Low-frequency noise observations in the deep ocean

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Simultaneous measurements of ocean-bottom infrasonic and ocean-bottom and sub-bottom seismic noise in the frequency band 0.1–20 Hz are presented. The data were obtained in 5.5-km-deep water in the South Central Pacific with a triaxial borehole seismograph and six triaxial ocean-bottom seismographs having externally mounted hydrophones. The borehole sensors were emplaced 54 m within basement rock overlain by 70 m of pelagic clay. In the band 0.1–1 Hz, noise propagation as seismic modes trapped in the seafloor is supported by observed spectral coherences, cross phases, and ratios between ocean-bottom pressure and vertical ground motion, and by the relatively lower noise levels in the borehole. Noise variations in this band are clearly correlated with changes in local wind direction and speed, presumably through ocean-bottom pressure fluctuations caused by nonlinear wind wave–wind wave and wind wave–swell interaction.

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INTRODUCTION

In principal, seismic seafloor and acoustic ocean motion are related through kinematic boundary conditions at the benthic interface. Therefore, one may reasonably conclude that ocean-bottom seismic and infrasonic noise are related in source and character. This relationship is confirmed by observations of these two types of ambient noise. Measurements of seismic noise at the ocean bottom reveal the presence of the ubiquitous microseism peak, actually an agglomeration of peaks, over the microseism band (periods between 1–10 s) (Bradner et al. and Latham and Sutton). Ocean-bottom pressure measurements of high resolution over the microseism band, of which very few are published (Latham et al., Nichols,4 and Cox et al.), also display the microseism peak. Observed amplitude and phase relationships between the ocean-bottom pressure and vertical-component seismic noise are consistent with those of normal modes trapped near the ocean–seafloor interface (Latham and Nowroozi).

Over the microseism band an intimate relationship between microseismic noise and nonlinear surface gravity wave interactions has been both hypothesized (Longuet-Higgins and Hasselmann) and observed (Haubrich et al., Kibblewhite and Ewan, and Cox et al.). This nonlinear interaction occurs between opposed gravity waves of the same period and produces significant ocean-bottom pressure fluctuations of half the gravity wave period. The microseism peaks are the manifestation of this phenomenon.

Although regional swell is most frequently identified as the primary source of microseisms in the deep ocean, recent observations have emphasized the role played by local wind-generated waves (Kibblewhite and Ewan). Hasselmann predicts a spectral form for ocean-bottom pressure fluctuations \( \alpha f^2 S^2(f/2) \) at frequency \( f \), where \( S(f) \) is the gravity wave spectrum. The asymptotic form of so-called equilibrium wind-generated waves is \( S(f) \propto f^{-5} \) (Pierson and Moskowitz, Hasselmann et al.) or, in some circumstances, \( f^{-4} \) (Kahma). Thus a falloff for well-developed wind waves \( f^{-5} - f^{-1} \) would be expected and, in fact, is observed (Cox, Latham et al., Fig. 6, and Nichols).

This paper presents ocean-bottom and sub-bottom observations from a recent ambient noise experiment in the South Central Pacific in which the dominant mechanism of ambient noise generation was clearly wind-related. The growth of spectral peaks and generally increased levels over the microseism band in both seismic and acoustic measurements followed sharp changes in wind direction. Activity in the microseism band abated as the wind persisted or died away. Amplitude and phase relationships between pressure and vertical seismic motion at the microseism periods imply that a portion of the noise propagated as trapped modes of the ocean seafloor elastic system. Spectral peaks, at periods corresponding to acoustic reverberations in the water–sediment column, suggest that some of the ambient noise may be due to (leaking) organ-pipe modes or, more likely, trapped modes of the seafloor.

These observations constitute the first comparison of simultaneous ocean-bottom and sub-bottom seismic noise and ocean-bottom acoustic noise over the microseism band in the deep ocean.

I. EXPERIMENT DESCRIPTION

The Defense Advanced Research Projects Agency developed a marine borehole seismograph, the Marine Seismic System (MSS), for seismic discrimination applications. The successful test of an MSS prototype tethered to the D/V Glomar CHALLENGER on Deep Sea Drilling Project (DSDP) Leg 78B (Ballard et al.) led to deployment of the fully operational MSS on DSDP Leg 91, which composed the first portion of the Ngendei Experiment.
A primary objective of the Ngendei Experiment was the appraisal of the relative seismic signal and noise characteristics at a marine borehole site and proximate ocean-bottom site. This behavior was at issue because the first generation of ocean-bottom seismograph (OBS) efforts during the 1960s revealed that ambient seismic noise levels in the microseism band were typically 10–20 dB higher at ocean-bottom sites than at nearby land sites. In the microseism band, seismic noise at land sites is dominated by fundamental mode surface waves, the amplitude of which diminishes with depth into more rigid basement rock (Sorrells\textsuperscript{15}). Since marine seismic noise is similarly composed, it was thought that the same advantage might be obtained at a marine borehole site over an ocean-bottom site.

The MSS sensors were deployed in DSDP Hole 595B, located midway between New Zealand and Tahiti at 28°49.34'S, 165°31.61'W. Further details of the Ngendei Experiment may be found in Volume 88/91 of the Initial Reports of the DSDP.

II. INSTRUMENTATION AND OPERATIONS

A. Ocean-bottom seismographs

The six OBSs are modifications of the four-component, digital event recorders described by Moore et al.\textsuperscript{16} (Orcutt et al.\textsuperscript{17}). Output from three components of ground motion (one vertical and two horizontal) and a hydrophone are amplified, bandpassed, and digitized with a 12-bit word at 128 samples/s. The gain of each channel is $2^N$, where $N = 0,1,...,9$. A channel's gain exponent $N$ is modified only when the long-term average of its digital output, corrected for dc offsets, falls outside pre-defined bounds of a few least-significant bits.

Semiconductor memory, nominally 13 s long per channel, continuously accumulates the four channel data in a cyclically overwritten buffer to avoid the mechanical shocking and electrical transients of recorder operation. Software writes events to tape at scheduled times or when the short term average of vertical component amplitudes exceeds a long-term average threshold. Additional RAM boards extended the 13-s buffers of several OBSs to 59 s.

The OBS employed two seismometer responses [Fig. 1(a)], selectable only prior to deployment. The "teleseismic" response emphasizes lower frequencies more than the "refraction" response in order to increase the sensitivity of the OBS event-triggering algorithm to moderate-amplitude, high-frequency depleted earthquake signals. The "refraction" and "teleseismic" capsules employed the same hydrophone response as shown in Fig. 1(a).

The seismometers are the 1-Hz free period Mark Products L-4-3D triaxial set, and the hydrophone is an Ocean and Atmospheric Sciences, Inc., model E-2PD deep-water sensor. A buoyant aluminum sphere mounted on a detachable, dish-footed tripod houses the OBS instrumentation. Explosive bolts, detonated either by an acoustic or programmed command, free the instrumentation sphere, which then floats to the surface for recovery.

B. Marine Seismic System

The MSS is discussed in more detail in Adair et al.\textsuperscript{18} The MSS comprises a borehole instrumentation package connected by coaxial cable to either an ocean-bottom recording unit or to two shipboard recording systems employed for diagnostic checkout and recording redundancy. One of the shipboard recording systems, referred to as the Teledyne system, specifically monitors the borehole package, while the other, the Gould system, emulates the ocean-bottom recording unit. A special tool at the end of the drill string carries the sensor package to the seafloor, although simple lowering of the mechanically strong coaxial cable, "wireline" deployment, provides an alternate means to achieve this end. To date, only drill string deployments have been employed, although successful recovery of the borehole package has been achieved by wireline.

The MSS sensors are four Teledyne model S-750 seismometers configured as a triaxial set (two orthogonal horizontal and one vertical) and a redundant backup vertical component. The backup sensor is located 175.5 cm below the primary vertical sensor. Output from the three primary sensors is filtered in mid-period (0.5–10 s) and short-period (0.05–2 s) bands while output from the backup sensor is passed through the short-period band only. The resultant seven data streams are then digitized and transmitted to the recording devices via coaxial cable. For convenience, the sensors and data channels are labeled in subsequent discussion. The horizontal sensors are X and Y, and the primary and backup verticals are Z and B, respectively. The prefixes "S" and "M" will refer to the passband of a particular channel.

![FIG. 1. Displacement responses of (a) ocean-bottom seismograph (OBS) and (b) Marine Seismic System (MSS) instrumentation used during the Ngendei Experiment. (a) OBS seismometer and hydrophone responses. Two responses were available for the OBSs. (b) The Gould MSS mid-period and short-period responses differ from the Teledyne responses by the low-pass digital filter shown with dashed lines.](image)
Data recorded on the Teledyne system are continuously gain-ranged to yield a 21-bit dynamic range with 14-bit resolution. No low-pass (anti-alias) filter is employed other than the roll-off of the instrument response. The mid-period and short-period bands are digitized at 4 and 40 sample/s, respectively, although either of the two short-period vertical channels may be digitized at 80 sample/s to the exclusion of the other.

Output recorded by the ocean-bottom recorder and its shipboard counterpart are oversampled 256-fold, transmitted to the recorder, digitally convolved with a low-pass filter, resampled at the proper rates of 4 sample/s and 40 sample/s for the mid-period and short-period bands, respectively, and buffered to tape as 24-bit integers with 24-bit resolution. The data are oversampled to ensure that the roll-off of the instrument response prevents aliasing prior to application of the digital filter. The transfer function of the digital low-pass filter is nearly flat to 75% of the ultimate Nyquist frequency, and thereafter falls at approximately 750 dB/decade. The low-pass filter coefficients are computed in an identical manner for both passbands. The dashed lines in Fig. 1(b) illustrate the augmentation due to the low-pass filter.

C. Operations

The first 20 days of the 60-day-long Ngendei Experiment comprised a seismic refraction experiment. The OBSs recorded refraction signals and teleseismic signals with two distinct arrays and the MSS operated with its two shipboard systems. The remaining 40 days were devoted to a teleseismic experiment during which the OBSs formed a single array for earthquake and noise recording, and the MSS employed its ocean-bottom recording system.

The D/V Glomar Challenger conducted MSS installation and shipboard recording operations. The Scripps Institution of Oceanography R/V Melville tended the six OBSs and recovered all instrumentation. Figures 2 and 3 show instrumentation locations throughout the experiment. In these figures, octagons centered on ellipses depict estimated OBS locations and their 95% confidence bounds based on least-squares modeling of satellite and acoustic ranging data (Creager and Dorman19). Diamonds and arrows denote OBS drop points computed from satellite navigation and Melville radar ranging of Challenger. The borehole location is computed relative to the best satellite fix of Challenger's location.

During the refraction experiment, four OBSs were placed 0.6–0.7 km from the MSS site [Fig. 2(a)], and two at approximately 30 km [Fig. 2(b)]. The former quartet of capsules employed the "refraction" seismic response, 13-s event memory, and scheduled event recording, while the latter pair employed the "teleseismic" response, 59-s event memory, and triggered event recording with scheduled noise samples at 3-h intervals.

The borehole package was emplaced at the bottom of DSDP Hole 595B, approximately 54 m within a basaltic basement overlain by a 70-m-thick layer of zeolitic pelagic clays bearing hard stringers of porcellanite and chert near its base. The respective compressional and shear wave speeds of the sediment layer are 1.6 (Kim et al.20) and 0.116 km/s (Sereno and Orcutt21). The underlying basement structure is adequately modeled as normal oceanic crust (Whitmarsh et al.22).

The borehole package was not clamped in place in Hole 595B. The package's holelock/cable isolation mechanism was found to damage the coaxial cable and was therefore not employed. Comparison of MSS and OBS waveforms indicates that MSS sensor coupling with the ground was adequate. A coil of the coaxial cable payed out on the seafloor aided in cable-motion isolation.

During the teleseismic experiment, the six OBSs were redeployed in a triangular array with single capsules forming two 25-km legs at approximate right angles to the south and east of the MSS site [Fig. 3(a)]. The other four capsules were clustered within 3 km to the west and north of the MSS.
in a trapeziform array [Fig. 3(b)]. All OBS employed the "teleseismic" response and triggered event recording, two with 13-s memory buffers and four with 59-s buffers. Daily noise events were recorded at Universal Time midnight. The MSS ocean-bottom recording system failed two days after entering the water due to a leak.

Successful recovery of all OBSs and the MSS recording system took place from MELVILLE over the period 22-26 March 1983. Following a 2-day shipboard recording period, the MSS recovery mooring was redeployed with a dummy load replacing the recording system. Both the borehole package and recovery mooring are still in place for future use.

III. NOISE DATA

Totals of 119, 43, and 83 h of seismic data were obtained from the Teledyne system, the Gould ocean-bottom recording system, and its shipboard counterpart, respectively. During the refraction experiment, 114 and 70 h of shipboard data were recorded on the Teledyne and Gould systems, respectively. OBSs recorded a total of 1400 noise events at 740 distinct times, 140 concurrently with MSS data.

Various MSS data channels were unavailable at certain times. The horizontal sensor Y malfunctioned throughout the entire Ngendei Experiment for unknown reasons and provided no usable data. A shipboard source of electrical interference contaminated the subsequently disconnected Gould channel SZ. A Gould electronics failure in its borehole package bay disabled Gould channel MZ at the start of the teleseismic experiment. Finally, as mentioned above, one of the Teledyne short-period Z channels was not available whenever the other recorded at 80 sample/s.

IV. DATA QUALITY

Data quality was assessed with both time- and frequency-domain criteria, the latter applied to power, cross, and coherence spectra.

Reliable spectral estimates were computed in the manner described by Welch23 and Carter et al.24 Nonoverlapping, demeaned segments of digital data were tapered with a normalized Hanning window and Fourier transformed. The resulting spectra were averaged to achieve a reduction of estimate variance. Power spectra, defined only for non-negative frequencies, are normalized so that the total power in the Nyquist-frequency bandwidth equals the time-domain variance, a normalization which essentially folds the power of negative frequencies into the positive.

Estimates of statistical reliability were calculated under the assumption that the data were derived from a stationary, Gaussian stochastic process. Power spectral estimates are then distributed as $\chi^2$ random variables with $2N$ degrees of freedom, where $N$ is the number of spectra averaged to obtain the estimate (see, e.g., Bendat and Piersol25).

Cross-spectra between different time series are calculated in order to estimate their coherence and cross-phase. The coherence shall be used as a measure of the degree to which the time series are linearly related. The coherence spectrum, actually the magnitude-squared coherence spectrum, is defined as the ratio between the squared cross-spectrum magnitude and the power spectra of the constitutive time series. A useful application of the coherence is to define some threshold-coherence value below which the estimate is no different from zero. In other words, we wish to determine the conditional cumulative probability $P(\hat{C} < T | C)$ that our coherence estimate $\hat{C}$ is less than the threshold value $T$ given the actual coherence $C$. The complicated statistics of the magnitude-squared coherence are essentially those of the linear-regression correlation coefficient (Carter et al.,24 Fischer25), and the expression for $P$ involves a sum of hypergeometric functions. If the actual coherence $C$ is zero, this expression is quite simple. Then, $P$ is given by ([Carter et al.24]

$$P(\hat{C} < T | C = 0) = 1 - (1 - T)^{N-1},$$  

where $P(\ )$ is the probability that the estimate $\hat{C}$ does not exceed the value $T$ given the actual coherence $C$ and $N$ is the number of spectra averaged in the coherence estimate. Thus
The threshold coherence $T$ is given at the 100P% confidence level by
\[ T = 1 - (1 - P)^{1/(N - 1)}. \] (2)

In computing spectral estimates from OBS data, the number of spectra averaged, and stationarity, must trade off because OBS noise samples are available only at discrete times: $N$ must be large enough to yield estimate stability and useful error bounds, but not so large as to sample over a time interval such that the noise field has statistically evolved. $N = 10$, a 30-90-min interval, was found to be generally optimal. The 95% confidence interval is then $-2.0$ dB and $+2.7$ dB. This trade-off is not very important for MSS data since they were continuously recorded, except during brief interruptions. However, where comparisons of simultaneous observations are desired, the statistics and frequency resolution of the MSS estimates were made similar to those of the OBS.

Linear regressions of Gould data against Teledyne data, in conjunction with sensor calibrations, revealed a 3-dB shortfall of Gould amplitudes. The calibrations also show that instrument response below 0.2 Hz is not good for any MSS channel.

Figure 4 compares power spectral density estimates of simultaneous vertical-component Teledyne and Gould data with instrument noise spectra which have not been corrected for instrument response. The peaks above 8 Hz in the SB spectra are ship-generated. Peaks in the SB Teledyne spectrum [Fig. 4(c)] which are absent in the SB Gould spectrum [Fig. 4(b)] are manifestations of aliasing in the Teledyne data due to inadequate high-frequency filtering. Gould and Teledyne spectra have comparable values near the Nyquist frequency, indicating probable malfunctioning of Gould’s low-pass filter.

Also shown in Fig. 4 are estimates of system noise, provided by H. B. Durham of Sandia Laboratories. The system noise is not significant over the entire short-period passband for either Teledyne or Gould data. The flat spectrum of quantizing noise for the Teledyne data, associated with uniformly distributed round-off errors of digital representation (see, e.g., Bendat and Piersol,25 pp. 231–232), has a level near $10^{-2}$ counts$^2$/Hz, well below the Teledyne signal level. Estimates of Gould’s digitizing noise spectrum, that of a delta modulator, are not available. The MZ spectrum [Fig. 4(a)] is similarly untainted by system and sampling noise. The MZ instrument response rolls off sufficiently at high frequencies to avoid aliasing.

Spectral coherences (Fig. 5) further delineate frequencies usable for ambient noise estimates. Coherences exceeding 0.28, illustrated with a dash line, are considered significant at the 95% level. Mid-period coherences [Fig. 5(a)] typically exceed 0.95 below the low-pass filter knee at 1.5 Hz, except at 1 Hz, where it drops sharply over a very narrow band of frequencies. The sharp drop is due to energy present in the Gould data, but not in the Teledyne, and may be related to the Gould digitizing scheme. Gould and Teledyne SB coherences [Fig. 5(b)] exceed 0.8 between approximately 0.2–4.0 Hz. The coherence at higher frequencies declines amidst peaks due to ship-generated signals. The sharp drops at approximately 2 and 4 Hz are similar to that at 1 Hz in the mid-period coherence. The frequencies bounding the short-period band of high coherence varied a great deal during the MSS deployment, but generally were those shown in Fig. 5. Coherence estimates between SB and SZ Teledyne data [Fig. 5(c)] were similar. Differences at frequencies greater than 4 Hz are probably due to the combined effects of aliasing and actual signal differences over the 175.5-cm sensor separation. The high coherence between the Gould and

![Figure 4](https://example.com/figure4.png)

**Figure 4.** Spectral power density estimates of typical MSS noise samples compared to system noise levels. The spectra, given in counts$^2$/Hz, have not been corrected for instrument response. (a) Gould Channel MZ, (b) Gould Channel SB, (c) Teledyne Channel SB.

![Figure 5](https://example.com/figure5.png)

**Figure 5.** Coherence spectra, corrected for alignment, between (a) Teledyne and Gould channel MZ, (b) Teledyne and Gould channel SB, and (c) Teledyne channels SZ and SB.
Teledyne data at the Nyquist frequencies indicates a possible deficiency in the Gould low-pass filtering scheme.

OBSs employing the refraction response can measure lower-amplitude ground motion at high frequencies than those employing the teleseismic response. The refraction response was specifically designed to "whiten" the ambient signal in the presence of typical deep-sea microseismic noise (Moore et al.17), while the teleseismic response was designed to low-pass it. Filtered high-frequency amplitudes are consequently less than those represented by a least-significant bit as recorded with the teleseismic response. Figure 6 compares spectra measured with the different OBS responses (upper curves) and shows the corresponding spectra of least-count noise (lower curves). The spectra, corrected for instrument response, were not measured simultaneously. Although most of the data composing the teleseismic capsule's spectrum (solid upper curve) were recorded after ship departure, their spectral levels exceed ship noise levels measured with a refraction capsule (dashed lower curve) at most frequencies. In the teleseismic spectrum, sampling noise clearly dominates at frequencies greater than 15 Hz, and is probably significant in the range of 7–18 Hz.

The lowest valid frequency of OBS microseismic measurements is taken to be 0.1 Hz. At lower frequencies, least-count and system noise dominate. The system noise was inferred from the lack of estimate stability over a wide range of numbers of spectra averaged to compute power density estimates.

Ship-generated noise must be estimated on a case-by-case basis because it depends on ship speed, activity, and range. Generally, sharply distinct peaks superimposed on a broadband level characterize ship-generated noise. In the Ngendei data, most peak frequencies appear to be multiples of some fundamental, indicating that propellant cavitation caused the peaks (e.g., Urick27). CHALLENGER generated peaks at multiples of 10 Hz while maintaining station, and two others at approximately 6.5 and 7.5 Hz while under way.

**FIG. 6.** Comparison of noise spectrum obtained with the teleseismic (solid upper curve) and refraction (dashed upper curve) OBS responses. The corresponding instrument corrected sampling noise spectra (lower curves) are also shown.

**FIG. 7.** Demonstration of MELVILLE's ship noise during the period 034:0430-035:0000 (Julian date). (a) Smoothed, integrated OBS Suzy vertical-component power density between 17.9 and 18.2 Hz. (b) MELVILLE's range from the MSS site.

MELVILLE's generated peaks at multiples of 6 and 9 Hz. Figure 7 shows the growth of a peak at 18 Hz due to MELVILLE's proximity to the MSS.

Figure 8(a) illustrates ship-generated noise with a comparison of vertical-component power density spectra, corrected for gain but not for instrument response, of OBS Phred data recorded before (solid curve) and after (dashed curve) both ships departed. The capsule, located 1.6 km west of the borehole, employed the teleseismic response. The measurement prior to ship departure was made as both CHALLENGER and MELVILLE move to within 2 km of the OBS. In this instance, ship-generated noise clearly domi-
nates the spectrum at all frequencies above approximately 4.5 Hz. After ship departure, least-count noise dominated above 7 Hz, although the lowest frequency at which it dominated varied by ±1 Hz since the level depends on OBS gain.

Some ambiguity exists concerning ship noise in the MSS data. Because the ocean-bottom recorder failed 2 days after its deployment, when CHALLENGER was 13 km away drilling Hole 596 and MELVILLE was en route to Tahiti, MSS noise measurements in the complete absence of ships are not available. However, CHALLENGER was briefly silent c. 0510Z, 13 Feb., while monitoring an ocean-bottom navigational transponder, thus providing a sample of low ship noise for the MSS. Figure 8(b) compares the MSS spectrum of this time period with that measured a short time later when CHALLENGER began using its dynamic positioning system thrusters. Also shown in Fig. 8(b) is an estimate of short-period system noise which has been scaled to fit the "quiet" spectrum. The good fit suggests that above 7 Hz system noise dominates the spectrum.

The dominance of ship-generated noise is, in many instances, diagnosed by an abrupt change in spectral slope, such as is seen in Fig. 8 at approximately 4 Hz. On this basis, it is apparent from these and other spectra that ship noise may be important at frequencies as low as 2 or 3 Hz in both the OBS and MSS data.

In summary, reliable spectral estimates of ambient seismic noise are available from MSS mid-period data between 0.2–1.5 Hz; from MSS short-period data between 0.2–7 Hz; and from OBS data between 0.1–7 Hz. These bands are subject to the variability described above due to ship proximity, OBS gain range, and instrument response. At frequencies greater than 7 Hz, sampling noise dominates OBS teleseismic data, so that lower bounds on ocean-bottom microseismic noise levels must be established using data recorded with the refraction response.

V. RESULTS AND DISCUSSION

A. Spectral noise characteristics

Figures 9–12 compare displacement power densities for all available components in units of nanometers$^2$/Hz(nm/$^2$ Hz) for simultaneous MSS and OBS data. Each spectrum is the average of ten individual spectra. The individual spectra were each computed from approximately 13 s of data obtained at 5- and 10-min intervals over a 70-min period. The OBS, Suzy, was 0.6 km north of the MSS site. Based on considerations mentioned above, ship noise is assumed to dominate these spectra above 4 Hz.

The spectra all display the ubiquitous microseism peak near 0.2 Hz, where the observed noise levels are comparable, approximately 10$^8$ nm$^2$/Hz. At higher frequencies, ambient ocean-bottom microseismic noise levels exceed the borehole levels by 5–30 dB (Figs. 9 and 10). Common to the spectra, though not prominent in all of the OBS spectra, are peaks centered on 0.5 and 0.7 Hz. At other frequencies, horizontal noise levels generally greatly exceed vertical levels (Fig. 11). In contrast, horizontal levels in the borehole approximately equal the vertical levels (Fig. 12). Latham and Sutton find...
similar depth variation for the computed displacements of low-order trapped modes in an oceanic crust model, suggesting that these observations are consistent with the noise comprising such modes.

The ocean-bottom pressure spectrum measured simultaneously with the above spectra also displays the microseism peak (Fig. 13). The fall-off towards higher frequencies decreases with frequency from the microseism peak.

The microseism peak and a 0.4-Hz-wide band centered at 1 Hz display high coherence and a cross-phase of approximately 180° [Fig. 14(a) and (b)] between the pressure and vertical displacement. These observations are diagnostic of the type of seismic waves comprising the microseism peak. At the ocean bottom, the pressure and vertical displacement of the trapped modes of the ocean-seafloor elastic system have a cross-phase of 180° (Bradner, 28 Schneider et al. 29). On the other hand, plane waves exhibit a cross-phase of 90° (Schneider et al. 29).

The spectral ratio of pressure and velocity amplitude is another quantity diagnostic of noise composition. If, for example, the noise consisted of body waves traveling from afar, then the low compressional wave speed of the sediment would result in nearly vertical incidence of these plane waves at the ocean bottom. For vertically traveling plane waves, the pressure-particle velocity ratio is constant with frequency and equals the product of density ρ and acoustic speed α in the water (Schneider et al. 29), 1.5 × 10^6 in mks units. For trapped modes, this ratio may vary with frequency and exceed ρα (Latham et al. 3 and Latham and Nowroozi 6). Figure 14(c) plots the ratio between the pressure and vertical-component velocity power density spectra, normalized by (ρα)^2. Over the two bands of high coherence, this ratio is on the order of 10, rather than 1, suggesting that in this instance they represent trapped seismic modes. Thus, in this case, the practice of comparing seismic ground velocities V(f) to a "plane-wave equivalent" pressure P(f) via the relation P = ραV is quite incorrect.

Figures 15 and 16 show the coherence and cross-phase between the vertical and hydrophone channels of OBSs Suzy and Lynn, separated by 0.6 km, for the same time interval as above. Again, energy at the microseism peak is highly coherent. The cross-phase at these frequencies is not significantly different from zero, implying either simultaneous arrival of microseism energy at the two capsules or a wavelength much larger than the capsule separation. Latham et al. 6 calculate the dispersion for trapped modes in an ocean-bottom structure similar to that of the Ngendei site. At these frequencies the fundamental mode has a phase velocity C_p of 1.0–1.5 km/s, corresponding to a wavelength λ = C_p/f on the order
of 5 km. The capsule separation, roughly $\lambda/10$, is too large to account for the observed small cross-phases, which rules out oblique arrival of the microseism energy from a distant source. It therefore appears that the microseism source location in this instance is either local or along a line that is approximately normal to a line defined by the OBSs.

**B. Temporal noise variations and noise sources**

It has been generally observed that the sea state affects microseismic and infrasonic noise levels. Ample evidence is found that the microseism peak is due to nonlinear interaction of opposed ocean surface waves of a given frequency. The interaction generates acoustic waves with fluctuations of twice the waves' frequency (Hasselmann). This has been conclusively demonstrated for swell reflected from a shoreline (Haubrich et al.), and for wind waves (Kibblewhite and Ewan; Webb). Decklog observations provide observations pertinent to local and distant noise source discrimination. Figure 17 summarizes CHALLENGER decklog observations of wind and swell between 24 January and 13 February 1983 (Julian dates 24–44) at 2-h intervals except for 8–10-h swell gaps corresponding to local night-time, and for hourly observations during days 33–37. The prevalent swell arrived from the south and had 6–7-s periods. The highest amplitude swell had periods between 8–10 s, predominantly 9 s, and arrived from the south between dates 29–31 and 34–35. Very little northerly swell was observed.
The wind direction changed both suddenly and gradually, the latter being marked by smooth changes in wind speed.

Figure 18 shows the variation of OBS Phred vertical-component acceleration spectra with time at 3-h intervals, except for gaps due to triggered event recording. The spectra, single realizations corrected for instrument response, are computed from 58 s of data. The noise clearly changes with time, particularly at the prominent peaks near 0.2 and 0.5 Hz, which have comparable levels. Figure 18(b) shows the time variation of the power in a 0.2-Hz band centered on these peaks, along with wind [Fig. 19(a)] and swell [Fig. 19(c)] observations over the same period. Abrupt increases in the power of both peaks, e.g., days 28.0, 30.0, and 33.0, coincide in several instances with sudden changes in wind direction. The onset of high-amplitude 9-s swell coincides with the wind change and power increases at day 30.0, but opposing 9-s swell, perhaps masked by other swell, were not observed. The onset at day 33.0 of the high-speed wind episode, from the SSE, coincides with a large power transient of the 0.5-Hz peak and a gradual ramping of power of the 0.2-Hz peak. Noise observations of 13-s length, collected at 3-rain intervals from the "refraction" OBSs near the MSS site during day 33.0, show the peak at 0.5 Hz building quickly over the course of an hour. Power for both peaks gradually diminished as the wind weakened and wound northward. The OBS Phred pressure spectra varied similarly with time. The correlated behavior of the wind and noise clearly implicates the wind as a major source of microseisms.

The observed variations of the wind are clearly related to those of the noise in Fig. 19. A sharp change in the wind vector immediately precedes, to within the 2-h resolution of the decklog, noise increases at frequencies greater than the microseism peak band. Hasselmann shows that the direct conveyance of pressure from the atmosphere at these frequencies is unlikely, leaving only the nonlinear interaction of wind-generated gravity waves or of wind waves with background swell as a likely source for the seismic variations. In the case of wind–wave interaction only, Hasselmann derives the following expression for the spectral density \( S_p(f) \) of the ocean-bottom pressure fluctuations at frequency \( f \),

\[
S_p(f) = 8\pi^2 g \frac{\omega^2}{c^2} S_{w}\left(f - \frac{\omega}{2}\right) G(\theta) G(\theta + \pi) d\theta ,
\]

where \( \omega = 2\pi f \), \( S_{w} \) is the wave height power density, \( \rho \) and \( c \) are the density and sound speed of water, \( g \) is the gravitational acceleration at sea level, and \( G(\theta) \) is the angular dependence of the wind–wave spectrum. \( S_p(f) \) may be thought of as a forcing function for seismic wave generation. Note that the nonlinear interaction between exactly opposed waves of frequency \( f \) [cf. the angular integrand in Eq. (3)] leads to the double-frequency dependence \( 2f \) of the ocean-bottom pressure fluctuations. Implicit in Eq. (3) is a white wavenumber spectrum which contributes only at phase speeds exceeding the sound speed of water, that is for wavenumbers \( <\omega/c \).

A large body of observations (Pierson and Moskowitz, Hasselmann et al. support self-similar form for the wind–wave spectrum \( S_{w}(f) \) of

\[
S_{w}(f) \propto f^{-5/3} e^{-\left(\frac{5/3}{f_m} f\right)} ,
\]

The function described in Eq. (4) falls exponentially for \( f < f_m \) and as \( f^{-5} \) for \( f > f_m \), where \( f_m \) is the peak frequency. Empirically, \( f_m \) decreases and power density increases with time under the influence of a steady wind. Tyler et al. find that \( G(\theta) \) is adequately described by

\[
\cos^q(\theta/2) ,
\]

where \( q \) is a function of frequency and \( \theta = 0 \) corresponds to the mean wind direction.

A more detailed view of the interplay between the wind and noise may be obtained from the continuously collected MSS data. Figure 20 shows the temporal variation of mid-period vertical acceleration in the borehole using data from Teledyne channels MZ. Averages of 14 spectra computed from consecutive 128-s samples, with nearly 30 min of total data, form the individual spectra of Fig. 20. The dominant feature of these spectra is the microseism peak near 0.2 Hz, which with time steadily migrated towards lower frequencies, and grew larger in amplitude and narrower in width. Note the "saddle" in the peak's amplitudes near day 41 coinciding with generally increased levels at higher frequencies. In contrast to the ocean-bottom accelerations, levels at the 0.5-Hz peak [Fig. 18(b)] are an order of magnitude lower than at 0.2 Hz. Other peaks are observed at 0.35, 0.65, and 0.80 Hz (Fig. 20, detail).

Figure 21(b) compares the temporal variation of power for the microseism ("primary") and 0.5-Hz ancillary peaks.
of Fig. 20. Power was estimated by integrating over 0.10-Hz-wide bands centered on 0.22 and 0.50 Hz. Power variations in these two bands were roughly parallel until day 40.7, after which they were complimentary, corresponding to the microseism peak’s “saddle.” The power of these peaks decreased in an unknown manner early in the MSS shipboard recording period, then began growing at day 39.0. Power in the ancillary band diminished during day 39, but resurfaced day 40.0. Power in both bands changed little in the half-day preceding day 40.7, then decreased in the primary band and increased in the ancillary band. At day 41.0, power leveled off in the microseism peak band and generally increased at higher frequencies, which peaked day 41.2. The rate of power growth in the two bands increased at day 41.5, so that primary power increased and ancillary power falloff slowed. By the end of the shipboard recording period, power in both bands attained levels comparable to those observed near day 40.5.

Figure 21(a) and (c) shows weather observations during the MSS shipboard data collection period which began just as the epoch of the strongest winds observed during the Ngendei Experiment was concluding. Southerly swell of 6-7-s period dominated most of the time, although northwestern swell briefly dominated on day 41 [Fig. 21 (c)]. The wind, on the other hand, varied a great deal [Fig. 21 (a)]. A moderate northeasterly wind blew during most of days 37 and 38 at a nominal 10 kn. During day 39, the wind direction was confused, though still northeasterly. The wind blew northwesterly throughout day 40 and southwesterly for the rest of the shipboard recording period. The highest wind speeds during the MSS shipboard recording period, 18 kn, accompanied the shift in wind direction from northwest to southwest at day 41.0.

The available data permit only a gross description of the sea. However, the foregoing discussion of wind–wave evolution permits drawing reasonable inferences of the wind–wave behavior from CHALLENGER decklog observations. The increases in noise power at frequencies greater than those of the microseism band, which follow soon after sharp changes in the wind vector, are probably due to the wind degrading the pre-existing wave field. The subsequent power increases at these frequencies evident in Fig. 21, on days 39.0, 40.0, and 41.0, are due to the newly evolving wind–wave field.

The behavior of the microseism peak band is not as easily explained by the available data. The growth of power throughout days 38 and 39, as well as the shift towards lower frequencies apparent in Fig. 20, suggests an evolving wind–wave source. Similarly, the fall during day 40 and the subsequent rise during day 41 suggest the demise of the old wave regime and domination of the new regime in response to

FIG. 20. Temporal variation of vertical acceleration in the borehole in the frequency bands 0.1–1.0 Hz and 0.4–0.9 Hz (detail).

FIG. 21. Temporal variation of (b) peaks at 0.2 and 0.5 Hz of the MSS data shown in Fig. 20 compared with (a) wind and (c) swell observations shown in Fig. 17.
winds which blew approximately 90°, then 180°, with respect to the old wind.

If steady wind waves were solely responsible for the observed seismic noise variations, then Eq. (4) implies that a falloff of the wind–wave spectrum would result in an ocean-bottom pressure spectrum \( \alpha f^{-2} \), rather than the observed \( f^{-3} \) (Fig. 13). However, the ragged falloff observed in all spectra suggests either that the seafloor responded resonantly to the wind–wave forcing, or that a steady sea state was not achieved. Sharp peaks at approximately 0.35, 0.5, 0.65, and 0.8 Hz were observed in the borehole data. It will be recalled that the peaks at 0.5 and 0.65 Hz were prominent in one or more of the OBS spectra in Fig. 11.

The presence of equispaced peaks in ocean-bottom microseisms has also been reported by Bradner and Dodds, who attributed them to so-called "organ-pipe" modes due to the near-vertical reverberations of acoustic waves between the sea surface and the sediment–basement interface. Organ-pipe modes belong to a class of seismic waves known as "leaking" modes, in which the seismic energy radiates away into the deepest structure. In contrast, a "trapped" mode remains confined within the capping crustal structure. Abramovicz found qualitative agreement between conspicuous peaks in the microseism observations of Bradner and Dodds and calculations of organ-pipe mode dispersion curves and transfer functions. Organ-pipe modes are an important constituent of the seismic phase \( P_n \) observed in the ocean which, in this context, is composed of \( P \) waves traveling almost entirely in the upper mantle before turning sharply to the vertical in the sediment (Sereno and Orcutt). The resulting spectral peaks occur at frequencies \( f_n = (2n + 1)df/2, n = 0,1,2,... \), where \( df \), the separation between peak frequencies, is the reciprocal of the two-way travel time of a single reverberation. At the MSS site, \( df = 0.135 \) close to the observed 0.15-Hz peak spacing in the microseism data. However, these modes are damped in time \( t \) by the term \( e^{\tau (R)} \) (Rosenbaum), where \( R \) is the reflection coefficient for normally incident plane waves at the sediment–basement interface, and \( \tau = 1/df \) is the two-way travel time of a single reverberation. \( R \) is given by

\[
R = \frac{\rho_s \sigma_2 - \rho_2 \sigma_1}{\rho_1 \sigma_2 + \rho_s \sigma_1},
\]

where the subscripts 1 and 2 refer to the sediment and basement, respectively, and where \( \rho \) and \( \sigma \) denote density and compressional wave speed, respectively. For the MSS site, \( \rho_1 = 1, \rho_2 = 2.7 \) g/cc, and \( \sigma_1 = 1.6, \sigma_2 = 5.0 \) km/s (Shearer, personal communication). We find, then, that successive reverberations are damped by a factor of \( e^{-4.2} = 1.5 \times 10^{-2} \). If we assume constructive interference of successive reverberations, the net contribution is found from the geometric series in the damping to be only 1.015 times the initial amplitude. It appears unlikely, therefore, that organ-pipe modes contribute significantly to the local generation of microseisms.

On the other hand, some crustal structures exhibit regular peak spacing for trapped modes because, as Rosenbaum and Haddon show, they are essentially continuations of leaking modes. At some transitional phase speed, the trapped-mode spacing is approximately that of the organ-pipe modes. Modeling the relative contributions to the ambient noise of these complementary wave types will be the topic of further research.

VI. CONCLUSIONS

The observations reported here constrain the source mechanisms and propagation modes of infrasonic and microseismic noise in the deep ocean. High coherences, 180° cross-phases, and ratios between displacement and pressure at frequencies below 1 Hz imply a trapped-wave propagation mode. Modeling of these measurements, as well as the horizontal-component measurements, may further constrain the wave type, although a definitive experiment to measure the wavenumber spectrum may be required.

Episodic variations of the wind direction and speed correlate with changes in both acoustic and microseismic noise. In particular, a peak at 0.5 Hz responded immediately to sudden changes in wind direction, both in ocean-bottom and sub-bottom observations. The microseism peak is also affected by the wind, an observation that is in contrast to multitudes of previous reports focusing attention on swell as the chief agent of noise generation. The mechanism for transferring energy from wind to microseisms is presumably the nonlinear interaction of wind-generated waves. Spectral shapes and the temporal evolution of the pressure and displacement spectra support this hypothesis.

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