Infragravity Waves in the Deep Ocean

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Energetic pressure fluctuations at periods longer than 30 s are a ubiquitous feature of pressure spectra from instruments sited on the deep seafloor in both the Atlantic and the Pacific oceans. We show these pressure fluctuations are caused by freely propagating ocean surface waves. The waves are generated in the near shore region along the entire coastline of an ocean basin through nonlinear transfer of energy from short-period waves. This view contrasts with some earlier work, which described these long-period pressure fluctuations as trapped waves tied to groups of short waves. We have constructed a model based on the average energy in the short (wind driven and swell) wave band along the North Atlantic coast to predict the energy in the long wave band at a site in the Atlantic. Maximum likelihood wave number-frequency spectra calculated on data from an 11 element array in the North Pacific confirm that the long wave energy is confined to wave numbers corresponding to the surface gravity wave dispersion relation. We have used the wave number spectra to isolate particular regions of the Pacific Ocean which are sources of long wave energy. Energetic short-period waves are incident on the coastline in these regions. Long waves are detected at the array which originate in the Gulf of Alaska, the northwestern Pacific, and at the southern tip of South America.

INTRODUCTION

The energy in the ocean wave height spectrum falls into a deep trough at frequencies between the tides at 2 cycles per day and ocean swell at frequencies higher than about 0.04 Hz. The term "infragravity wave" has been used to describe these surface waves at periods longer than the swell and wind-driven waves. In the surf zone, infragravity waves (also described as "swash" or "surfbeat") can contribute as much as 50% of the variance of sea level from surface waves. For this reason many measurements of infragravity waves in the near-shore region have been made beginning with *Munk* [1949] and *Tucker* [1950].

We focus on the infragravity wave spectrum as measured by instruments on the deep seafloor. The pressure signal from infragravity waves is easily detected using differential pressure gauges and is very evident at frequencies below 0.03 Hz in spectra from sites in both the Atlantic and the Pacific (Figures 1 and 2). Since these waves are surface waves, the pressure signal decays away from the sea surface exponentially with some characteristic length. Waves shorter than 30 s period are too greatly attenuated to be seen at the deep seafloor. Wave heights in the infragravity wave band inferred from the pressure spectra are very small (1 cm or less) when compared to typical wind waves (a few meters) or deep-ocean tides (30 cm). The very small amplitudes associated with deep-sea infragravity suggest the origins and physics of infragravity waves are only of esoteric interest. However, the deformation of the seafloor under infragravity waves establishes a lower bound on the amplitudes of seismic signals that can be detected at the seafloor and under the seafloor in boreholes. The resurgence of interest in establishing permanent seismic stations on the seafloor [Purdy and Dziewonski, 1988] provides a motivation to study infragravity waves to enable one to predict infragravity wave amplitudes at various sites in the oceans.

Our interest results from the realization that the deformation under deep-ocean infragravity waves provides a new method to

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study the shallow elastic structure of the oceanic crust. The infragravity waves induce deformations of the seafloor that can be measured using a low-noise, long-period seismometer such as the LaCoste-Romberg gravimeter [Webb and Hildebrand, 1988]. One major goal of the research described here was to show that the low-frequency pressure signals detected at the deep seafloor can be identified with freely propagating surface gravity waves so that the wavelengths of the disturbances can be determined as a function of frequency from the surface gravity wave dispersion relation. Measurements of the transfer function between vertical acceleration and the pressure as a function of frequency from a site on Axial Volcano, Juan deFuca Ridge, have been inverted to provide constraints on the shallow elastic structure under this volcano [Webb and Hildebrand, 1988]. This technique was pioneered by Yamamoto and Torii [1986], who showed that seafloor displacement and pressure measurements under ocean waves in shallow water could be used to determine the shear modulus in the upper few tens of meters of sediment.

Sutton and Barstow [1990] have recently reanalyzed longperiod pressure and displacement signals obtained from the OBSS system installed on the seafloor off Pt. Arena in 1965. Spectra of pressure and seafloor displacement from this site are also consistent with infragravity waves as the source. Infragravity wave energy at the site was found to vary by nearly 20 dB when storms were present in the North Pacific. Sutton et al. [1965] had previously linked low-frequency pressure signals detected by the OBSS system to forced waves traveling with groups of short waves because the amplitude of the long-period signals was correlated with short-period wave height near the site. Larsen [1978] has predicted the amplitude of the forced long wave component.

Our results suggest forced waves are not a significant component of the pressure signal. In this paper, we will discuss the sources of deep-sea infragravity waves using two data sets, one from the North Atlantic, and one from the North Pacific. We believe deep-water infragravity waves are primarily generated in shallow water, by the same nonlinear mechanisms that drive edge waves in the near shore [Gallagher, 1971; Symonds et al., 1982]. Waves detected in the deep ocean must propagate off the shelf in "leaky" modes [Munk et al., 1964] or escape by scattering from an irregular shelf or coast. The wavelength of a freely propagating wave will be distinctly different from a forced or trapped wave. Understanding the role of the free versus trapped waves is crucial



Fig. 1. Power spectra of pressure fluctuations from the sea floor at the HEBBLE site: (a) an interval of moderate current at the seafloor (5 cm/s), and (b) an interval of unusually strong seafloor currents (30 cm/s).

to predicting the energy in the infragravity wave band from environmental measurements of wind and waves and in the use of bottom deformations to infer the elastic strength of the seabed.

A 70-day record of pressure fluctuations from the western Atlantic (Figure 3) will be used to illustrate the day to day variability in energy in the infragravity wave band. A single instrunent collected data from late September to mid-December of 1985 and provides measurements over a frequency band from 0.003 to 8 Hz. The energy varies by about a factor of 40 during the interval spanned by the Atlantic record. The energy in the infragravity band detected at this site can be related to wind waves in the near shore region along the perimeter of the North Atlantic. In general, the generation mechanisms and the propagation problem are sufficiently complex to preclude direct calculations of the deep-water infragravity energy; however, we will present a simple model directed at predicting infragravity energy in the North Atlantic. The method is based on calculations of the average short wave energy incident along the coastlines of the North Atlantic. The model provides an explanation as to why the energy in the infragravity band from this site in the western Atlantic is always 20 to 30 dB less than at sites in the eastern Pacific. Wave height estimates were derived from the global spectral ocean wave model [Clancy et al., 1986] of the Fleet Numerical Oceanography Center. Hindcasts from this wave model have been used to locate possible sources, and to examine correlations between the (short) wave height predicted above the instrument site and at other sites in the North Atlantic and the pressure signal seen at the seafloor.

The Pacific array provides a more powerful tool for locating sources and sinks of infragravity wave energy. The Pacific array consisted of eleven instruments deployed and recovered during a three week period in November, 1988 at a site 1000 km West of Point Conception, California (Figure 4). The results from beamforming on the Pacific array using maximum likelihood techniques illustrate that the energy is indeed concentrated on the surface gravity wave dispersion curve. The infragravity waves are primarily from a few sources in localized directions.



Fig. 2. A typical pressure power spectrum from the seafloor in the eastern Pacific Ocean. This spectrum is from the Galapagos ridge.

PREVIOUS WORK

Measurements of long-period surface gravity waves began in the late 1940s, when *Munk* [1949] and *Tucker* [1950] independently discovered small-amplitude, low-frequency oscillations in sea surface height near the shore. Tucker found a correlation between the amplitude of incident short waves and the amplitude of the long-period waves observed offshore. Measurements from within a few kilometers of shore show peaks and valleys in the spectra of wave height identified by *Munk* [1960] as related to trapped "shelf waves". *Munk et al.* [1964] measured the along shore wave number spectrum of these modes and provided a theory including both trapped and "leaky" modes on a straight and uniform shelf and shoreline.

Work on infragravity waves has focused on tsunamis and on waves in shallow water, since these are the only infragravity waves sufficiently energetic to affect the world at large. Tsunamis are large amplitude infragravity waves generated by the displacement of the seafloor during earthquakes, or by landslides. The work on tsunamis has focused on three problems: the generation by an earthquake source, the bathymetric steering of the waves and their dissipation. The tsunami work has shown that long wave dissipation is negligible in deep-water, and also that waves are greatly attenuated during interaction (reflection) with the coastline [VanDorn, 1987].

The first measurements from the deep seafloor of the pressure fluctuations from infragravity waves were made by *Sutton et al.* [1965] using a very large ceramic hydrophone with a pass band from about 0.005 Hz to 0.1 Hz. They interpretated their observations as the result of forced waves associated with groups of higher frequency surface waves. *Filloux* [1980] measured the spectrum of pressure fluctuations below 0.03 Hz at a site in the deep sea near Hawaii using a Bourdon tube gauge and discovered



Fig. 3. The HEBBLE site. The dark line connecting the HEBBLE site and the coast of Portugal is a great circle route. The light dashed line is the 100 m contour.



Fig. 4. (Right) The PEGASUS experiment site. (Left) The PEGASUS array.

the pronounced rise in spectral levels at frequencies below 0.01 Hz. *Filloux* [1983] reports on observations from other sites and describes a tsunami detected with the pressure sensor on the seafloor. Filloux noted the shape of the spectrum matched that expected from the hydrodynamic filtering of free surface waves, and identified these low frequency pressure fluctuations as caused by surface waves. The Bourdon tube measurements have continued during a series of recent experiments directed at measuring ocean currents and tides [*Filloux et al.*, 1990; *Luther et al.*, 1990]. A program has been recently established by NOAA [*Gonzalez et al.*, 1987] to monitor tsunamis near the Aleutian Islands using several continuously occupied deep sea stations equipped with quartz crystal pressure sensors.

Infragravity waves have been measured at the sea surface in the open ocean only once. *Snodgrass et al.* [1966] describe observations of the propagation of ocean swell from large storms from the southern oceans to Alaska. The floating instrument platform (FLIP) was instrumented with pressure gauges and accelerometers to measure low-frequency waves. FLIP is essentially a large manned spar buoy and the response of FLIP to low-frequency ocean waves is sufficiently well known that an estimate of the infragravity wave spectrum in the band above 0.005 Hz could be derived from measurements of its vertical acceleration. The energy in the infragravity wave band over the open ocean in the Pacific was found to be very small (less than 1 cm in amplitude).

The long wave amplitude measured from FLIP is 7 to 25 times larger than wave heights inferred from the pressure measurements from the floor of the Atlantic (0.3 to 1.2 mm). Seafloor measurements of pressure fluctuations from several Pacific sites confirm the infragravity wave spectrum is consistently more energetic in the Pacific than at this Atlantic site [Webb and Cox, 1986]. We will show later that this may be a result of the location of the North Atlantic site on the Nova Scotia Rise. The instrument may ite "in the lee" of the Grand Banks of Newfoundland and shielded from infragravity waves propagating from the northern seas.

Infragravity waves may also exist on the Arctic Ocean. Gravimeter measurements of the vertical displacements of ice islands show amplitudes of about 0.5 mm in the period range from 5 to 40 s [*Crary et al.*, 1952]. There is evidence of propagation of these motions at surface wave velocities; however, displacement amplitudes are strongly correlated with the wind, suggesting the infragravity waves are generated locally or the motions are caused by direct forcing of the ice islands [*Hunkins*, 1962].

DEEP-OCEAN PRESSURE SPECTRA

A differential pressure gauge was developed by *Cox et al.* [1984] for pressure measurements of low frequency acoustic signals. The gauge was found to be useful in a band from about 0.0005 to 32 Hz. The gauge was designed to avoid the difficulties at long-periods with conventional hydrophones caused by increasing drift from electronic noise and temperature fluctuations. The gauge has been deployed during a series of experiments in the eastern Pacific [*Webb and Cox*, 1986]. Most estimates of the prescure spectrum from these experiments are remarkably similar in appearance (Figures 1 and 2).

Energy in the pressure spectrum above 0.08 Hz is associated with acoustic waves. Above 15 Hz, noise is dominated by shipping and levels are quite variable. An energetic peak is always present in the pressure spectrum between 0.1 and 5.0 Hz and is called the microseism peak. Signals in this band propagate as elastic waves with energy partitioned between the ocean and seafloor rocks depending on wave mode and frequency. The source of the energy in this band has been shown to be derived by a nonlinear mechanism from surface waves [Kibblewhite and Ewans, 1985, Webb and Cox, 1986].

Below 0.1 Hz the spectrum falls abruptly (40 to 60 dB) into what has been termed the "noise notch". The ambient noise level between 0.03 and 0.1 Hz is typically very low and may lie at some times below the noise level set by the electronic noise of the differential gauges of about 10-3 Pa2/Hz. The signal level in this band is probably very low because hydrodynamic and acoustic scales are either too large (acoustic) or too small (hydrodynamic) for energy to be easily coupled into propagating waves. Usually, we see only Rayleigh waves from occasional large earthquakes in this band, although flow noise can be important at some locations including the Atlantic site [Webb, 1988]. The seafloor currents must be greater than about 5 cm/s to produce measurable effects. Most hydrodynamic signals are restricted to lower frequencies; for example the typical deep ocean buoyancy frequency is near one half cycle per hour, far below the frequencies considered here, so internal waves do not generate pressure disturbances in the deep sea in this band.

The energy in the deep-ocean pressure spectrum is much larger below 0.03 Hz, and this signal is caused by infragravity waves. The sharp increase in energy below 0.03 Hz can be easily understood by examining the physics of surface gravity waves. The amplitude of the pressure signal at the seafloor of depth h, is related to the surface wave height ζ by

$$P_b = \rho g \zeta / \cosh(kh) \approx P_0 e^{(-kh)}$$
(1)

where k is the wave number and ρ is the water density. Waves with wave numbers comparable or smaller than the inverse of the water depth $(kh \approx 1)$ generate pressure fluctuations which can be detected at the seafloor. The pressure signals from waves of slightly higher frequency are greatly attenuated because of this exponential decay with depth; we will refer to this effect as "hydrodynamic filtering". If the pressure fluctuations are caused by freely traveling surface gravity waves then the dispersion relation can be used to determine the wave number

$$\omega^2 = g k \tanh(kh) \tag{2}$$

where ω is the frequency of the wave in rad/s. If h is 5 km, then kh=1 for waves of 160 s period (0.006 Hz) and kh=5 for waves of 63 s period (0.015 Hz) in agreement with the observed spectral shape (Figure 1).

The pressure spectrum is flat or slightly humped down to frequencies of about 0.002 Hz (Figure 2). There is a sharp break in the slope of the spectrum at this frequency, and the spectrum rises slowly toward ever lower frequency until interrupted by the many lines associated with the tides [Filloux, 1980, 1983]. The mechanism generating the pressure fluctuations below 0.002 Hz may be different from the mechanisms discussed here for the infragravity wave band. Filloux [1983] has suggested that the pressure sensors are detecting the first baroclinic mode of internal waves, but D. S. Luther (personal communication, 1990) believes the measurements show very long-period surface waves driven directly by atmospheric pressure fluctuations. Recent measurements by Filloux et al. [1988] show very large correlation length scales at these low frequencies apparently ruling out a significant internal wave component to the pressure measurements. Pressure spectra measured at the Arctic seafloor under ice cover show similar amplitudes as the Pacific below 0.002 Hz, but much lesser amplitudes above 0.002 Hz, again suggesting a direct atmospheric origin for the very lowest frequency infragravity wave components [Schultz et al, 1990].

THE HEBBLE DEPLOYMENT

A small patch of the Nova Scotia Rise has been the site of a ten year, multidisciplinary experiment designed to study the effects of bottom currents on the geology and biology of the area. The experiment was given the acronym HEBBLE for "High Energy Benthic Boundary Layer Experiment". This experiment is described in a special volume of *Marine Geology* (1985, vol. 66). A deployment of several bottom lander tripods during the fall and winter of 1985/1986, provided an opportunity to obtain a long record of seafloor pressure fluctuations.

A tripod was deployed in 4817 m of water at the HEBBLE site on the Nova Scotia Rise (40°26.58'N, 62°21.57'W) on September 21, 1985, and recovered April 1986. A small cartridge tape drive on the tripod was used to record 83 days of pressure data starting September 22, 1985, and ending December 14, 1985. Battery and tape capacity limited the length of the record; some of the data in the early part of the record were unreadable because of problems with the tape recorder. The recording schedule was such that six contiguous, 1-hour-long records sampled at 1 Hz and one, 225-slong record sampled at 16 Hz were collected each day. The primary reason for this deployment was to study the effects of strong bottom currents on seafloor pressure measurements. Pressure spectra derived during periods of strong bottom currents show elevated levels in the noise notch [Webb, 1988]. We have now used this record to study the variation of energy in the infragravity wave band. The seafloor currents were too weak to effect pressure measurements in the more energetic infragravity wave band.

The energy in the infragravity wave band varies during the record by about a factor of 40, with a characteristic time scale of 6-10 days. Figure 5 displays the spectral density in the bands between 0.0039 Hz and 0.016 Hz versus Julian day. The jagged nature of the traces in Figure 5 is due to the irregular spacing of the spectral realizations and the statistical variability of the estimates. The shape of the infragravity wave spectrum is essentially invariant, so the energy in the 0.012 Hz band can be used to predict the energy in adjacent bands and as an indicator of total energy in the infragravity wave band. We have derived a measure of the wave height variance in the infragravity wave band by multiplying the spectral density at 0.012 Hz by a bandwidth of 0.03 Hz and then multiplying the result by 60 to correct for the hydrodynamic filtering to refer the pressure spectrum to the free surface (Figure 6a). We have used this one frequency rather than integrating over a band of frequencies because we fear some con-



Fig. 5. The pressure spectral density in four bands between 0.0039 and 0.0156 Hz between day 280 (October 8) and day 350 (December 15, 1984). The hydrodynamic filtering increases toward increasing frequency, so the amplitude of the pressure fluctuations is lower in the higher frequency bands.



Fig. 6(a). An estimate of the infragravity wave height variance (inferred from the pressure power spectral density in the band near 0.01172 Hz) versus day of the year showing the variability in the infragravity wave band energy with time. (b) The mean squared wind stress averaged over the surface of the North Atlantic during the same time interval as in Figure 6a. The infragravity wave energy is correlated with the wind stress. (c) The square of the significant wave height predicted from the FNOC wave model at the grid point nearest the HEBBLE site. Correlation between Figures 6a and 6c is poor. (d) Mean square wave height (wind wave or swell) averaged along all grid points nearest the coastlines which can be connected to the HEBBLE site by a great circle path without crossing the Grand Banks of Newfoundland.

tamination to the spectrum at longer periods due to electronic noise (a problem particular to only the Atlantic data set). One Pascal corresponds to a wave height of 0.1 mm.

We see three major peaks and several other smaller peaks during the record. The 6-10 day time scale of the variability is characteristic of major weather systems. We look for a correlation between the presence of large storms and the intensity of low frequency seafloor pressure fluctuations. A simple measure of the intensity of storms is the mean squared wind stress averaged over the North Atlantic (Figure 6b). The infragravity wave energy (Figure 6a) and the mean wind stress are surprisingly well correlated (correlation coefficient of 0.65, Figure 7), but the infragravity wave energy lags the mean wind stress by nearly 3 days. The origin of this time lag could depend on both the rate of growth of waves and some characteristic propagation time. It must be explained by our model.



Fig. 7. Time lag correlation of the mean squared wind stress averaged over the surface of the North Atlantic and the energy in the infragravity wave band. The two series are correlated at the 0.65 level and wind stress leads the infragravity wave energy by about 3 days.

Several storms can exist over the North Atlantic at the same time. We find that the presence of an intense storm is not necessarily correlated with more energetic low-frequency pressure fluctuations at our Nova Scotia Rise site. Figure 8a shows an example of a large storm (November 15; day 318) with large waves (>11 m) near the center. During this storm the infragravity wave height variance in the band below 0.04 Hz is estimated to be a moderate 0.25 mm². This and other wave height maps were derived from the Fleet Numerical Oceanography Center global spectral model results. Four days later (Figure 8b; November, 19; day 322), a hurricane North of Puerto Rico is associated with large low frequency pressure signals at the HEBBLE site (infragravity wave height variance 1.1 mm²). This pressure signal all but disappears the following day, after the hurricane leaves the Atlantic and enters the Caribbean Sea (Figure 8c; wave variance 0.2 mm²). We will look at two models: forced waves over the site and free waves derived from the coasts to explain these observations.

Fig. 8. Charts of significant wave height calculated by the FNOC model for (a) day 318, (b) day 322, and (c) day 323 in 1985. Contours are approximately in meters. Figure 8a shows an intense winter storm in the Norwegian-Greenland Sea. A smaller storm is located over the Grand Banks. A hurricane has generated large waves in the Caribbean Sea in Figure 8(b); the northern storm has moved much further north. In Figure 8(c) the hurricane has moved into the Gulf of Mexico.



FORCED WAVES

Larsen [1978] has proposed that the deep-sea infragravity wave signal is caused by forced waves associated with groups of shortperiod surface waves. The slight nonlinearity of the surface gravity wave boundary conditions requires the mean sea surface be depressed under the peaks of the envelope of a modulated wave train [Longuet-Higgins and Stewart, 1962]. An equivalent description is to say that a forced wave with a wavelength equal to the length of the wave group travels with the group, at the group velocity of the carrier wave with the trough of the forced wave phase locked to the peak of the envelope. The pressure signal associated with this forced wave is attenuated with depth into the ocean following equation (1), as is a free wave. Following Larsen, we assume $\varepsilon = ka$ is a small parameter describing the slope of the carrier wave $(k \text{ and } a \text{ are the wave number and wave ampli$ tude). We diverge from Larsen's notation to write the inverse length scale of the modulation of the wave group as l=k/n, where n is the number of cycles in a wave group. The amplitude of the pressure signal at the seafloor at depth h is

$$P \approx \frac{2\rho\varepsilon^2 C^2(l/k)}{\cosh(lh)} \tag{3}$$

with ρ the water density and C the phase velocity. The signal depends on the square of the wave slope since the nonlinearity is quadratic.

If the period of the short waves is 15 s and the amplitude is 6 m, then $\varepsilon = 0.1$. We take the wave group to be 10 waves (n = 10). Equation (3) predicts the amplitude of the pressure fluctuations at the floor in a 4.8 km deep ocean will be about 0.4 Pa (0.04 mm). This result is smaller than typical values for the root mean square wave height in the infragravity wave band determined from the Atlantic measurements. The forced wave moves with the group velocity of the short carrier waves. The group velocity of a 15-s wave is about 12 m/s. The period of the forced wave must be the length of the wave group divided by this group velocity. The group velocity of a deep-water surface wave is one half of the phase velocity; therefore the period of the forced wave is twice the period of the carrier wave times the number of waves in the group. The carrier wave is 15 s, and there are 10 waves to the group, so the forced wave period is 300 s, and the wavelength is 3450 m.

The wavelength of a forced wave is always much smaller than the wavelength of a free wave of the same frequency because the forced wave travels at the group velocity of the short, carrier wave, whereas the free wave travels at the much larger free wave phase velocity. We see now that forced waves cannot explain the sharp roll-off observed in the deep-water pressure spectra because a model based on forced waves must predict a much lower corner frequency than is observed. The corner frequency is well explained by hydrodynamic filtering if one assumes the wave number k is associated with freely propagating surface gravity waves.

We expect that if the infragravity wave band pressure signal were associated with forced waves then the amplitude of the fluctuations should be correlated with the (short) wave climate over the site. The significant wave height derived from the FNOC model at the grid point nearest the instrument site is shown in Figure 6c along with a measure of the infragravity wave energy near 0.01172 Hz in the pressure record (Figure 6a) from the HEBBLE experiment. The predicted wave height over the sensor varies greatly, but there is little correlation apparent between the wave

record overhead and the amplitude of the pressure fluctuations seen at the seafloor. A major peak in the infragravity wave spectrum (Figure 6) at 338 days is also present in the local wave data (although perhaps a day early), but the major peaks at 316 and 323 days are not seen in the local wave data. A major peak in the local wave height record at day 308 does not show up on the infragravity wave record. The duration of each peak in the local short wave record is shorter than the duration of the peaks in the infragravity wave energy, again suggesting the infragravity wave signal is not local in character. No correlation is apparent between the short waves overhead and the infragravity wave variance inferred from the bottom pressure spectrum in the scatter diagram of the two variables (Figure 9). The least squares best fit of a line to the data in this scatter plot has a slope of 0.34 and a correlation coefficient of 0.26 (fitting to the logarithm of the two variables).

We will show next that energetic intervals in the infragravity wave pressure record are well correlated with large short wave amplitudes along various coasts. The phase delay between the occurrence of waves along a coast and the energy in the pressure spectrum is too short to allow for the propagation of the short wave components to the HEBBLE site. This analysis suggests locally forced waves are not important in the band measured in this experiment. The measurements do not preclude the possibility that forced waves are an important component at periods longer than we observed; periods more comparable to the 300 s predicted from the simple example presented here.

Forced waves must be important on the shelf. The trapped wave corner frequency occurs at periods of only a few seconds in the measurements of *Treverrow et al.* [1989]. These authors describe array measurements of long waves in shallow water (12.5 m) off the coast of New Jersey. In these measurements infragravity waves were detected at periods from just above the swell (at 15 s) to periods as long as 90 s. Directional spectra calculated rom the array measurements showed the long waves at 90 s were propagating in the same direction as the swell at 15 s (both on-



Fig. 9. A scatter plot of the square of the significant wave height over the HEBBLE site and the infragravity wave energy measured at the HEBBLE site. The two series do not appear to be correlated.

shore), although not in the same direction as the very shortest waves (5–9 s). The water is too shallow around this array to discriminate between free and forced waves on the basis of phase velocity or wavelength of the waves, but *Treverrow et al.* [1989] relate these observations to forced waves under wave groups. Longuet-Higgins and Stewart predict the forced wave under a short wave group in shallow water should have amplitude

$$\zeta = -\frac{3ga^2}{2\omega^2 h^2} \tag{4}$$

The short wave height (a) from the published data is only about 0.1 m, in the band near 15-s period (ω =0.4s⁻¹). The forced wave amplitude in 12.5-m water depth (h) should be about 5 mm. The long wave amplitude in the infragravity wave band from these same measurements is about 5 mm. This estimate of long wave height is about 4 times larger than the largest rms wave height observed at the HEBBLE site, and about equal to the most energetic sites in the deep Pacific.

A MODEL FOR THE GENERATION OF INFRAGRAVITY WAVES ALONG COASTS

Surface waves become steeper and more nonlinear in shallow water, and so drive a larger forced wave component. The short waves break and dissipate at the shore, leaving the long wave components as low-frequency changes in the mean elevation [Longuet-Higgins and Stewart, 1964; Gallagher, 1971]. The long waves reflect at the shore line and couple primarily into waves trapped along the shoreline by the beach slope (edge waves), since the phase velocity of shallow water waves increases with water depth. The sloping beach can be described as a waveguide with many modes of different along-shore and cross-shore structure [Oltman-Shay and Guza, 1987]. Some energy escapes into deepwater in "leaky" modes [Munk et al., 1964].

A second mechanism for the origin of long-period variability of the mean elevation along a beach has been proposed by *Symonds et al.* [1982]. The variability in the height of crests in a wave group leads to a variation in the distance offshore of the break point. Bigger waves break further out, which leads to differences in the height of wave setup. For this discussion it is not important which mechanism generates the low frequency waves.

Most of the energy is trapped within a few hundred meters of shore in edge waves. A reviewer has pointed out the existence of a class of progressive waves on beaches which are too short to be edge waves [Oltman-Shay et al., 1989]. These vorticity balanced waves presumably do not couple into infragravity waves in deepwater. The edge wave amplitude has been shown to be linearly related to the amplitude of the incident short waves in the simple case of a narrow-band modulated wave train [Battjes, 1974]. Observations suggest a simple linear relationship may be adequate to predict long wave energy from swell height for broadband incident waves as well [Guza and Thornton, 1982]. Some dependence on the slope of the beach is expected from these theories, but we ignore this problem in the very simplified model we present below. The infragravity wave spectrum has been measured in the near shore region under limited ranges of conditions [Huntley et al., 1977; Guza and Thornton, 1982]. The spectrum was found to fall off at frequencies above about 0.03 Hz as ω^{-3} or ω^{-4} , and to be approximately constant at lower frequencies. We find these spectra are not inconsistent with deep-water results. Deep-ocean infragravity wave height spectra estimated from seafloor pressure measurements are essentially white (flat) in the band below 0.03 Hz after correcting for the effect of the hydrodynamic filtering. The spectrum of long waves at higher frequencies in the open ocean is unknown because the long waves are too strongly filtered to be detected at the floor in deep-water. These measurements contrast with the shelf measurements described by *Munk* [1960], which showed evidence of the modal structure in the frequency spectrum.

Having shown trapped waves are ineffective in deep-water, we now model the generation of deep-water infragravity waves using predictions of the short wave component (wind wave and swell) from the grid points in the FNOC wave model closest to shore along the entire coastline of the North Atlantic. We assume the infragravity waves are generated by waves in shallow water and the amplitude of the outgoing long waves is linearly related to the amplitude of the shorter incident waves. If the relationship is linear then the deep-water wave energy should be some weighted average of the short wave energy along the shore. We assume all shorelines are identical on average and have calculated the average variance in the significant wave height along the perimeter of the North Atlantic for each day during the HEBBLE site record (Figure 10). This average is corrected for the variation in grid spacing with latitude and longitude, but ignores the direction of propagation of the waves relative to the coast. The length of the coast line is roughly approximated by the choice of grid points. In this model, we have ignored the propagation delay from the coasts to the site, deep-water attenuation, and the South Atlantic as a source or sink of infragravity wave energy. More important, we have ignored dissipation on the shelf, and reflection of waves at coastlines.

The day to day variability of the energy in the infragravity wave band in the open ocean can be partly explained using this model. Most of the major peaks and valleys in the infragravity band energy (Figure 6a) also appear in this simple estimate of the strength of the forcing along the coast (Figure 10). The contribution from each coast varies depending on the location of various storms. The result from no single coast explains all the peaks seen in the pressure record.

The fit is improved when we average over only those sections of the coast which are within the "line of sight" of the Nova Scotia Site (Figures 6d and 10). This improvement is evident in a comparison of scatter diagrams of variance in the significant wave height averaged over the entire North Atlantic and over those sections in the line of site (Figures 11a and 11b). By "line of sight", we mean those sections which can be connected by a direct long wave ray path without intersecting another coastline or the shelf. We have used great circle paths to approximate the ray paths across the Atlantic. The great circle route from the Nova Scotia site to Europe, just tangent to the edge of the Grand Banks off of Newfoundland intersects the European continent as far south as Portugal (Figure 3). The great circle route only approximates the true path because the long wave phase velocity varies with the water depth. We find the line which fits the data in Figure 11a, has a slope of 1.8 and a correlation coefficient of 0.68. The best fitting line to the data in Figure 11b has a slope of 1.25 and a correlation coefficient of 0.71, a slight improvement, but the line fit to Figure 11a has a much larger intercept, a less satisfying iesult.

We determine the power law between the mean square wave height and the infragravity wave energy by fitting a line to the logarithm (base ten) of the variables. The best fitting line to the data in Figure 11*a* (in the least square sense) has a slope of 0.38, an intercept of 1.52, and a correlation coefficient of 0.66. The best fit line to the logarithm of the data in Figure 11*b* has a slope of 0.75, an intercept of 0.70, and a correlation coefficient of 0.76. Both



Fig. 10. The mean squared significant wave height averaged over different sections of coastline during the time interval spanned by the HEBBLE data. The darker of the dashed lines includes data from the entire coastline of the North Atlantic. The solid curve is the series derived from the grid points adjacent to sections of coastline which are not shielded by the Grand Banks from the HEBBLE site. The lighter of the two dashed lines included data only from the east coast of North America south of the Grand Banks. The models shown by the dashed lines are less well correlated with the energy in the infragravity wave band than are the data from the solid line.

estimates are a considerable improvement over a estimate based on a prediction for the short wave height over the sensor (Figure 6c, the correlation coefficient is 0.26). Because the line fitting procedure fits the logarithm of the data, the effect of the very energetic peaks is subdued, with the many small values determining the slope of the lines. The correlation coefficients generated by this method do provide some guidance, but we put little credence on the power coefficients generated by this method. We hypothesize a linear relationship, and instead plot the best fit line with a slope of one on Figure 11b.

The model derived from the "line of site" data fits the infragravity observations better than the model based on data which includes the northern ocean. These results suggest the infragravity spectrum at the Nova Scotia site is less energetic than at stations in the eastern North Pacific because the eastern Atlantic is essentially shadowed by the shallow Grand Banks from sources along the northern European and Greenland coasts. The preponderance of storms in the North Atlantic are too far North to be "seen" at the Nova Scotia site. Without this factor it is very difficult to explain the large difference between the eastern Atlantic and the western Pacific seen in typical spectral levels below 0.03 Hz. The short wave climate in the Pacific may be slightly more energetic than in the Atlantic, but not by the order of magnitude or more required by the observations. The decay time for long waves is slightly faster in the Atlantic than in the Pacific, but again not by an order of magnitude (more below).

The direct relationship of the coastal waves to infragravity wave energy is illustrated by the sequence of wave height maps under a hurricane (Figures 8a-8c). An intense storm in the North Atlantic (day 319) apparently contributes little to the infragravity wave spectrum at the HEBBLE site. The infragravity wave energy increases rapidly as a hurricane develops north of Puerto Rico. The energy increases by a factor of 5 as the hurricane and large waves approach land (Figure 6b, days 321-322). Very shortly after the hurricane enters the Gulf of Mexico through the Florida straits, the infragravity energy falls to a much lower level.

The decay time of the infragravity energy will depend on both the propagation velocity and the efficiency of reflection at coastlines. Long period waves with wavelengths comparable to the ocean depth travel at a phase velocity $C \approx \sqrt{gH}$. These waves are nearly nondispersive and the group velocity is nearly equal to the phase velocity. In 5000 m of water, a long wave propagates at a group velocity of about 220 m/s. A short-period wave travels at a phase velocity $C \approx \sqrt{g/k}$ and group velocity $U \approx 1/2\sqrt{g/k}$, so that a 12-s wave travels at a group velocity of about 9 m/s.

The distance from the Florida shoreline to the HEBBLE site is about 2500 km. This is probably a good estimate of the typical distance from most major storms to the HEBBLE site. A 150-s period surface wave will propagate this distance in about 3 hours. The 12-s wave will travel the same distance in about 74 hours. We have calculated the time lag correlation between the average squared significant wave height (Figure 6d) and the energy in infragravity wave band at the HEBBLE site over the length of the record (Figure 12). The maximum correlation is about 0.8 and occurs at a lag of less than 0.3 days. It is difficult to assess the errors in these calculations because the correlation coefficients are dominated by a few large peaks in the data and because the data are not evenly distributed during the day. However, the results do seem to suggest that the infragravity wave energy seen at the deep-sea site is related to faster propagating, long, coastally generated free waves, rather than forced waves tied to slowly propagating swell waves.

The scatter plot (Figure 11b) shows a linear trend between the infragravity wave energy at the HEBBLE site and the mean



Fig. 11. (a) Scatter plot of infragravity wave band energy versus the mean squared wave height averaged over grid points along the entire coastline of the North Atlantic. (b) Scatter plot of infragravity wave band energy versus the mean square significant wave height derived from FNOC grid points nearest the coastline, for all sections of the North Atlantic coastline which can be connected by a great circle path not crossing land or the Grand Banks. The apparent slope of the relationship between short wave and long wave energy is very different depending on which model for the short wave energy is used.

square significant wave height (line of sight calculation) from Figure 6d. We had expected a linear relationship because measurements from the surf zone suggest a linear relationship between low-frequency sea level variability and the amplitude of the waves in the short wave field. The coefficient of the relationship between the infragravity wave height variance inferred from the pressure spectrum and the mean square significant wave height along the shore is about 0.1 mm²/m². The mean square significant wave height is usually about 16 times the variance of the wave amplitude [*Neumann and Pierson*, 1966]. These results suggest that for this North Atlantic site, the average amplitude of waves in the infragravity wave band is about 800 times smaller than the average wave height along the coastline in the short wave band.



Fig. 12. Time lag correlation between infragravity wave energy and the mean squared wave height along the North Atlantic coastline. The wave height series leads the infragravity wave energy series by a small fraction of one day.

This contrasts with the surf zone where the amplitude of swash can be 70% of the incident short wave amplitude, and with observations from a few hundred meters offshore were the long waves may be 10% of short wave amplitudes.

Conservation of energy requires the amplitude of a long wave propagating into deeper water to decrease because the group velocity becomes larger. The amplitude varies with the inverse fourth root of the water depth (Green's Law; see Lighthill [1978]). The wave amplitude should decrease by about a factor of 6 from just outside the breaker zone (about 4 m) to deep-water (5000 m), but the infragravity wave measurements still suggest only a tiny fraction (<1%) of the energy leaks away from the coast into the deep ocean.

We have so far ignored attenuation in this model. Infragravity waves propagate in the deep ocean with essentially no attenuation [Lighthill, 1978]. In shallow water, long wave attenuation is thought to result primarily from bottom friction, although one can propose other mechanisms such as interaction with turbulence and breaking short waves. Long wave attenuation in shallow water is difficult to calculate, since the frictional losses are not linear with velocity. The long waves are superimposed on other phenomena; the attenuation may depend on the amplitude of the tides, currents, or internal waves and short-period waves [Cox and Sandstrom, 1962]. We look to work on the dissipation of tides and tsunamis for guidance. The tsunami originates with an earthquake. The direct wave arrivals are first observed, but after an initial period of "diffusion" the tsunami is observed to lose energy synchronously across the basin with an e-folding time for the Pacific of about 22 hours [Van Dorn, 1984, 1987]. The e-folding time in the much smaller Atlantic is about 13 hours. Munk [1963] proposed that tsunamis lose energy during multiple reflections from the coastline at a rate equal to an e-folding per reflection thus the characteristic time for the Pacific is about 22 hours, and the smaller Atlantic, about 13 hours. Recent work has confirmed this view, but estimating the reflection coefficient (which is associated with attenuation on the shelf) is still mostly ad hoc [Van Dorn, 1987]. Van Dorn's estimates assume the dissipation is linear with the width of the shelf so that the energy loss is proportional to area. No frequency dependence for this attenuation has been detected, but all tsunami decay observations have been based on much longer periods than the observations presented here.

Reflection at coastlines seems to be a very inefficient process on average for tsumanis with 63% of the energy lost during each reflection (an e-folding). We are apparently justified in ignoring reflected waves in our simple model. The tsunami attenuation estimates include the effects of propagation twice across the shelf as well as losses in harbors and straits and so may overestimate the losses for infragravity waves initially propagating away from a coast. Attenuation may be important in deriving infragravity wave amplitudes, but including realistic attenuation in the modeling appears to be impossible. The tsunami estimates do suggest an upper bound for the attenuation of infragravity waves. If 63% of the energy on average is lost traveling twice across a shelf, then at most 32% will be lost during one crossing. The estimate for the attenuation in one crossing of the shelf is sufficiently small to convince us that the equal weighting of waves along all shorelines in our model is not entirely without justification, although waves crossing some sections of the shelf will be more attenuated before reaching the deep sea than waves crossing other sections.

One can argue that the width of the shelf should not be a factor in the dissipation of long waves. The amplitude of a reflecting long wave reaches a maximum very close to shore. An envelope of a normally incident wave on a linearly sloping beach in the absence of dissipation can be described by $J_0(2\omega\sqrt{X}/\alpha g)$, where X is the distance off-shore and α is the beach slope (Guza and Bowen, 1976). Refraction will tend to turn all incoming waves nearly normally incident to the beach. The orbital velocity reaches a maximum at an argument of about 1.37, or a distance off shore of only 24 m at a frequency of 0.01 Hz on a beach of slope 0.02. Perhaps all the dissipation of long waves occurs essentially in the very energetic surf zone. The surf zone is very energetic compared with most of the shelf, and it is here that the infragravity waves are largest, so interaction with the bottom should be the strongest. No measurements appear to exist which discriminate between Van Dorn's view of dissipation distributed over the shelf, and this view of most of the energy loss occurring in the surf zone. Of course, the frequencies we consider here are one or 2 orders of magnitude higher than in the tsunami studies.

The tsunami results suggest the infragravity wave energy should decay after an impulse forcing with an e-folding constant of about 13 hours. The hurricane's passage through the Florida straits (Figure 8) drives the most rapid change in infragravity wave energy. The decay time is close to one half day, but the decay is obscured by the intermittent collection of data during each day.

THE PEGASUS EXPERIMENT

Eleven instruments were deployed in a 80 km by 80 km array on the seafloor 1000 km west of Pt. Conception, California, during the PEGASUS cruise (Figure 4; 34°30'N, 132°30'W). The average depth at the site is about 5100 m, corresponding to a free surface wave phase velocity of 223 m/s. The instruments measured fluctuations of pressure and the horizontal components of the electric field in a band from 0.0005 to 1 Hz. The purpose of the array was twofold, first, to detect propagating signals in the infragravity wave and microseism bands and, second, to detect electromagnetic signals generated by a towed electric dipole as part of an active source electromagnetic sounding experiment. This latter experiment was conducted by C.S. Cox and S.C. Constable. A gravimeter was also deployed on the seafloor by J. Hildebrand as part of a "fifth force" gravity experiment completing the PEGASUS acronym (pressure, electromagnetic, gravity, active source, underwater survey). The array was deployed during a 3-week period during November 1989. We will focus on the week of November 14–21, 1989, during which the array was fully deployed on the seafloor.

The resulting array is of appropriate dimensions to study the wave number-frequency spectrum of infragravity waves in the band from 30- to 512-s period. The number of elements in the array is inadequate for good results from conventional beamforming techniques. We have used instead the maximum-likelihood method as developed by Capon [1969].

The results of the maximum-likelihood beamformer show that the energy in the infragravity wave band is indeed constrained to the surface gravity wave dispersion relation. Calculations of the wave number spectrum in various frequency bands and over various intervals demonstrate the wave energy is constrained to lie on a ring of constant magnitude in wave number corresponding to the free gravity wave wave number (Figure 13). The wave number

a) November 15, 0600Z, T = 256 s



Fig. 13. Contour plots of wave number spectra during different days and in various frequency bands derived using maximum likelihood techniques. The wave numbers are normalized so that magnitude 1 is equivalent to a wave traveling at the deep-water free surface wave velocity of 220 m/s. Waves at higher frequencies travel slower than this velocity, so the energy lies slightly outside the wave number magnitude equal to one circle in all of these plots.



Fig. 13. (continued)

axes have been scaled so that a wave number equal to 1 corresponds to a wave traveling at the long wave (shallow water) phase velocity of 220 m/s. The shallow water wave phase velocity is the limiting velocity at very low frequency, at the PEGASUS site, waves at a frequency of 0.0039 Hz (256 s period) travel about 6% slower than this velocity and waves at 0.0064 Hz (157 s period) travel 14% slower. We see no evidence for significant energy in waves traveling much more slowly than the

free wave velocity (larger wave number), which suggests forced waves are a negligible component of the pressure spectrum. One peak in Figure 13d suggests there are waves traveling from the northeast, perhaps 15% too slowly; however, it may well be an artifact of the array binning.

The limited resolution and aperture of the array is evident in the broadening of the peaks in the wave number spectrum at the lower frequencies. At 512 s, the array dimensions are smaller than a wavelength (about 110 km). At higher frequencies the discretization in wave number is apparent in the contour plots as the maximum-likelihood technique tends to put energy in isolated wave number bins.

The largest peak in the estimates of the wave number spectrum in Figure 13*a* is associated with waves propagating toward an azimuth of 210–220°. A second major peak corresponds to waves propagating toward an azimuth of 260–270°. The first set of waves originate in approximately the region of Vancouver Island, the second set from the California coast north of Point Conception. The spectrum from a few days later, shown in Figure 13*e*, shows energy in waves propagating toward a broad range of azimuths extending from 130° to 290°. The waves originate along a broad section of coast extending from northern Japan along the Aleutian Islands and the west coast of North America as far south as the tip of Baja California. Little energy originates from the southern ocean, or the tropical western Pacific.

A chart of wave heights from the Pacific for this period shows energetic waves in the Northern Pacific and quiet conditions throughout the rest of the Pacific (Figure 14), except for two storms at 50° S, one at 180° W and the other at 90° W. We have not detected long wave energy in the spectra originating from either of these southern storms except perhaps at the very longest periods. In Figure 13c there is a broad shoulder extending from an azimuth of 270° through north. Other wave number-frequency plots shows an isolated a peak associated with waves propagating toward an azimuth of about 345°. The coastline of Antarctica may be a poor source of long waves, since the short waves which are required to generate the long waves are attenuated by pack ice before reaching the coast. The waves from storms in the roaring forties and fifties may be directed primarily toward the east, and so interact with no coastlines except the southern coast of South America. The great circle path to the southern tip of South America starts from the PEGASUS site at an azimuth of about 345°. We can suggest no reason why only these very long waves reach the PEGASUS site from the South Pacific, but there is no evidence for energy in waves from the south at periods shorter than about 270 s.

We do not see any energy in waves originating elsewhere in the Pacific, and although difficult to quantify, this suggests scattering and reflection from coastlines and islands is small, in general agreement with the tsunami dissipation measurements [Van Dorn, 1987].

CONCLUSIONS

Pressure fluctuations at periods longer than 30-s period are a ubiquitous feature of pressure spectra from instruments sited on the deep ocean floor in both the Atlantic and the Pacific. We have shown that these pressure fluctuations are predominantly caused by freely propagating surface gravity waves, and not the result of forced waves driven by the nonlinearity of the surface wave equations and associated with groups of wind-driven waves or swell. The energy in the pressure spectrum at 0.01 Hz measured at the HEBBLE site in the northeastern Atlantic varies 40-fold during an 80-day record. These variations in the long wave energy can be explained well by a model which assumes the long wave energy



Fig. 14. A chart of the significant wave height over the Pacific Ocean at 0000UT, November 15, 1988. Contours are approximately in meters.

originates near shore, driven by the same mechanisms which cause "surf beat". The model assumes the long wave energy is linearly related to the incident short wave energy averaged over all coastlines which are within the "line of sight" of the site. Storm waves in the far north Atlantic do not contribute to the long wave energy seen at the HEBBLE site. The HEBBLE site may be "in the lee" of the Grand Banks, and this may be the reason this Atlantic site is 10–30 dB quieter at infragravity wave periods than sites in the eastern Pacific.

Wave number-frequency spectra calculated from data from the PEGASUS array show the energy at infragravity wave periods confined to the free wave dispersion curve. Peaks in the spectrum suggest the Gulf of Alaska and the Californian coastlines are the two primary sources of infragravity wave energy during this paricular week in November 1988. Some wave energy originates along the entire coastline from northern Japan, along the Aleutian Islands and the western coast of North America to the tip of Baja California. A chart of wave height in the Pacific show large wind driven waves may be incident along this entire stretch of coast at this time. Storm waves incident on the southern tip of South America generate waves at very long-periods (270–512 s) which were detected at the Pacific array.

The entire infragravity wave source problem will be investigated in a multi-array experiment called SAMSON during the fall of 1990 off the east coast of the United States. The arrays will start in the surf zone and extend 600 km off-shore in a large joint acoustic and oceanographic experiment supported by the Office of Naval Research.

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