more completely, it would be necessary to make a greater number of similar measurements over a period of time, covering preferably the greater part of a year, in order to include as many different sea states as possible. Observation of predominant directions of propagation of the waves offshore would be desirable. In addition, measurements could be made at other sites along the Oregon coast to see whether the microseismic activity observed at Newport, Oregon, represents the effects of a process going on along the entire shoreline, or whether local effects are involved, such as wave reflection off the reef or jetties.

It has been implied throughout this thesis that coastal wave reflection is responsible for standing-wave activity along the shore, and hence for the microseisms observed in the area. Although this is most probably the case, it is strictly only an assumption, since nothing is known about the actual amount, if any, of wave reflection occurring at the coast. Reflected waves would inevitably be quite small and difficult to measure. Simple reflection of sea waves at a fairly straight shore, such as that of the Oregon coast, would imply, or strongly suggest, an influence of the incoming, deep-water wave directions on the microseisms, with obvious implications on using seismic methods of measuring sea wave heights. The scatter in Zopf's plot of seismometer-inferred sea wave heights versus wave-sensor and visually observed heights (see Figure 2.4-1) may well be due to variations in the direction of the waves. In general, however, the relation between standing-wave activity, microseism generation and offshore wave direction is not a simple one. It depends also on the configuration of the shore, on the geography, as well as the topography.

In addition to regular reflection off straight and smooth surfaces, it is also possible that wave energy may be scattered in various directions by small-scale undulations of the nearshore sea bed. The possibilities of energy transfer between generalized wave-motions (mainly long surface and internal waves) in the ocean by sea-floor irregularities and orbital currents was studied from a theoretical point of view by Cox and Sandstrom (1962). It was found that energy transfer may occur from one mode to another for waves of the same frequencies, if the wave numbers of the sea floor variations are in the same order of magnitude as the difference between the wave numbers of the two coupled modes. Cox and Sandstrom concluded that 9 to 69 percent of the power of the surface tide could be converted to internal waves by the irregular bottom of the Atlantic Ocean. These results were based on solutions developed for the inhomogeneous equations of wave motions (the inhomogeneity being due to the variable water-depth), by means of a perturbation approach. It may also be possible to arrive at a theoretical estimate for a reflection coefficient for the shore by the same method (Bodvarsson, personal communication), when the near-shore bottom topography is known in some detail. Such an approach has not been made for gravity waves in a coastal region and it is not known what orders of magnitude would be obtained for wave reflection coefficients or wave scattering at a shore like that on the Oregon coast.

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