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ULF/VLF (0.001 to 50 Hz) Seismo-Acoustic Noise in the Ocean

Proceedings of a workshop at the Institute for Geophysics,
University of Texas, Austin
November 29 to December 1, 1988

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G. H. Sutton

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Executive Summary

Scientific Objective

The objective of the ONR Accelerated Research Initiative (ARI) in ULF/VLF seismo-acoustic noise in the ocean is to improve our quantitative understanding of the physical processes that generate the acoustic/seismic ambient noise field in the frequency band from 0.001 to 50 Hz as a function of location and time (pp. A2-12):

- to experimentally identify and quantitatively define the important noise sources including the discrimination between noise from local and distant sources;
- where possible, to theoretically establish and describe the physical processes which are the noise source in the ULF/VLF frequency band; and
- to determine the effects of ocean, bottom and sub-bottom acoustic/seismic parameters on the observed noise field, i.e., the effects both of propagation and of local site conditions.

It is understood that only "natural" noise sources will be emphasized and that, e.g., sources from shipping will be considered only as required to evaluate their contribution to the noise field.

General characteristics of important noise sources and of the noise power spectrum are already fairly well established over much of the frequency band of interest; most are related to ocean waves and their meteorological sources (see, e.g., reports of Frequency Oriented Working Groups, pp. 2-9 and talk abstracts, pp. A97-136). Ice motions, earthquakes, volcanic eruptions, turbulent tidal (or other) water currents, and atmospheric acoustic disturbances (also discussed in the Proceedings), are other possible sources that might be considered where appropriate.

Propagation through the ocean/crustal waveguide between source region(s) and receiving instrument arrays produces important effects on the observed ambient pressure and the seismic motion components. Relationships among these components depend

strongly on the wave type (e.g., free traveling waves, such as body, Rayleigh, Love, Stoneley/Scholte, or forced deformations of the ocean-bottom produced by short-wavelength pressure disturbances) and upon the detailed velocity/attenuation structures of the oceanic waveguide; in general, the relationships also are strongly frequency dependent.

The elastic properties of the ocean floor differ by orders of magnitude between soft sediments and igneous rock - sometimes over very short distances; such local site variations can degrade array performance. Improper coupling of seismic sensors to soft sediments results in serious signal distortion and excess noise, especially for the horizontal components of motion.

The program outlined below of 1) theoretical investigations, 2) analysis of existing data, 3) development and testing of improved measurement techniques, and 4) integrated field experiments, if followed during the ARI, should produce a significant improvement in our knowledge of the ULF/VLF ambient noise field and our ability to predict its geographical and temporal variations. Each of the four elements of the program are considered to be of equal scientific priority.

1. Theoretical Investigations (pp. 10-12, A92-97)

Many details concerning the sources and propagation of ULF/VLF ambient noise in the ocean are poorly understood. This is the result both of inadequate observational data and of incomplete theoretical models. To meet the objective of the ARI it is necessary that both be improved to the point that reliable, testable predictions can be made of expected noise from given sources and vice versa. Studies using land-based seismometers have identified two marine noise sources. In the .05-0.3 Hz frequency range, microseismic noise is associated with seismic surface waves from coastal sources. In the .15-.5 Hz frequency range, large storms at sea have been identified as sources of body

waves. Sea-floor observations (A-15, A-31, A-50), however, indicate that the noise field in the 0.05-30 Hz range is quite strongly controlled by local wind and swell conditions. The dependence between noise level and wind (or wave height), at first glance, seems to be at variance with current models (A-97, A-15). Resolving this paradox will require ocean wave directional spectra concurrent with sea-floor noise observations.

- Because of the large changes in waveguide structure across continental and island margins realistic two- (and three-) dimensional propagation models are needed.
- The importance of incoherent scattering must be investigated.
- Quantitative theories for proposed source mechanisms must be developed and evaluated.
- Contributions of local site conditions to observed spectra must be determined.

2. Analysis of Existing Data (pp. 18-24, A15-23, A31-39)

In addition to relevant data from recent field experiments that are currently being analyzed, other existing long-term, broad-band seismo-acoustic data (both land and marine), meteorological data, and physical oceanographic data (e.g., wave spectra from spaceborne synthetic aperture radar) provide valuable information on temporal and geographic variability of observed ambient noise and its sources. (Data to be obtained from related current programs, e.g., LFASE, SWAPP, SWADE, SRP on acoustic reverberation, and other related Navy programs, would also be useful.) Analysis of some of these data is currently being supported through this ARI; it is recommended this and related research be directed toward specific outstanding recognized problems such as:

- the origin, time variability, and source regions of single (5-10 sec) and double (10-20 sec) frequency microseisms;
- effects of intense storms;
- estimates of wind, wave, and swell activity in the vicinity and season of proposed field experiments; and
- relationships between ambient noise observed on land and the ocean floor.

3. Development and Testing of Improved Measurement Techniques (pp. 13-17, A77-91, A97-105)

A major component of this ARI is the development of a standardized microprocessor controlled, digital OBS system for use in coherent array studies. The modular construction of this system permits the use of various configurations of up to six seismometers and/or hydrophones (or differential pressure gauges, DPG). Testing of the new instruments at very early stages is important to ensure that instrument performance does not hamper later large-scale uses of the new instruments.

Hydrophones and DPG's adequate to monitor the whole ULF/VLF frequency band are currently available; current OBS seismometers are not adequate to cover the ULF range. (However, apparently adequate seismometers do exist). Signal distortion and excess noise, especially on horizontal motion components, remains a problem with OBS. Electric field and pressure gradient measurements can provide data related to horizontal component motion. Planned tests of OBS coupling and of improvements from shallow burial will be useful. Except for a few cable-based systems, no long-term (several months or longer) OBS have been deployed. Short experiments, e.g., one month duration, do not adequately sample all weather conditions at a site. In addition to improvements in far-field noise measurements, significant improvements are needed in techniques for near-field measurements of noise from swell and breaking waves, and of directional spectra of the waves themselves from near-surface and satellite or aircraft remote sensing techniques. Implementation of the following recommendations will clearly benefit attainment of the ARI goals:

- development of a technique for high-fidelity measurement of horizontal motion;
- development of a ULF seismic sensor package;
- development of long-term monitoring capability to provide adequate temporal statistics;

- development of near-source acoustic and ocean wave measurement techniques;
- development of adequate remote sensing of directional wave spectra.

Testing of these recommended improvements should be coordinated with the field experiments recommended in the following element.

4. Field Experiments (pp. 25-48, A24-26, A40-76)

A coordinated sequence of field experiments is proposed to obtain the data required to meet the objective of the ARI in ULF/VLF seismo-acoustic noise in the ocean. Where possible, this field work also will be coordinated with other ongoing experiments in order to maximize efficient use of personnel, ships and instrumentation. Results from earlier experiments discussed in the Appendix, provide background and guidance for current planning.

The primary source of seismic noise is the earth's atmosphere, which in turn drives the ocean waves and current, and, thus, it is important that noise measurements be related to physical oceanographic (PO) observations. Since, in many cases, the PO measurements are at least as costly as seismic measurements, there is great economic advantage in marrying the observational programs for the two. The ONR has underway an Accelerated Research Initiative in Ocean Surface Wave Dynamics. This ARI is sponsoring two field programs in 1990. The first one is Surface Wave Processes Program (SWAPP) and its field program will be carried out off Pt. Conception in February-March 1990. The second is the Surface Wave Dynamics Experiment (SWADE) which will be conducted off the East coast during the fall of 1990. SWAPP is a deep-sea experiment instrumented with high-frequency Doppler sonars (A-97) and radars mounted on FLIP. The sonar will be used for measurement of the ocean wave directional spectra and the radar will detect wave-breaking events. The Canadian vessel Parizeau will deploy instruments (A-123) to make high frequency acoustic measurements of breaking waves in

addition to other PO observations. SWADE offers an instrumentation suite focused on processes at oceanic margins and, in the context of this program, is well suited for the study of edge waves (A-134) which may be the limiting noise process in the noise 'notch.'

The proposed suite of field experiments, described in eight separate reports, are designed to address six major problem areas:

- generation and propagation of ambient noise near continental margins (and active plate boundaries);
- geographic distribution of deep-ocean ambient noise over given time periods;
- near-source investigation of wind-waves and noise during severe weather conditions;
- coherent array studies of the directional spectrum of deep-ocean ambient noise;
- strong effects of local bottom material on observed ambient noise; and
- effects of ice cover on observed ambient noise and characteristics of ULF/VLF ambient noise in the Arctic.

With the possible exception of active plate boundaries (whose tectonic interest should attract other funding agencies), all the above problem areas seem to be of approximately equal importance for attaining the objectives of the ARI. Scheduling and resources should be assigned according to practical considerations of funding, personnel, ship, and instrument availability, and of possible cooperative opportunities with other projects.

Some of the eight described experiments, as mentioned in the reports, are natural candidates for combined or coordinated field programs, e.g., the Arctic Ocean experiment with the Ice margin experiment and the Distributed OBS Array and/or Continental Margin experiments with the severe weather experiment.

Introduction

This report presents results from an ONR sponsored workshop to discuss scientific issues, experiments and theoretical work for an Accelerated Research Initiative in ULF/VLF seismo-acoustic noise in the ocean. The workshop was held November 29 to December 1, 1988 at the Institute for Geophysics, University of Texas, Austin. A list of the 41 participants is included in Appendix B at the end of the report. The interactions of marine acousticians and seismologists, physical oceanographers, and specialists in satellite remote sensing of the ocean surface produced an interesting and mutually informative meeting.

The two and a half day workshop began with a series of 24 short invited talks providing background for the rest of the workshop. Extended abstracts of these talks are included in Appendix A; they contribute useful support for the body of the report. After the background talks, four working groups were organized to discuss scientific issues, possible experiments and required theoretical work in four frequency bands: 0.001 to 0.01; 0.01 to 0.3; 0.3 to 10; and 10 to 50 Hz. After a short time, because of strong mutual interests and requirements the 0.001 to 0.01 and 0.01 to 0.3 Hz groups combined. In the middle of the second day the groups were rearranged to define specific observational, instrumental and theoretical problems and to develop proposed experiments, to be conducted during the ARI, to address those problems.

The working group reports that follow are organized in five categories; frequency oriented reports; theoretical considerations; instrumentation considerations; use of existing data; and proposed experiments (see Table of Contents). As is to be expected, there is some overlap among the reports, e.g., some theoretical work and experimental observations are recommended in the frequency oriented and instrumentation reports.

As mentioned earlier, the extended abstracts in Appendix A provide important background material on the history and current status of knowledge and experiments in ULF/VLF seismo-acoustics. The first three papers review ONR plans for the ARI and related interests of NORDA and IRIS. Following papers discuss historical and current research activities; ONR supported instrumental development; analytical techniques; ocean wave noise sources; and VLF noise in the Arctic Ocean.

0.001 to 0.01 Hz and 0.01 to 0.3 Hz Working Groups

Participants: Spahr Webb, Co-Chmn; Joan Oltman-Shay, Co-Chmn;
Tokuo Yamamoto, Co-Chmn; George Sutton, Co-Chmn;
Dean Goodman; Robert Beal; H. M. Iyer; Adam Schultz;
Robert Holman; Jerry Smith; Robert Guza; Charles Cox;
Mark Riedesel; and Rick Adair

0.001 to 0.01 Hz

Seafloor pressure noise in the band from 0.001 to 0.01 Hz appears to be caused solely by waves on the surface of the ocean. Seafloor displacement noise also appears to be set by ocean waves which directly force seafloor motions. Other possible sources of noise in this band are seafloor currents, earthquakes and infrasound (acoustic) waves in the atmosphere.

Low frequency ocean waves are called "infragravity" waves. These waves can either be free waves, or forced waves associated with wave groups. The free waves travel at phase velocities set by the surface gravity wave dispersion relation; the forced waves travel at much slower phase velocities. Either kind of wave must be of a wavelength comparable to the water depth to be detectable at the seafloor. Recent observations suggest both types of waves can be important. Pressure measurements from the shallow continental shelf show low frequency waves propagating toward the coast, presumably associated with wave groups. Deep water observations show a well defined high frequency cutoff in the spectrum of infragravity waves. This sharp decrease in energy at shorter periods would not necessarily occur if most of the energy was associated with forced waves. A primary goal for research in this frequency band should be to determine the roles of these two types of waves in determining ocean pressure and seafloor displacement signals.

Some work has been done in estimating the amplitudes of seafloor motions associated with infragravity wave pressure signals. This work shows the displacement signal depends on the detailed elastic structure below the seafloor and focuses on determining this structure from the pressure and displacement measurements.

Ocean currents can effect seafloor measurements in two ways. Strong currents induce a turbulent boundary layer with significant pressure fluctuations, but the flow also interacts with the instruments causing vibrations, and eddies are shed from the instrumentation causing pressure fluctuations. The instrumentation can be designed to reduce the "flow noise", and burial of the instruments should greatly alleviate the problem. Some signals associated with the boundary layer pressure fluctuations may still effect buried instrumentation. Acoustic noise in the atmosphere has also been proposed as a source of seafloor noise. Infrasound is easily measurable at terrestrial sites. These signals are of sufficient wavelength to generate significant pressure signals at the seafloor also, but have not yet been identified in seafloor observations.

Instrumentation

Coherent arrays of pressure transducers can be used to study the infragravity waves. In shallow water, pressure transducers can also be used to measure the wind wave and swell components of the surface gravity wavefield to permit a study of the effect of wave groups. In deep water, pitch and roll buoys and acoustic doppler and radar measurements from stable platforms such as FLIP could be used to make comparable surface wave measurements. Displacement measurements in this frequency band require very low noise, long period seismometers such as the Guralp, or LaCoste and Romberg instruments. Current meters could be deployed with other seismic instrumentation to study flow induced noise. Instrumented surface buoy arrays might be used to study the component associated with acoustic waves in the atmosphere.

0.01 - 0.3 Hz

A primary feature in the deep-water seafloor pressure spectrum in this band is a band between about 0.03 and 0.1 Hz with very low spectral levels. The pressure signals from infragravity waves establish the spectral level at low frequencies, and microseisms establish the level above 0.1 Hz. At intermediate frequencies, no primary mechanism for noise generation has been determined.

The pressure signal associated with infragravity waves decreases away from the surface exponentially, with an e-folding distance equal to the inverse of the wavenumber. The pressure signal at the seafloor becomes very small for waves of wavelengths much shorter than the ocean depth. There is therefore a sharp decrease in spectral levels above about 0.03 Hz in deep water, but surface waves may be detected with frequencies as high as 0.3 Hz at a depth of ten meters. In shallow water the infragravity waves fill in the spectrum and no intermediate frequency trough is observed in the pressure or displacement spectra.

A peak in the spectrum near 0.06 Hz is occasionally observed in deep water measurements. This peak is recognized from terrestrial measurements as the "single frequency microseism" peak. The origin of this energy appears to be waves breaking on the coast and the peak is called the single frequency peak because the frequency follows the primary ocean swell frequency. The physics of this peak is not very well understood. The problem requires that energy in very slowly propagating ocean waves (10 m/s) should be coupled into seismic waves at phase velocities exceeding 4 km/s.

Spectral levels rise sharply above 0.1 Hz into the "microseism" peak. The seismo-acoustic noise in this band is generated by nonlinear interactions of surface gravity waves. There is a frequency doubling associated with this mechanism so the oceanic noise occurs at frequencies equal to twice ocean wave frequencies. At frequencies between 0.1 and 0.2 Hz the ocean noise is associated with ocean waves of periods between 10 and 20 seconds and therefore primarily with the swell rather than the local wind waves. This mechanism requires waves propagating in nearly opposing directions. Swell, by definition, comes from distant sources and so is primarily from a single direction. The noise is thought to be the result of swell interacting with the same swell

reflected off nearby coastlines.

Ocean turbulence is also a possible cause of noise in this band. Another, unexplored source may be waves breaking overhead which might produce pressure signals with periods twice the primary swell frequency. Earthquake surface waves may also contribute to the noise in this band.

Instrumentation

The instrumentation appropriate for experiments in this band are similar to the instrumentation required in the 0.001 to 0.01 Hz band. The direct verification of the physics of the double frequency microseism peak will require very accurate measurements of the directional spectrum of the swell components. These measurements might be attainable with acoustic doppler sonars, radar measurements or perhaps pressure transducers arrays (in shallow water). Mid-water column measurements with freely drifting sensors may be useful for studying the microseism peak as well as any components that may be associated with wave breaking or atmospheric acoustic waves. The single frequency peak may require extensive surface wave measurement in the nearshore to determine those sections at which the coupling between surface waves and seismic waves occurs. Nearshore wave staff or pressure transducer arrays might be used to measure the swell components reflected from the coastline. Arrays of seismic instruments either on land or in the sea might be used to determine source regions also. Less expensive seismometers such as the Mark Products L-4 become useful at frequencies above about .05 Hz.

0.3 to 10 Hz Working Groups

Participants: Fred Duennebier, Co-Chmn; Leroy Dorman, Co-Chmn;
Peter Shearer; Tony Schreiner; Chip McCreery;
Dale Bibee; Robert Cessaro; Charles Cox;
Marty Dougherty; Jan Garmany; Brian Lewis;
Yosio Nakamura; Antares Parvulescu; G. Michael Purdy;
and Altan Turgut

Science Questions

Forcing Functions:

Wind Waves: (0.2-6 Hz) Observed usually as a saturated level at higher frequencies (3-5 Hz) with spectral slope indicating wave-wave nonlinear double-frequency interaction.

Swell: (<0.5 Hz) Is local swell observed, or is the energy propagating as Rayleigh waves from distant storms?

Breaking Waves: (>3 Hz, acoustic, and lower frequencies when breaking on shores) Do breaking waves generate acoustic noise observed in this part of the spectrum?

Ships: (>1 Hz) Observed as sharp spectral lines.

Ice: (?) Relatively unstudied in this band.

Atmospheric Turbulance: (?) The acoustic component may propagate to the ocean floor.

Bottom Currents: Create noise on instruments, but may be avoided with proper design.

Bottom Turbulance: Creates noise around obstacles, in conjunction with bottom currents.

Volcanism: Regional importance with possible propagation to quieter areas.

Earthquakes: All frequencies, transient. may present quiet noise floor at some frequencies.

Boundary Conditions:

Shorelines: How do the relative importance of forcing functions change as shorelines are approached?

Water Depth: How do the relative importance of forcing functions change as water depth changes?

Bottom Rigidity: How does the noise spectrum change from regions of soft bottom to regions of hard bottom? How do these parameters change with sediment thickness?

Bottom Roughness: How does the noise spectrum change from regions of smooth bottom to regions of rough bottom?

Ice: When the free surface is bounded by ice, what is the effect on the noise spectrum?

Coupling to Propagation:

- A. How does wave energy get transferred to the ocean floor?
- B. How are Rayleigh waves generated at the bottom from wind waves and swell at the surface?
- C. What are the importance of Stoneley waves, Love waves, and scattering in the propagation of noise along the ocean floor.
- D. What is the directionality of noise propagation? (Local vs. coastal generation of noise at swell periods.)

Possible Experiments

Long Term:

1. Partial Ice-Over: observe change in noise spectrum as ice covers the seas over an OBS array.
2. Array studies in variable sea states: observe changes in noise spectrum as sea state changes over an OBS array. Requires good environmental data.
3. Quiet-site recording: observe noise spectra in quiet regions (deep arctic, tropics), identify noise sources when there are no waves.
4. Large storm: observe noise spectra changes directly below a large storm, and observe modes of noise propagation to other sites using widely spaced array. The directionality of arrivals would aid in identification of sources as from the storm itself, particular shorelines, or whatever.

Short Term:

1. Hard/soft bottom: Observe variations in noise spectra as bottom characteristics change from hard to soft bottom and thin to thick sediment.
2. Depth transect: Observe variations in the noise spectrum with changes in depth of the ocean.

Environmental Measurements

- A. Cartesian Diver: pressure acoustic near-surface
- B. Waverider (pitch and roll buoy): ocean wave spectrum

- C. FLIP acoustic doppler sonar: ocean wave directional spectrum
- D. Wind speed and direction
- E. Atmospheric pressure (array)
- F. Bottom current
- G. NOAA buoys
- H. Sonobuoys and XCTDS's from planes when ships not available

Analysis of Existing Data

Several data sets exist that can be studied further to aid in answering some of these questions.

The Columbia OBS data are a long uninterrupted time series that could be analyzed for directionality of approach, correlation with storm activity, and temporal variation of the noise spectrum. These data are in-hand and require only funds for digitization and analysis.

The Wake Bottom Hydrophone Array is a working array in deep water that is routinely recorded by HIG. Data from the array yield information from 0.1 to 30 Hz, and cover much of the range of interest. Data are currently available for the past five years, and only small sections have been analyzed. The array could be used as a fixed-point standard for long-term experiments, and, if coupled with several more instruments and environmental measurements, could be valuable in determination of noise sources. It has the particular advantage that it is in-place now, and data are available in real-time.

10 to 50 Hz Working Group

Participants: George Frisk, Co-Chmn; John Orcutt, Co-Chmn;
Jan Garmany; Winston Chan; David Farmer;
Antares Parvulescu; Yosio Nakamura; and Henrick Schmidt

Scientific Questions

What are the primary natural sources of noise in this band? What is the energy budget for these sources? What are the physical mechanisms by which it propagates?

Major Postulated Sources of Noise

Shipping

This is a major source of noise in this band on which a considerable amount of applied research has been conducted. Therefore it does not fall within the purview of this basic research program which concentrates on natural noise sources. However, it is necessary for the community to be briefed on the state-of-the-art knowledge of shipping noise in order to evaluate its contribution to the energy budget.

Wind

The determination of the role of wind-induced noise will require the solution of a number of problems associated with air-sea interaction. For example, wave breaking has been shown by Farmer and Melville to be an important source of noise at higher frequencies, and Duennebier's measurements suggest that it is significant in the 10 to 50 Hz band as well. Careful acoustic, directional gravity wave, and atmospheric boundary layer measurements (e.g., wind speed and stress) will be required in order to clarify this issue. Measurements made under extreme conditions, such as hurricanes, would be very helpful in solving these problems. The impact of distant storms must also be assessed. A suitable theory of sound generation by wind must be developed. It is also important to study the role of bubbles in noise generation, perhaps in cooperation with the ONR reverberation program.

Ice

Ice noise arises due to a variety of mechanisms which include: a) thermal cracking, b) ice collisions, c) the formation of leads and pressure ridges, and d) gravity waves impinging on the marginal ice zone. A suitable approach to studying ice cracking noise, e.g., would be to: a) develop theories of noise generation for an individual crack and for an ensemble of cracks, b) make measurements of the ice noise using geophones on the ice and hydrophones in the water, and c) simultaneously make independent measurements of the cracks, e.g., by taking pictures of the cracks induced by an icebreaker.

Other Miscellaneous Sources

Other sources of noise in this band worthy of investigation but considered to be of lower priority are: a) biological (e.g., whale sounds), b) seismic (e.g., earthquake,

volcanic and thermal vent), and c) turbulent flow over irregular topography.

Theory of Noise Propagation

A complete, three-dimensional theory of noise propagation must be developed. This should include the effects of propagation in a strongly scattering medium in order to account for the "noise" introduced by random inhomogeneities. In addition, in order to understand the noise source mechanisms, source effects must be distinguished from propagation effects.

Instrumentation

It is important to minimize the use of cable-mounted instruments because of the significant amount of strumming noise in this band. Freely drifting sensors, such as noise-measuring SOFAR floats and sonobuoys, are desirable alternatives. Arrays could be synthesized by tracking several of these sensors. Both horizontal and vertical arrays are essential to the determination of noise directionality. The importance of making environmental measurements simultaneously with the acoustic measurements cannot be overemphasized.

General Comment

This band is unique in that sound by man-made sources can be readily generated in it, thereby providing the opportunity to ground truth certain postulated ideas on noise generation and propagation. For example, ships of opportunity could potentially be used as controlled noise sources.

Theory of Generation from Atmosphere and Ocean Surf

(Charles Cox)

Theoretical analysis is needed for further understanding of the hydrodynamic processes involved in the generation of noise in the ocean. Existing theory (Longuet-Higgins, Hasselmann) accounts well for the generation of double-frequency pressure fluctuations from opposed swell trains in the frequency range .11 to .5 Hz. The double frequency mechanism must be modified for high frequency waves where the dispersion relation of the sea surface waves is no longer well defined. The importance of non-linear interactions between components of the surface wave spectrum suggests that higher order terms beyond the second may play a role in the subsurface pressure field.

Other mechanisms give rise to possibly important noise generation over a wide frequency range. For example the fact that waves of the sea surface superpose so as to create randomly spaced groups of unusually high amplitudes leads to variations of mean sea level between regions of large and small amplitude waves. It can be shown under the restriction to second order theory that this process does not lead to coupling to the deep ocean acoustic field but it certainly leads to pressure fluctuations in shallower water, and the higher order effects are unknown. As an extreme example, the existence of whitecaps must lead to pressure pulses probably associated with groups. The typical repeat period of whitecaps is twice the principal wave period and thus these pressure pulses may be a contributor to the noise in the .03 - .1 Hz noise notch. The wind field is strongly distorted over breakers and the associated pressures acting on the sea surface may also play a role in generating noise in this frequency range.

Deep sea pressure fluctuations are essentially uncoupled from slowly moving pressure components in the atmosphere because movement of the upper ocean in response to the pressures annul the deeper pressure field. On the other hand pressure components with phase velocities equal or greater than the speed of long surface waves can reach all depths in the sea. An example of rapidly moving pressure disturbances in the atmosphere is a sound wave. Theoretical work is needed to work out the magnitude of sound waves to be expected in the atmosphere. Some will undoubtedly be associated with the turbulent shear flow of the wind over the water even though the rms turbulent velocity is very far from the speed of sound.

Propagation of Microseisms

(Henrich Schmidt)

The ambient noise measured by means of seismic sensors is a convolution of the source spectrum and the transfer function of the environment.

The transfer function, accounting for the propagation from source to receiver, has significant frequency dependence due to the waveguide nature of the ocean and continental crusts. Basic understanding of these propagation features is essential for mainly two purposes:

1. Determination of the source mechanisms and level from the microseismic noise measurements.
2. Explaining the spatial distribution of microseismic noise in the ocean and on the continents.

Research Issues

1. Basic issues such as the effect of water depth and seabed geology on microseism propagation. To what depth is the geology important?
2. Propagation from storm areas and coastal source areas across the ocean-continent boundary.

Methods

Some of the basic issues may be addressed using existing theoretical and numerical models, including the effect of water depth and the vertical heterogeneity of the ocean crust.

Since the coastal sources such as wave breaking tend to behave like time-sources parallel to the coastline, the problem of propagation from these into the ocean and onto the continent may be addressed using 2-dimensional finite element or finite difference codes. Due to the computational requirements of these techniques, the development of more efficient hybrid approaches should be considered.

To address the propagation of microseisms from storm center to ocean as well as continental arrays will require inclusion of 3-dimensional effects. 3-dimensional seismic propagation codes, tailored to 5th generation computers must be developed to make such propagation modeling feasible.

Propagation in Random Media

(Jan Garmany)

Noise studies seek statistical descriptions of observed wavefields such as power spectral estimates at single receivers and cross-spectral estimates between pairs of receivers. Array studies of wavenumber spectra effectively depend on cross-spectral estimates and their comparison with simple models of predicted phase variations (plane waves). These estimates are based on second-order moments of the observed field (the correlations of two series at a time). Higher order moments of the wavefield (fourth order and above) are studied very little, since they are derivable from the lower order moments in simple propagation problems. The theory of propagation in weakly scattering media is well developed and provides useful estimates of deterministic spatial fluctuations in a medium (e.g., migration in seismic exploration) and of the statistical fluctuations in media which may be too complicated or too rapidly varying to warrant analysis by deterministic means. In ocean acoustics, there has been much effort expended on deterministic studies of propagation in range-dependent media, with increasing attention to strongly reverberating (scattering) media. These important studies make it possible to study the origins of strong fluctuations that are seen in actual field data, and to draw some conclusions about the fate of acoustic energy in the oceans. However, at some point we must resign ourselves to our ignorance of the detailed structure in the water and the crust, and of the sources of noise that give rise to the ambient wavefield. Statistical properties of the wavefield can be predicted from the statistics of the medium, but only if we can include strong scattering effects. The propagation of waves over thousands of kilometers are clearly subject to these effects. We must also consider local effects of strong scattering, especially in the case of very slow Scholte waves, which will be aliased in any practical arrays to be considered in the proposed work. Propagation in strongly scattering random media is therefore a relevant component of the SNAG effort, particularly in the medium- to high-frequency range studies.

ULF OBS Sensor/Instrumental/Burial

Participants: Tokuo Yamamoto, Co-Chmn; George Sutton, Co-Chmn;
Fred Duennebier; Chip McCreery; Antares Parvulescu;
Mark Riedesel; Altan Turgut; and Dean Goodman

Objective

To discuss the technical requirements for developing buried ULF OBS arrays to measure the ambient noise field, the "ULF notch" level in particular.

Available Seismic Sensors

The committee recognizes that ULF seismic sensors satisfying the technical requirement of this objective are available commercially. They include GURALP CMG-3, Streckeisen (perhaps) and Teledyne-Geotech KS-54000 seismometers.

Housing Configuration and Design

The committee recommends that the seismometer housing be partially evacuated to prevent disturbances generated by turbulence inside the pressure package. It is also a design consideration to determine the required length of the umbilical cable between the seismic sensor and the recorder/battery package in order to reduce package induced noise and distortion.

Installation/Burial

The committee strongly recommends that the sensor package should be buried within the seabed. This is required to eliminate the rocking of the OBS by current. Also, the exposed sensor housing causes uneven settlement of the package due to consolidation. Burial is absolutely necessary for accurate measurements, especially of the horizontal components, at ULF (0.01-0.1 Hz) band. The required depth of burial of the seismic package is to be assessed depending on the sediment types and disturbance characteristics at the site. In any event, it is estimated that burial depths up to 1 m below the seafloor is sufficient. This requirement can be met by existing University of Miami burial system. The only additional requirement for deep water application is the acquisition of a deep water submersible pump for the University of Miami burial system.

Miscellaneous Recommendation

The committee recommends acquisition of preamplifiers for the Wake Island Hydrophone Array in order to improve ULF response down to 0.05 Hz. The committee also suggests development of a fiber optic cable for supporting permanent (or semi-permanent) OBS arrays in addition to ONR deep water OBS systems.

Conclusions

The committee strongly recommends that buried OBS capabilities are developed during the ARI in order to measure temporal and spatial variations of total seismic noise levels of the deep ocean bottom, particularly at the ULF notch. This is the single most

important objective of the ARI. Tokuo Yamamoto, George Sutton, Fred Duennebieer and other interested researchers from the scientific community could take responsibility for this task.

Proposed Experiments

Two self contained ULF OBSs will be built for the deep water experiments. A new deep water hydraulic jet OBS burial system will be built by following the design concept of the existing shallow water burial system at University of Miami.

Burial experiments in deep water will be conducted on the abyssal plain near Baltimore Canyon where both sandy and muddy bottoms are found within a short distance in deep water (5 km). One OBS will be buried while the other OBS will be deployed in a conventional manner. At each site, 40 hours of continuous recording will be made of the ULF (0.001-1 Hz) seismo-acoustic ambient noise fields. By comparison of data from the two OBSs, the effects of burial on the ULF seismic noise measurements will be evaluated.

Burial System

University of Miami's burial system has been tested and improved throughout three sea experiments to bury their OBS seismic packages in any type of sediment on the continental shelves. A feasibility study has been done to modify the burial system for deep water application. A 3 kilowatt submersible DC motor will be used to pump the sea water through hydraulic jets to liquefy the sediment under the seismic package.

The proposed OBS has a main package (recording and battery packages) and a seismic package (see Figure 1). Two packages are attached together with a carriage rod which will be released automatically when the BOS is on the ocean bottom. The seismic package will be buried 1m below the seafloor by pumping the sea water for two minutes. After the burial operation, the burial system will be disengaged by a timed magnet release and recording will be started. The main package and seismic package will be recovered by releasing the acoustic releases. Two acoustic releases have a timed back-up option so that in case of a failure in acoustic interrogation the main package and seismic sensor will be recovered after a certain set-up time.

Estimated Budget

Buried ULF OBS Experiments in Deep Water

Personnel	200k
Equipment	
2 ULF OBSs	100 k
Deep Water Burial System	100 k
Shiptime	100 k
	<hr/>
Total	500 k

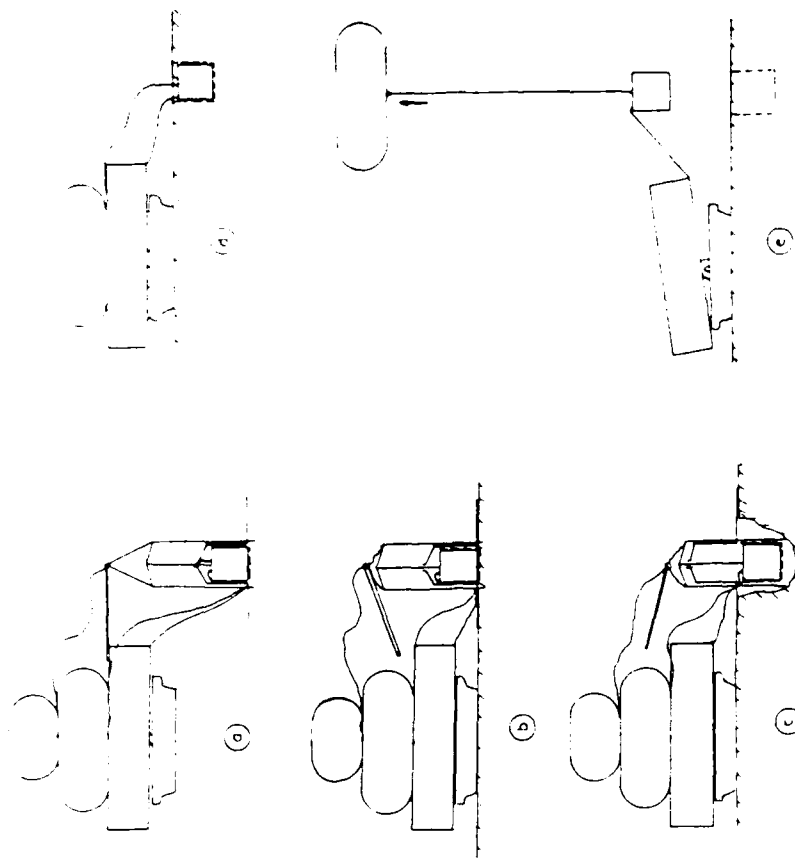
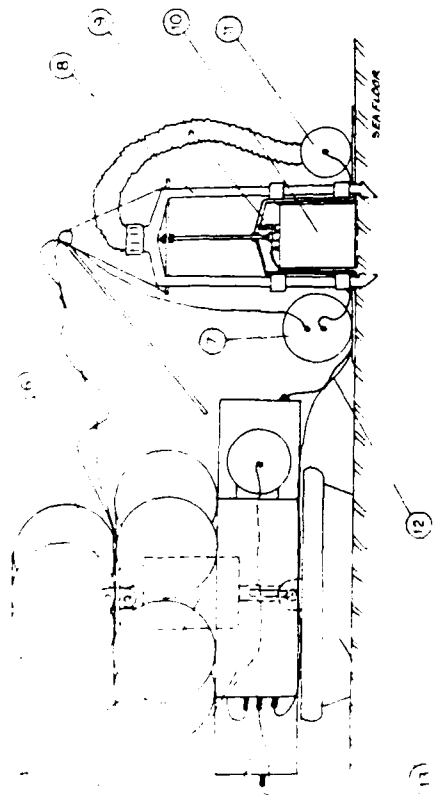


Figure 1b Deep water OBS deployment and recovery.



- 1- BATTERY PACKAGE, RECORDING SYSTEM
- 2- ELECTRONICS HOUSING (DIGITIZERS, AMPLIFIERS, ETC.)
- 3- MAGNETIC RELEASE WITH TYPED RELEASE BACK-UP
- 4- CLASSIFICATION BUOYS (17 INCH DIA.) FOR MAIN PACKAGE RECOVERY
- 5- CLASSIFICATION BUOYS (17 INCH DIA.) TOP BURIAL SYSTEM RECOVERY
- 6- MAGNETIC RELEASE
- 7- BATTERIES PACKAGE FOR DE-SUBMERSIBLE PUMP AND MAGNETIC RELEASE
- 8- BURIAL ROCKET
- 9- MAGNETIC RELEASE
- 10- SELF-BURIAL PACKAGE
- 11- DE-SUBMERSIBLE PUMP
- 12- TYPED RELEASE CABLE

Figure 1a RSMAS deep water, self-burial OBS.

An Air-Deployable Acoustic Drifter for Low Frequency Sound Studies

(David Farmer)

In order to understand the sources and subsequent propagation of sound close to the ocean surface, it is necessary to develop ways of acquiring observations in this difficult environment. Wave-breaking is thought to be a primary sound source, at least above about 1-2 Hz, but there has been little or no measurement of this process at high wind speeds. The need for near surface measurement, so as to resolve the fine structure in the signal which is normally lost through spatial filtering when sensors are placed at depth, imposes special demands on deployment, recording capability and data recovery. There is a need for the development of an air-deployable device that could acquire essential data with which to resolve surface acoustic noise sources and the related generation mechanisms. One possible configuration is outlined below.

The device would be small and light enough to be deployed, like a sonobuoy, from an aircraft in the path of an intense storm or hurricane. It would, however, include both active and passive sensors; the passive sensor would be a hydrophone capable of useful acoustic sensing down to at least 5 Hz. The active sonar would transmit a high frequency pulse at 1-2s intervals vertically upwards so as to acquire both wave and bubble cloud data, a technique that has proved very effective from ship deployed systems. For short periods data could be related directly to the deploying aircraft. However, it would certainly be desirable to have on board processing and data reduction capability, so that key results could be downloaded via VHF to an aircraft on a subsequent flight. This would require an appropriate positioning device on the surface float.

Previous acoustic measurements of surface generated sound have either been obtained from the sea-floor, in which case the temporal variability of the signal is lost by the spatial filtering of the effective listening area; or the measurements have been obtained from instruments deployed by ships, which understandably avoid very severe weather. Sea-floor instruments also suffer from lack of any active sonar sensing of the ocean surface. Nevertheless, available data have shown that individual breaking waves are responsible for acoustic radiation at least down to 40 Hz.

An air-deployable version equipped as indicated above, would allow us to determine whether wave breaking was in fact responsible for the known sound level increase at lower acoustic frequencies during storms. It would also allow us to learn more about the behavior of the ocean surface at very high wind speeds, and to measure the distribution of bubbles. Bubble clouds have been shown to influence higher frequency sound propagation at lower wind speeds. It is probable that in extreme conditions their influence extends to much lower frequencies.

Use of Existing Instrumentation and Data

(H. M. Iyer, George Sutton, and Robert Cessaro)

Available broadband, 3-component (land and ocean bottom) and array seismic data from past (now discontinued) seismic networks can be particularly useful to address questions relating to single-frequency microseisms in addition to other ULF/VLF studies.

- What is the spatial pattern of their occurrence? Specifically, are there "bright spots" of single-frequency microseisms?
- How do the spatially distributed sources vary as a function of time?
- In what azimuth (from the seismic stations) do they occur? Can the sources be located? Do the sources correlate with coastal ocean wave activity or storm positions in the deep ocean?
- What is the composition of single-frequency microseisms?

The available networks include arrays such as LASA, ALPA, and 3-component stations such as LRSM, WWSSN, and SRO stations (we should prepare a list of available data and in what form they are available. Much of the data may be in analog form only and may need digitization). Long term ULF/VLF data from the Columbia OBSS and from the Wake Island hydrophone array provide unique information on temporal variability.

The types of analysis as above should also be carried out on an on-going basis using broadband, three-component seismic networks, which are currently in operation in North America and Hawaii (Figure ?). The advantage here is that the data may be of much higher quality than in old seismic arrays and networks, and may be available in digital format. We propose that analysis of data from currently operating seismic stations be carried out to understand the nature of the source(s) and composition of both single-frequency and double-frequency microseisms for one whole year covering a full meteorological cycle. Of special interest will be intensified analysis during the hurricane season in the Atlantic.

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Directional Ocean Wave Spectra and Groupiness With Spaceborne Synthetic Aperture Radar

(Robert Beal and David Tilley)

Introduction

One of the primary unanswered questions of acoustic noise research concerns the source of the "double frequency" microseisms in the 0.1 to 0.3 Hz frequency region. Noise energy in this region is often highly correlated with the presence of large storms, sometimes located nearby, but often remote. Longuet-Higgins (1950, 1952) and Hasselmann (1963) proposed non-linear wave-wave interaction to explain the frequency doubling, but the detailed character of the actual generation region, and its evolution in time and space, remains obscure. Although the suggested mechanism requires interaction between essentially opposing surface wave systems, the existence of these wave systems has yet to be experimentally confirmed in the open ocean.

There are very few techniques available for estimating the full directional wave spectrum in the open ocean. The most valuable estimate requires a combination of high angular resolution and large spatial coverage. Beyond the spectrum itself, spatial correlation in the surface wave field amplitude and phase (wave groupiness) may provide a generation mechanism that penetrates to the ocean bottom. Thus, a statistical characterization of the actual wave field may be as crucial as the directional spectrum to a better understanding of the bottom microseisms.

There is no ideal sensor with which to address this problem, but of those few techniques available, high resolution ocean imagery collected from a spaceborne synthetic aperture radar (SAR) appears to offer several advantages. Techniques for converting radar backscatter maps to ocean wave height maps and their corresponding energy spectra have progressed rapidly during the last decade. The SAR offers an opportunity to study wave groups simultaneously with their associated evolving directional spectrum, over large areas and through the active regions of storms. The SAR spectrum is radially ambiguous, so it cannot separate opposing wave systems.

Nevertheless, spaceborne SAR spectra when combined with wind-wave models, offer a capability for revealing complex storm morphology which is probably not available with any other technique.

Existing Data Sets

During the last decade, two important spaceborne SAR data sets have been collected: from Seasat in 1978 and from the Shuttle Imaging Radar-B (SIR-B) in 1984. Of these two, the SIR-B data set is by far the more valuable, since it was substantially less affected by Doppler motion, and thus of inherently higher resolution in the along-track dimension. Within the SIR-B data set, a simple pass through Hurricane Josephine around 1630 GMT on 12 October 1984 represents an exciting opportunity to study a rapidly evolving spectrum and simultaneously examine the groupiness question. An additional relevant data set was collected over the Sea of Japan.

Figure 1 shows the hindcast wind field associated with Hurricane Josephine at the time of the SIR-B overpass. The storm was well developed, containing winds of up to 60 knots (30 m/s) and surface wave heights up to 10 m. The associated directional wave spectrum, as estimated by SIR-B, showed a rapid spatial evolution, and exhibited trimodality in its most active region, where the local wind was nearly orthogonal to a pre-existing swell. The associated (enhanced) radar imagery occasionally contains strong evidence of wave groups; it is likely that the statistics of these groups evolved in some way related to the storm morphology. However, the details of the groupiness statistics remain to be extracted from the data.

Strategy for Analyses

The participants in the ULF/VLF Workshop have identified a variety of surface sources forcing the propagation of acoustic waves to the ocean bottom where a characteristic frequency spectrum appears verified by several instruments over a bandwidth of 0.01-40.0 Hz. Distinct spectral peaks attributed to ocean swell, steep wind-waves and foaming breakers seem especially prominent, after frequency whitening; resulting in comparable pressure oscillations near the ocean floor. Pressure spectra are interesting for shape features, such as the double frequency microseism, that are relevant to surface source interactions which enable sound to penetrate to appreciable depth. Surface dynamics also create broadband noise, best observed in low power notches in the pressure spectrum, that is relevant to the random properties of sea states.

Seismic data pose several interesting questions that may be ultimately resolved using ocean bottom seismometers, suspended hydrophones and differential pressure gauges assisted by an orbiting radar altimeter (GEOSAT), synthetic aperture radar (SAR) and various surface instruments documenting environmental conditions at experimental sites. However, prior to sensor deployment, several questions can be explained using SAR data now archived at JHU/APL as a result of our participation in the Seasat and SIR-B programs. For example, the effective resolution of the two SAR systems could be assessed relative to requirements of the seismic community. The SAR data may also be assessed for characteristic spatial and temporal estimates of swell wave and wind sea correlations and coherencies. The linearity of wind-wave and wave-wave interactions should also be addressed in deep and shallow water with considerations of amplitude and phase modulations that would assist acoustic propagation. Ocean wind and wave climate from GEOSAT altimeter data can be referenced to estimate typical meteorological conditions at various sites that have been suggested for ULF/VLF experiments.

The SIR-B Hurricane Josephine data can be interrogated for lumpy swell wave distributions and spatially isolated groups of breaking waves that, in the imagery, appear to collapse together. Fourier domain filtering algorithms will be applied to SIR-B image data to cover a 25 kilometer swath approximately 500 kilometers along the ground track of the space shuttle. Acoustic propagation in the near field (100-400 meter depth) will be considered for such single groups that are identified and random coherence of several spatially correlated groups will be considered for acoustic propagation in the far field. The directional properties of the swell wave field in the SAR data base and its associated wave age estimated from the hurricane wind field may lend support to the existence of group structures. Coherence properties of the wind-wave field over large spatial

dimensions may be usefully modeled using a two-dimensional formulation of Poisson's statement of equilibrium where the divergence of the surface elevation is balanced by the wind friction velocity acting on a distribution of sea facets.

A SIR-B data set over the Sea of Japan of a much more limited spatial coverage is also available. Several groups of coastal breakers have been identified. These sources of acoustic noise correlate spatially from wave to wave within groups. Temporal coherence is more likely from group to group along the coast as individual waves feel the local bottom at common depths. These events can be modeled as amplitude modulated sources of acoustic waves that might converge at a common distance out beyond the surf zone. At least one isolated wave group is also apparent on the shelf well away from the surf zone. Non-linear descriptions of a deterministic nature in the random SAR backscatter process might be relevant. If so, they might offer refinement of our current techniques for restoring azimuth resolution of the SIR-B images of wave groups.

Future Opportunities

Within the time frame of this Advanced Research Initiative, two additional spaceborne SAR missions will occur: the European ERS-1 in 1991 and the NASA SIR-C in 1992. Both sensors will be capable of monitoring the spatial evolution of ocean waves. ERS-1 is designed for a multi-year lifetime, but will be placed in an undesirably high altitude for obtaining good along-track resolution. Also, data accessibility from ERS-1 will likely be difficult in the context of this U.S. Navy research; the protocol for accessing the data can be expected to produce difficulties for foreign investigators, particularly for foreign military investigators.

The NASA SIR-C mission is currently scheduled for May 1992. Although SIR-C is planned for only a several day duration, Navy accessibility to the data would not be a problem. Agreements are already in place between ONR and NASA for acquiring a comprehensive ocean data set, with one tentative site covering a good fraction of the the North Atlantic (see Figure 3).

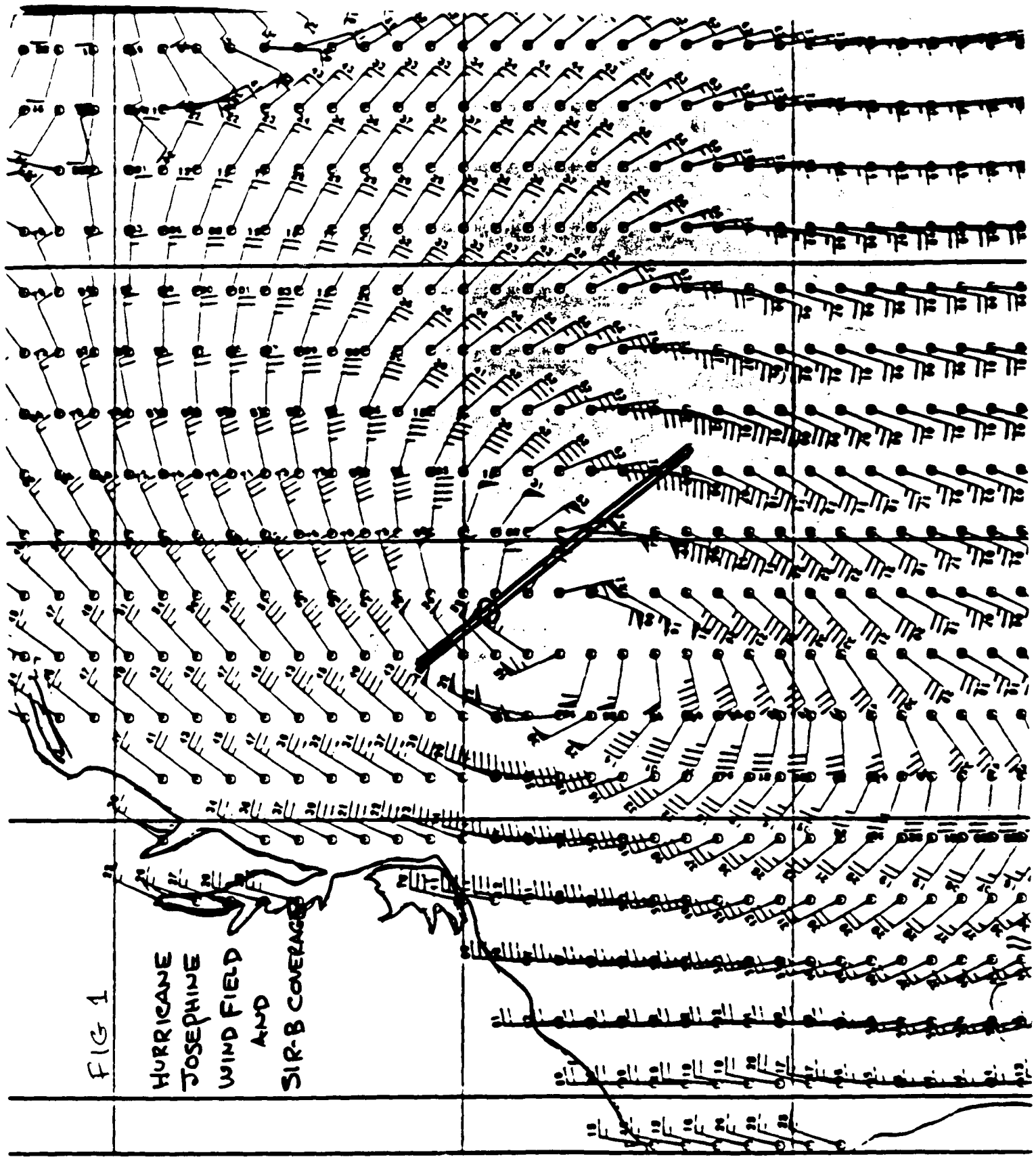


FIG 1

HURRICANE
JOSEPHINE
WIND FIELD
AND
SIR-B COVERAGE

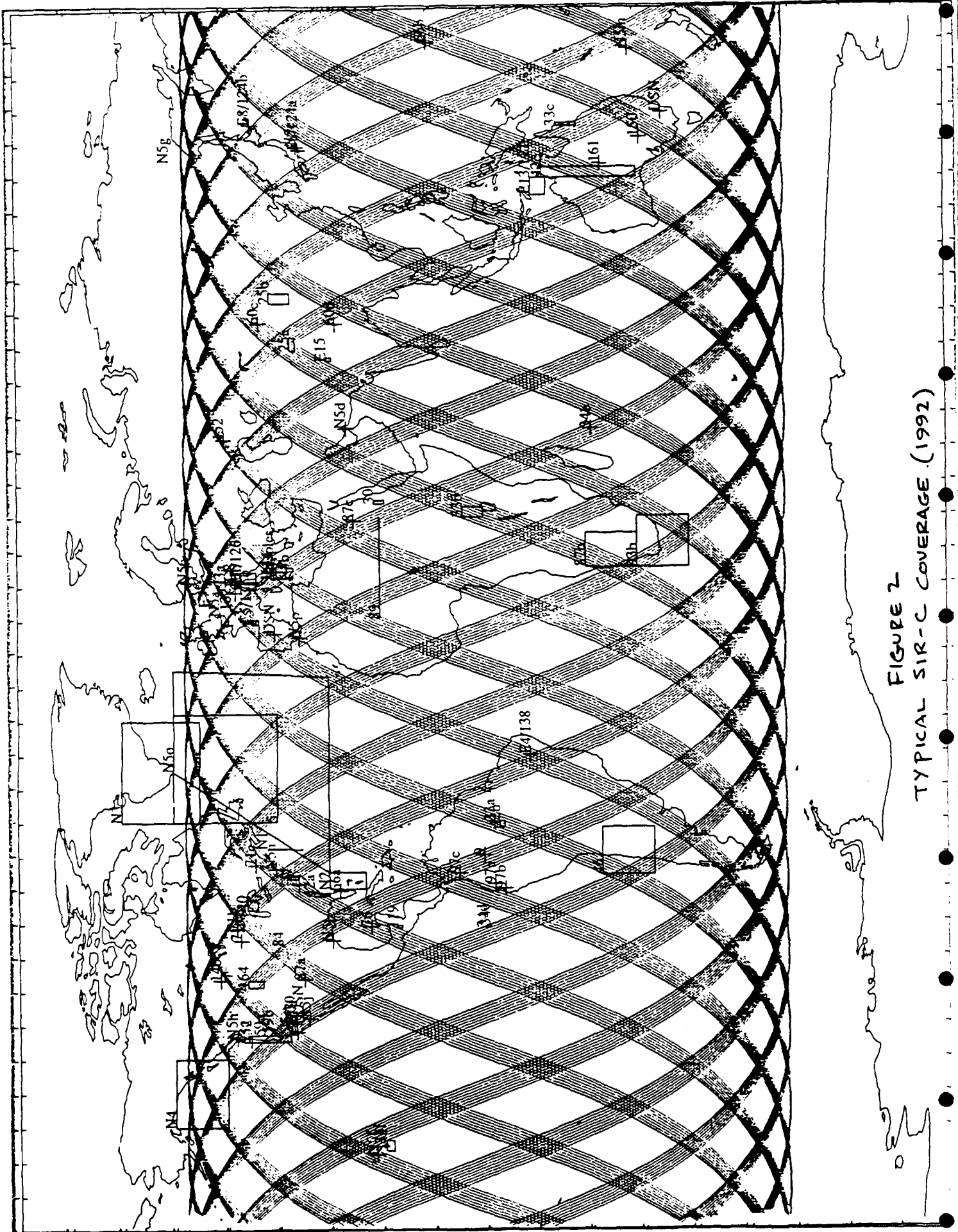


FIGURE 2
TYPICAL SIR-C COVERAGE (1992)

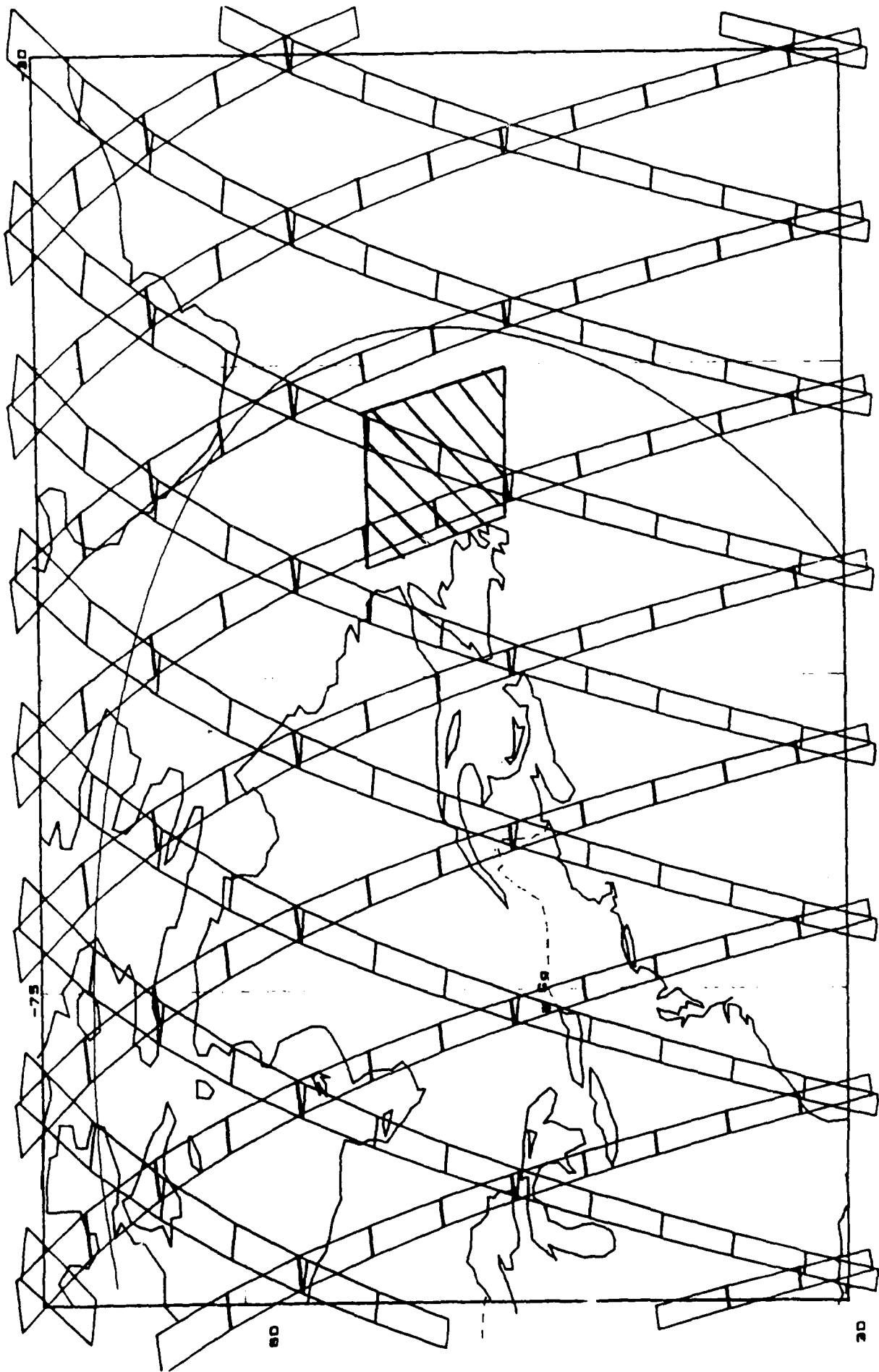
FIGURE 3

EUROPEAN ERS-1 (1991)

LABRADOR SEA WAVE CALIBRATION SITE (3-DAY SAR COVERAGE)

ERS 1

esa



Coherent Array Experiment

(Leroy Dorman, Adam Schultz, and Charles Cox)

The questions central to the understanding of sea-floor noise are (for each frequency range):

1. What is the energy source (wind, swell)?
2. Is the source local, and if not, how does the energy propagate from the source to the receiver?
3. Are Stoneley waves important as a propagation mechanism in the local area?
4. What is the relative importance of scattering?

These questions are amenable to attack by sea-floor arrays much as arrays of seismic instruments have been used on land with great success. The first step in most studies of noise is to partition the signal into different frequency bands or components through Fourier transformation. This partition requires multiple time samples at a single sample location in space and is easily accomplished. Complete decomposition (or synthesis) of a wave-field requires an additional decomposition (or integration) in a one or two dimensional wavenumber space.

This is much more expensive and complicated since it requires a Fourier transform in the space dimension and additional sample points in space are costly. It is the extensive spatial sampling which drives the instrument construction program for this project.

What then, can we achieve through this transformation that warrants this cost?

The prize that wavenumber (or slowness) decomposition offers is that we can immediately identify, given sufficient resolution, the mode of propagation. In particular, we can immediately identify the contribution of the slowly-traveling (50-100 m/sec), high-wavenumber Stoneley waves, faster (3-4 km/sec) Rayleigh waves, the continuous spectrum of broad-band energy radiated down toward the sea floor from the sea surface, and the proportion of scattered energy. An additional benefit is that, in the course of the analysis, we obtain the spatial statistics upon which array performance is based.

Pilot experiments with horizontal scales of ten meters to a kilometer have been conducted by Dorman, Bibee and Hildebrand, and with scales of a few kilometers and tens of kilometers by Cox, Webb, and Orcutt. These observations establish the scales of the arrays required for the several frequency ranges. These experiments have established that the coherence lengths are a few hundred meters in the 0.5-5 Hz band and a few tens of kilometers in the 0.05-0.5 Hz band. In the 0.5-5 Hz band there is no clear evidence of directionality.

The Instrumental Needs

Two-dimensional arrays can be designed for Stoneley waves or Rayleigh waves, but not for both with the number of instruments available because of the large difference in wavelength. In the design of land seismic arrays, it is commonly assumed that the

wave-field is space-stationary so that each vector interelement spacing need only be sampled once. Arrays designed using this philosophy are called minimally redundant arrays and provide the largest ratio of array aperture to minimum interelement spacing. The aperture of such an array is approximately the number of elements times the minimum interelement spacing.

The minimum instrument spacing is established by the sampling theorem as half the shortest wavelength present. Given this, the array will have resolution determined by the array aperture, which should be as large as possible.

The use of two-dimensional arrays in a three dimensional problem presumes the availability of some independent means to regenerate the third component of the wavenumber vector. In this case the identification of downward-traveling energy from the sea surface is vitally important so we will equip several of the new OBSs with vertical hydrophone arrays of a few hundred meters length.

If we take as a target maximum frequency 0.5 Hz and a phase velocity of 1.5 km/s, the minimum sampling spacing will be 1.5 kilometers. With 35 instruments we can obtain an array with aperture of about 50 km.

Environmental Control

To relate the observed field at the sea floor, we will need to know as much as we can afford about the wind and swell conditions at the sea surface during the course of the experiment. A NOAA data buoy can provide wind speed and direction and some wave information when instrumented as a wave-rider buoy. A Cartesian diver equipped with a broad-band hydrophone will assist in identifying energy propagating down from the surface. Some direct way of measuring surface swell directional spectra would be useful.

Experimental Scheduling

This experiment will require the full suite of instruments so we recommend that it be accomplished in the second field season of the ARI.

Experimental Siting

The experiment should be located in deep water in an area of minimal geologic complexity. The weather should provide periods of calm weather as well as storms so that energy partition can be observed under a wide range of conditions. An existing data buoy mooring would be a valuable asset as would be the availability of FLIP to provide ocean wave spectra.

Sites meeting some or all of these criteria are:

1. The far northeast Pacific (Alaska coast)
2. The Wake Island Region
3. The West Coast of the US (off California)

Experimental Costs

1. Ship costs: 30 days each for deployment and recovery
2. Instrument expenses
3. Other observational expenses
4. Analysis costs

Distributed OBS Array for Measuring Broadband Seafloor Noise

(Peter Shearer, Leroy Dorman, and Chip McCreery)

Scientific Questions

- What is the general spatial and temporal variability of seafloor noise over large regions of the ocean floor?
- Can features of the seafloor noise spectrum be correlated with distance to large storm systems?
- What effects do coastlines have on the pattern of noise observed in the deep ocean?
- What features of the seafloor noise spectrum can be predicted from global satellite imaging?

Proposed Experiment

A large scale ocean bottom seismograph (OBS) array should be deployed for an extended time in a region in which large storm activity is typical. No attempt would be made at beam-forming with this array, which would have dimensions much too large to maintain phase continuity across elements. (However, wave type and propagation directions could be obtained from three-component seismic plus pressure sensors in each OBS.) Rather, the idea would be to obtain a general picture of the large scale distribution of seafloor noise over regions up to 2000 km across. Variations in the pattern defined by the amplitude and shape of the noise spectra could be related to conditions at the sea surface (i.e., distance to large storms, coastline geometry, etc.). Satellite observations of swell height and wind speed would form an ideal data set for comparing surface conditions to seafloor noise levels over large areas. Empirical relationships between these data sets might then allow seafloor noise levels to be predicted at other locations for which satellite data are available.

Description

- A large number of OBS instruments should be deployed in order to obtain the best resolution of the seafloor noise pattern. The 35 OBS instruments currently being built under this initiative would be ideal for this experiment.
- The array should be made as large as possible in order to sample a significant fraction of the seafloor. The maximum size of the array is largely determined by the distance that the ship deploying the instruments can cruise during the deployment period. Instrument deployment would be relatively quick since precise locations are not required. If 20 days of ship time were available for deployment, and 20 hours of cruising time (10 knots) available per day between deployments, then an array approximately 500 km by 2000 km would be feasible. Such an array would be 12 OBSs long and 3 OBSs wide with a spacing of 180 km between grid points. An elongated array would be preferred over a square geometry, because it could be deployed perpendicular to the storm track direction in the region and thus enhance

the probability of storms crossing the array.

- The OBSs would be left on the ocean floor for a month or more, depending upon the capability of the instruments. Longer deployments (3 to 12 months) would be preferred due to the possibility of measuring seasonal changes.
- 3 seismometer components and a low-frequency pressure gauge would be sampled at selected intervals during each day. Recordings would be broadband with the upper frequency limit determined by the recording capacity of the instruments and the deployment time. A sample rate of 32 samples/s would allow observations of seafloor noise up to about 10 Hz.
- Observations would be correlated with global satellite and weather data sets. GEOSAT data could be analyzed to obtain surface wind speed and RMS wave height. ERS1 satellite data could provide the wave-number spectra of surface gravity waves through the analysis of Synthetic Aperture Radar (although with a 180 degree ambiguity).
- If the OBSs could be modified to also record amplitudes at 4, 10, and 20 kHz, then the wind speed above each OBS could be directly determined from known relationships between wind speed and the high frequency acoustic energy emitted by breaking waves. Changes in wind direction can also be resolved with this technique due to the increased thickness of the near-surface bubble layer. Such observations could be used to examine possible correlations between wind speed and seafloor noise spectra, as well as to check against the satellite data.
- Analysis of the data would attempt to derive quantitative (probably empirical) relationships between sea surface conditions and seafloor noise levels. These relationships could be used to test various theories of noise generation and propagation. In addition, they could be used directly to predict probable seafloor noise levels at other areas in the ocean, using global satellite data. Thus, the experiment promises to provide practical results that do not depend upon specialized or expensive sea surface measurements.

Location

Several possible regions should be considered for this experiment. These include:

- The Gulf of Alaska. This region is characterized by intense winter storm activity. Advantages include its proximity to West Coast ports and the possible availability of ERS1 Scanning Aperture Array data for measuring the sea-surface wave spectrum. Disadvantages are that the storm activity is so intense and continuous that quiet periods may not be recorded, discrete storm systems would be difficult to isolate and study, and the heavy sea conditions may preclude instrument deployment and recovery.
- Southwest of Mexico. Advantages of this region include proximity to West Coast ports and the large number of hurricanes which cross the region during the summer and early fall. Sea conditions are normally quiet, favoring reliable instrument

deployment and recover. Storms are generally intense and discrete which would allow their effects on the seafloor noise spectra to be easily identified. The contrast between quiet and noisy conditions should be very large. Disadvantages of this region include the less reliable weather information and possible lack of ERS1 SAR data, and the relative unpredictability of many of the hurricane paths.

- Wake Island. Advantages and disadvantages of this region are similar to the Mexican region with the added advantage of correlating the OBS data with existing data from the Wake Island Hydrophone array, and the added disadvantage of its relatively remote location.
- Equatorial Atlantic. Advantages and disadvantages are similar to the Mexican region.

This experiment could be done successfully in many different areas. Other factors from those considered above, such as instrument and ship availability, may favor a particular region.

Logistics/Time Line/Costs

- The experiment will require approximately one month of ship time for deployment and one month for recovery, with a time gap of one or more months between the two legs. The OBSs will need to be shipped to and from the starting and ending ports.
- 1991 would be a good time to conduct the experiment since the 35 OBS instruments are scheduled to be completed and tested in 1990, and ERS1 satellite data should be available in early 1991.
- If instrument availability or fiscal constraints limit the size of the experiment, the number of elements in the array could be reduced. This would best be done by reducing the width rather than the length of the array. For example, 12 OBSs could be deployed in a single line 2000 km long. In this case existing SIO/NORDA instruments could be used.

Acoustic Observations in Severe Weather .1 Hz-200 Hz

(David Farmer, Peter Shearer, George Frisk, Frank Herr, Robert Holman, Chip McCreery, Antares Parvulescu, and David Tilley)

Objective

To acquire simultaneous environmental and acoustical observations for the purpose of establishing the relationship between surface processes (wave-breaking, wind effects, etc.) and the ambient sound field.

Scientific Issues

We do not know how different portions of the sound spectrum depend upon surface wave-field and wind conditions, especially in the range of .1 Hz to \approx 200 Hz. For example, how does rapid veering of the wind contribute to the nonlinear frequency doubling mechanism? Does wave-breaking constitute the primary source above 1 Hz? Long term observations will include a wide range of sea-states; of particular interest is the ambient sound generation at very high wind speeds. Existing data suggest a 'saturation' of the frequency doubling source, but not of the signal at higher frequencies ($>$ 1 Hz). This supports the concept of wave-breaking as the primary generation mechanism above 1 Hz, but there is a need for well calibrated data together with sea surface environmental measurements.

Proposed Experiment and Description

The basic experimental concept proposed here would be combined with other experiment(s) to minimize logistics and increase overall effectiveness of the deployment. For this reason a specific site is not identified at this time. The plan is to install both an environmental buoy and an acoustic array with cable link to shore. The environmental buoy would include measurements of vector wind stress, wave conditions (preferably directional spectra), air and sea surface temperature and atmospheric pressure. Some form of precipitation measurement would be highly desirable (e.g., upward looking radar).

The acoustic array would consist of a vertical line array with uppermost components deep enough to be out of the strong currents of the surface layer. The number and spacing of hydrophones will depend on engineering considerations. A second component of the array will be a horizontal distribution of hydrophones on the sea-floor that can be used coherently. The two components will therefore include both a vertical and horizontal measurement capability. Both acoustic and environmental measurement systems will be connected to the cable, which would be of modern fibre-optic design and would connect the system to a shore station for data recording and transmission.

An essential additional measurement capability is provided by satellite coverage. RMS wave height, and over a large foot-print, wind speed, can be obtained from GEOSAT data. ERS-1 data is also a possibility and would yield directional wave spectra subject to a 180° ambiguity.

Continental Margin Transect - ULF/VLF Experiment

(Joan Oltman-Shay, Spahr Webb, John Orcutt, Rick Adair, Charles Cox, Tokuo Yamamoto, Jerry Smith, Robert Guza, and H. M. Iyer)

East Coast Transect

Ocean surface gravity waves are the primary source of seismo-acoustic noise in the ocean at frequencies below 10 Hz. Any experiment focussed on determining the sources and mechanisms associated with low frequency seismo-acoustic noise in the ocean requires the best possible environmental data on the surface wave field. It is important to emphasize that the detailed surface wave analyses required to make any sense of the complicated non-linear mechanisms causing seafloor noise require extensive instrumentation and are very expensive. We propose to site a major seismo-acoustic experiment off of the North Carolina coast in Fall of 1990 to take as full advantage as possible of measurements derived from both a long-term wind wave directional measurement network at the Army Corps of Engineers Field Research Facility (FRF) as well as an existing ONR sponsored program, Surface Wave Dynamics Experiment (SWADE) to be conducted in this area during the Fall-Winter of 1990-1991. Instrumentation from both the FRF and SWADE would provide wind wave [0 (10 sec)] measurements in the nearshore [0 (10 m) depth] and the offshore [0 (2 km) depth] in addition to aircraft coverage.

We propose an experiment to be conducted along an east-west transect, focused on the changes in the seismo-acoustic noise from shallow water into deep water and from the nearshore region to 600 km offshore. This array will be combined with a terrestrial array to study land-ocean propagation as well. The North Carolina coast is reasonably two dimensional in cross section and so poses fewer problems than many other sites. Numerical modeling of noise generation and propagation (e.g., finite differences) will be straight forward at these frequencies.

Instrumentation in this experiment will consist of a line of six to ten seismometers and/or ULF pressure transducers deployed along a 600 km transect beginning off of North Carolina, extending over the shelf break and 500 km into the basin. The instruments will measure signals across the entire ULF/VLF band (.001-50 Hz) but will be too far apart for coherent processing across most of this band. If additional instruments are available we would propose a small coherent array of seismometers or pressure transducers should be deployed in deep water (250 km offshore) perpendicular to the transect to measure the directional distribution of seismo-acoustic energy. An appropriate scale for the two dimensional seafloor array is probably a 15 km aperture with a 3-5 km inter-element spacing.

A second group of instruments will be a 1.5 km aperture array of seismometers and pressure transducers (DPG) deployed 5 km from the coast in 15 m of water. Elastic waves in the microseism peak (.1-5 Hz) and ocean surface waves in the infra-gravity wave band (.001-.05 Hz) band will be studied with this array. Within this array will be a smaller 250 m aperture array of pressure transducers. The purpose of the secondary array will be to measure the wind wave components. We believe these arrays provide the most important component measurement of infra-gravity and wind wave surface

waves.

The third major component should be 35 three-component seismometers (from PASSCAL) deployed in a 20 km aperture array at a near coastal site in North Carolina. These land based measurements are important to understanding those contributions to low frequency oceanic noise which are associated with distant noise sources.

The long-term FRF array will provide surface gravity wave directional information in the very nearshore (600 meter from the coast). The 15 meters depth array we propose to deploy will provide wave measurements 5 km offshore. The deployment 5 km offshore permits a larger spatial integration of waves propagating from the coast. The SWADE experiment will also include aircraft measurements and wavebuoy measurements at other sites along the Atlantic coast. There is a NOAA wave and meteorological buoy in 2000m, 100 km offshore. This buoy will provide some useful deep water surface wave data, but we believe it will be important to have other means of estimating or observing the noise contributions from the local wind wave field in deep water as well. One proposed solution is to use instruments deployed near the sea surface in the source region to determine that component. The source region at microseism frequencies appears to be the upper few hundred meters. The Cartersian diver instrument (C.S. Cox) is one instrument which can be used to measure noise within this region. Another type of instrument which is designed to measure acoustic noise generated at the surface is D. Farmer's acoustic drifter buoy (p.17).

West Coast Transect

The west coast transect experiment is envisaged as a smaller experiment than the East Coast Transect Experiment. We would propose that this experiment should be conducted in conjunction with the deployment of a major coherent array of seismic instrumentation sited in deep water. This latter experiment is described elsewhere in this document (Dorman, p.25).

As with the East Coast Transect Experiment, a major component of this experiment will be studying the changes in the VLF/ULF seismo-acoustic noise field across the shelf. We expect very different results from the west coast because of the vastly narrower continental shelf. We propose a small number of seismometers and ULF pressure transducers deployed in a line starting near the coast, extending over the shelf, continuing into deep water, and finally joining up with large array of seismic instrumentation deployed for the coherent array experiment.

Surface wave environmental data are essential to understanding the results from these experiments. The major component of this information must be derived from Doppler acoustic measurements from FLIP. The west coast experimental location should be north of Point Conception, to keep the boundary "simple", yet not too far north, to facilitate the operation of FLIP. FLIP can operate only within a few days steam of San-Diego because the vessel is nearing the end of it's design life. The Doppler acoustic measurements from FLIP can provide state of the art surface wave measurements to assist in evaluating direct, local sources, such as the double-frequency mechanism and "group forced" pressure fluctuations bound to groups of surface waves as they propagate. A surface wave array at the coast is also desirable, to evaluate the reflection/generation

of long period (order 100 seconds) free waves, thought to occur there. Unfortunately, the wave climate north of Pt. Conception is too rugged for surf-zone deployments. However, a near-shore array in about 50 to 100 m depth appears feasible, and could provide estimates of the low frequency surface wave variance. Other wave data is available from a series of waverider buoys along the California coast maintained by R. Seymour of SIO.

The six to ten seismic instruments deployed along the transect will measure pressure fluctuations in the band from 0.01 to 50 Hz and/or displacements in the band from about .05 to 50 Hz. We also propose to site 35 three-component seismometers (from PASSCAL) in a coherent array in California near the coast. These land based measurements are important to understanding those contributions to low frequency oceanic noise which are associated with distant noise sources.

ULF/VLF Noise on a Basaltic Sea-Floor and a Sedimented Sea-Floor

(Brian Lewis and Marty Dougherty)

Background

In the frequency band 0.1 Hz to about 2 Hz noise on the sea-floor appears to be dominated by microseisms (Cox *et al.*, 1984, Latham and Nowroozi, 1968). The microseisms are probably generated by the nonlinear interaction of oppositely directed sea surface wind waves at twice the frequency of the waves (Longuet-Higgins, 1950, Cox, this report) and they propagate as Rayleigh waves. These Rayleigh waves are dispersive, attenuate exponentially in the sea-floor, and are sensitive to the shear speed of the sea-floor.

At sea-floor where basalts are exposed, the basalts are typically highly fractured and represent a seismically inhomogeneous medium. At sea-floor where the basalts are covered by sediment, the sediment itself is relatively homogeneous and has a low shear strength, especially at the sea-floor. However, the rough sediment-basalt interface represents a significant inhomogeneity.

On basaltic sea-floor we expect the inhomogeneities to cause scattering which could generate interface waves (Stoneley waves) as well as radiation into the ocean. Although these Stoneley waves are likely to be of small amplitude, incoherent, and rapidly attenuated, they may reduce the coherence of the primary Rayleigh wave.

On sedimented sea-floor with sediment thickness less than a few hundred meters we expect the sediment-basalt interface to cause scattering which likely will manifest as Stoneley waves at the sea-floor. These Stoneley waves will have large amplitudes (because of the low shear strength), will be incoherent, and could greatly degrade the coherence of the primary Rayleigh wave. Latham and Nowroozi (1968) have suggested that sediments could also attenuate the primary Rayleigh wave because of high shear attenuation. Where sediment thickness is greater than a few hundred meters, scattering is expected to be reduced because of the reduced velocity contrast between sediment and basalt.

Based on these considerations we may expect the effect of sedimented sea-floor and basaltic sea-floor on microseismic noise to be significantly different. These differences should manifest themselves in terms of coherence and attenuation of the noise.

Objectives

The primary objective is to measure the coherence and attenuation of noise in the frequency band 0.1 to 5 Hz at a site where sedimented sea-floor and basaltic sea-floor are sufficiently close together (tens of kilometers) that the noise field in the ocean is similar at both sites.

The objective can be met by placing an OBS array over both types of sea-floor and spacing the elements of the array so as to allow the determination of coherence, attenuation and direction of propagation of the noise.

Experiment Description

We require an experiment site where sediment thickness of up to 500 meters can be found within tens of kilometers of relatively smooth basaltic sea-floor. Such a site exists on the east flank of the Juan de Fuca ridge on Cascadia basin.

Two arrays of 12 SNAG/OBS's will be deployed, one array on basaltic sea-floor and the other array on sedimented sea-floor. The array elements will be located so as to best determine the characteristics of Rayleigh wave propagation through the two environments. Preliminary closely spaced array studies by Dorman, Bibee and Hildebrand on sedimented sea-floor have revealed little or no coherence to the shorter wavelength interface waves. Therefore, extremely close element spacing is not necessary for this experiment.

An assumption of this experiment is that the noise field will be dominated by Rayleigh wave propagation. To verify this assumption one SNAG/OBS will be equipped with a 6-8 element vertical hydrophone array of about 400 m length. This array will provide an estimate of the noise wavenumber characteristics in the ocean. It may also be wise to conduct a small (five OBS) pilot experiment prior to the main experiment.

The Cascadia site is bounded on the south by the Blanco Fracture Zone and it can be expected that earthquakes will be recorded. These events will provide coherent sources that can be used at both arrays to independently measure coherence degradation associated with the array site itself. These natural sources will be supplemented by controlled sources to provide environmental and instrument location control.

Theoretical Developments

To evaluate expected array responses to ocean generated noise, models will be constructed which have distributed noise sources at the sea surface as well as point sources. The models will include scattering induced interface waves (Stoneley waves) and attenuation.

Timing

In order to ensure high energy noise periods the main experiment should be conducted in the fall/winter of 1991 and should allow for a two month deployment.

Costs

Ship time costs are not included but should allow for about three weeks of ship time with a large vessel.

Deployment and data analysis	\$400,000
Theoretical modeling	\$100,000

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Arctic Ocean ULF/VLF Noise

(Adam Schultz, Brian Lewis, Fred Duennebieer, and Marty Dougherty)

It is widely thought that noise on the seafloor in the frequency range 0.01 to 50 Hz is dominated by mechanisms related to ocean/wind stress coupling at the sea surface. Other significant noise generation mechanisms include shipping and elastic stresses produced at the sea floor by geologic processes. In the Arctic Ocean the ice covering the sea-surface will greatly dampen the coupling of wind stress and it will also considerably modify the acoustic reflection coefficient from its usual value of -1 , thereby modifying sea-surface acoustic boundary conditions. This unique situation could greatly alter the noise characteristics.

For instance, the double-frequency microseism peak might be shifted in frequency and greatly reduced in amplitude. Similar effects may be seen at lower frequencies (e.g., single-frequency microseisms; background spectral levels within the noise notch, etc.). The ubiquitous 18 dB/octave slope in the (presumed) wind-wave dominated spectral region at frequencies above the microseism double-frequency peak may be absent or greatly attenuated beneath the Arctic ice cover. On the other hand, the existence of microseism spectral peaks may prove to be independent of geographic location and largely independent of local wind stress coupling.

Establishment of the existence, or absence of microseism noise under the Arctic Ocean is the best test of the hypothesis that such noise is strongly coupled to local ocean/wind interactions.

Reduced noise levels are also to be expected in the Arctic since surface shipping is non-existent for much of the year. The largest natural noise sources are expected to be impulsive and related to the breaking of ice, although much of this energy may be concentrated at higher frequencies.

Because it is a low energy environment in terms of physical oceanographic parameters and man made noise sources, the Arctic Ocean provides a unique laboratory to study the physics of VLF/ULF noise generation and propagation. To our knowledge there are no VLF/ULF noise data in the Arctic.

The Beaufort Sea may be a particularly quiet part of the Arctic because it is remote from influences of the Atlantic Ocean and largely decoupled from the Pacific. In fact it may be the least noisy part of the world's ocean. For instance, the integrated internal wave energies appear to be greatly attenuated in this part of the Arctic Basin. This is not only interesting from a noise viewpoint but it offers opportunities for recording geologic sources with high sensitivity and at a location which would provide valuable constraints on deep-earth structure. This is particularly true in light of the current absence of broadband digital seismic coverage within the Arctic.

For these reasons a program of VLF/ULF noise measurements in the Arctic Ocean should be undertaken, with the initial phase of measurements to be made in the Beaufort Sea. Follow-up work includes a Marginal Ice Zone experiment (possibly west of Spitzbergen) - discussed elsewhere in this document; and an array study of VLF/ULF noise generation and propagation in the deep Arctic Basin.

Program Plan

1990: In March-April of 1990 one or two ONR OBSs shall be deployed for one month on the seafloor in the Beaufort Sea in conjunction with a currently planned program of physical oceanography and HF (20 Hz to 10 kHz) acoustics. Simultaneous with this, an instrument shall be deployed on the ice (sea level) concurrent with the seafloor experiment, to evaluate noise generated on the ice (both natural and cultural), and to study temporal and spatial noise non-stationarity as the ice cover relative to the OBS deployment sites. Ancillary measurements deemed to be of value include airgun/watergun seismic bottom profiling, gravity cores and complementary low frequency hydrophone measurements.

This initial deployment is a feasibility study. This does not test the feasibility of deploying and retrieving such instrumentation, since this is already a well tested technology. It tests the hypothesis that ambient noise levels in the VLF/ULF band are attenuated in the arctic, and tests the feasibility of a comprehensive program to measure VLF/ULF noise in the Arctic on a regional basis. This will provide valuable benchmark data for planning the subsequent Arctic VLF/ULF noise experiments.

1991: Observations of reduced VLF/ULF noise levels beneath the Arctic ice in the previous year's initial experimental phase would lead to a follow-up experiment in this year. In the early spring of 1991 approximately six OBSs will be deployed over a wider area of the Arctic Ocean to evaluate the general behavior of noise levels on a regional basis. A program of simultaneous meteorological and physical oceanographic observations, as well as bottom profiling, coring and hydrophone observations will be concurrently carried out during this period.

1992: Marginal Ice Zone OBS studies (tentatively set west of Spitzbergen) to study propagation of microseismic noise from the Atlantic ocean into the Arctic; noise associated with retreat of ice cover, etc. This is described in another section of this document.

1993: Analysis of data collected during previous years of ARI. Design of post ARI field experiments.

Logistic Support

The Applied Physics Lab at the University of Washington has operated a logistic support group in the Arctic for many years. Deployment and recovery of physical oceanography experiments through the ice onto the seafloor is a routine operation involving helicopters, divers, and ice cutting equipment, which are provided as part of the logistic support effort. Cost and scientific effectiveness can be increased by collaborating with other Arctic programs, and costs quite competitive with conventional ship-borne OBS experiments can be achieved. Field efforts are undertaken almost every year.

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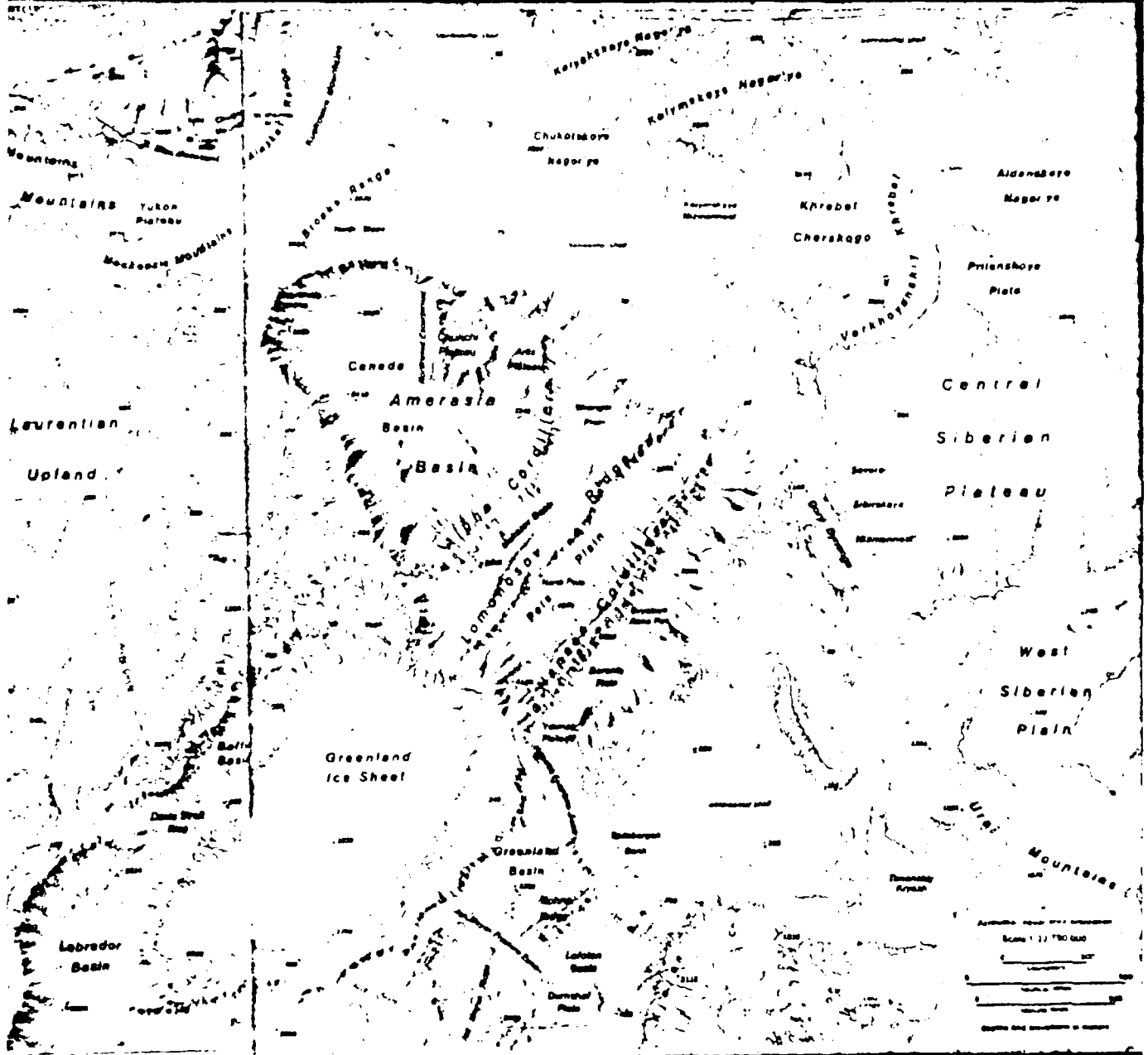
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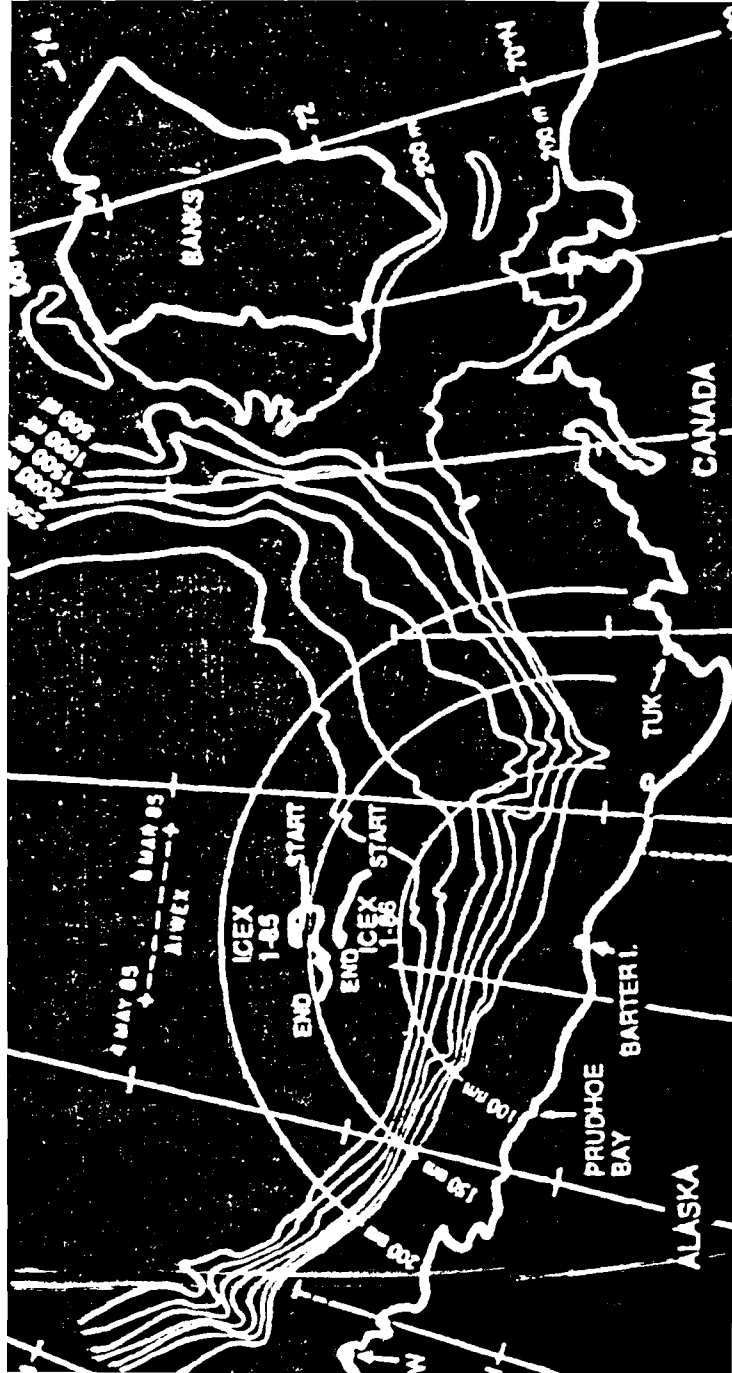
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Арктический район
 Масштаб 1:22 750 000
 Северная широта
 Южная широта
 Долгота восточная
 Долгота западная



Ice Margin Experiment

(Fred Duennebie, David Farmer, George Frisk, and Adam Schultz)

Purpose

Ocean bottom noise in the frequency range from 0.1 to 50 Hz is dominated by noise from waves at the ocean surface. The purpose of this experiment is to determine:

1. How far does this noise propagate in the deep ocean into areas where there are no waves or swell (under the ice)?
2. How does the noise spectrum change with distance from the ice?
3. How much noise is generated by the ice at and away from the margin?

This experiment should be conducted after the deep-Arctic quiet-site experiment to obtain baseline noise data under the ice.

Procedure

A deep-ocean site near the ice margin (probably west of Spitzbergen) will be chosen, and six OBS's containing both ULF (pressure) and VLF sensors will be emplaced through the ice in a line about 70 km long towards the ice margin. The two at the edge of the margin will be placed near each other to ensure that at least one working instrument will be operating in open ocean. These instruments will be emplaced during March, 1991, at the time of maximum extent of the ice, and in such a way that there will be a reasonable certainty that all the instruments will be in open water in September. The OBS's will be configured to record data periodically for six months. With sample periods about every two hours, depending on the amount of data storage. A ship will be sent to recover the instruments in September, 1991 (change to 1992?).

Auxiliary Data

Satellite data on the location of the ice and the sea conditions will be needed for the duration of the experiment. Sea conditions may be estimated from satellite weather maps, and from weather conditions on Spitzbergen. Data taken at the time of recovery should include 3.5 kHz to obtain information on the geologic setting of each instrument.

Logistic Considerations

1. Will need logistic support for ice work in March, 1991.
2. Will need commitment of six or more OBS's for eight months (six months on the bottom) for Spring-summer, 1991 (1992).
3. Will need ship support ice strengthened in September, 1991.

Noise Generated at Active Plate Boundaries and Intra-Plate Volcanoes

(H. M. Iyer, Sean Solomon, G. Michael Purdy, Adam Schultz, and Jan Garmany)

Adequate data do not exist to quantify contributions to the total ULF/VLF noise field made by the active processes occurring in the deep ocean floor: faulting; magma flow injection and emplacement; and fluid flow and jetting associated with hydrothermal fields. In addition, the extreme topographic relief associated with active plate boundaries significantly modifies in some unknown manner the horizontally propagated wave fields from other unrelated distant noise sources. The few existing observations prove that substantial spatial (few kilometers) and short-term temporal (nearly one hour) variations (greater than 12 dB) occur in noise levels in the .5 to 50 Hz band near mid-ocean ridges. (e.g., Riedesel *et al*, 1982; Bibee and Jacobson, 1986; Little, 1988). However, the sources of this noise, the propagation characteristics and how it is controlled by topography are all unknown. Examples are shown in the attached two figures. Figure 1 shows the substantial lateral variability observed in Axial volcano by Bibee and Jacobson, 1986. Figure 2 shows the consistent 12 dB difference in level between two hydrophone near the seafloor separated vertically by nearly 35 m observed at two different locations (Inferno and Hell) both within 100 m of a 50 m high scarp.

A series of modest experiments involving measurements both on and immediately above the seafloor is required to:

- establish the contribution that mid-ocean ridges make to the overall deep ocean noise field;
- determine the mechanisms that generate this noise and define their relative significance; and
- understand the impact that major topographic features have on the distribution and directionality of noise in the deep ocean.

The problem of identifying likely candidate sites is not trivial. Considerable time variability in hydrothermal vent construction, fluid temperature and possible variations in fluid velocity at individual hydrothermal sites has been shown in recent work including time-lapse video, temperature and fluid velocity time series measurements (Delaney, McDuff, McClain, Schultz and others, 1988). Figure 3 shows the variability in temperature and fluid velocity observed over a 50 day period in summer 1988 on a vigorously percolating hydrothermal polymetallic sulfide edifice at the Endeavour Segment, Juan de Fuca Ridge (Schultz, EOS abstract, Nov. 1988). The lower curve is the temperature recorded within 23 cm of the top surface of the sulfide body; the curve above it is temperature recorded 60 cm above the first point within a constrained flow chamber; above this is the background ambient water temperature; and the top curve is the vertical hydrothermal fluid percolation velocity in arbitrary units (currently being calibrated). Such temporal variability suggests that rational experimental design requires monitoring multiple hydrothermal structures, possibly in different stages of evolution, in order to maximize the probability of sampling structures that might produce appreciable levels of

acoustic noise.

Such experiments might take the following forms:

- a large, vigorous hydrothermal field consisting of several clusters of venting sites would be encircled by a ring of hydrophones/geophones - this would establish whether the source of noise was internal to the specific vent field cluster. This might be repeated at one or more locations;
- having unequivocally established that a particular vent field is a significant source of acoustic noise, an array will be designed with the intent of examining the characteristics of the noise, and for establishing the predominant mechanisms responsible for the production of such noise; and
- arrays both on, and off ridge axis can be established to study the propagation characteristics of noise sources associated with ridge crest hydrothermal processes, and sources external to the ridge crest.

A parallel set of experiments in the submarine intra-plate volcanic environment (e.g., Loihi submarine volcano) would also be highly desirable to assess the differences in behavior between inter-plate and intra-plate volcanic, hydrothermal and magmatic processes.

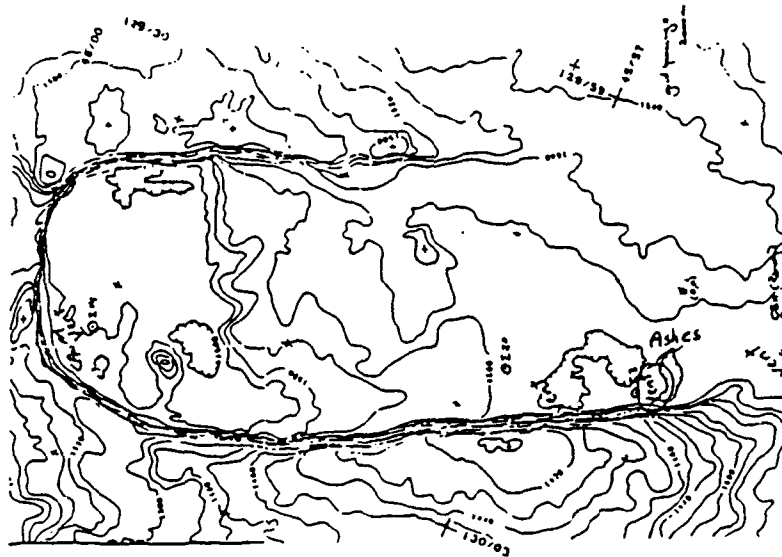
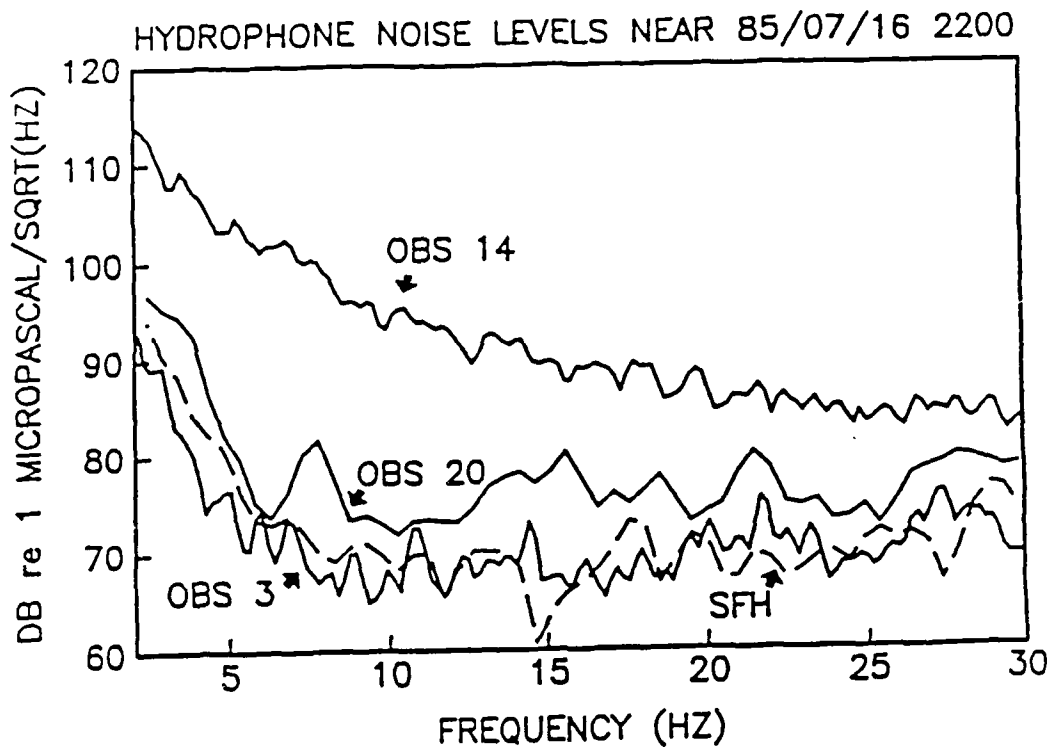


Figure B8a
Noise levels in Axial Volcano from Bibee and Jacobson's 1985 experiment. OBS 14 was in the northern part near a low temperature vent field. OBS 3 was near Ashes vent field.

Figure B8b
Site map of above experiment with Ashes vent field in southwestern part of caldera and CASH low temperature field in northern part.

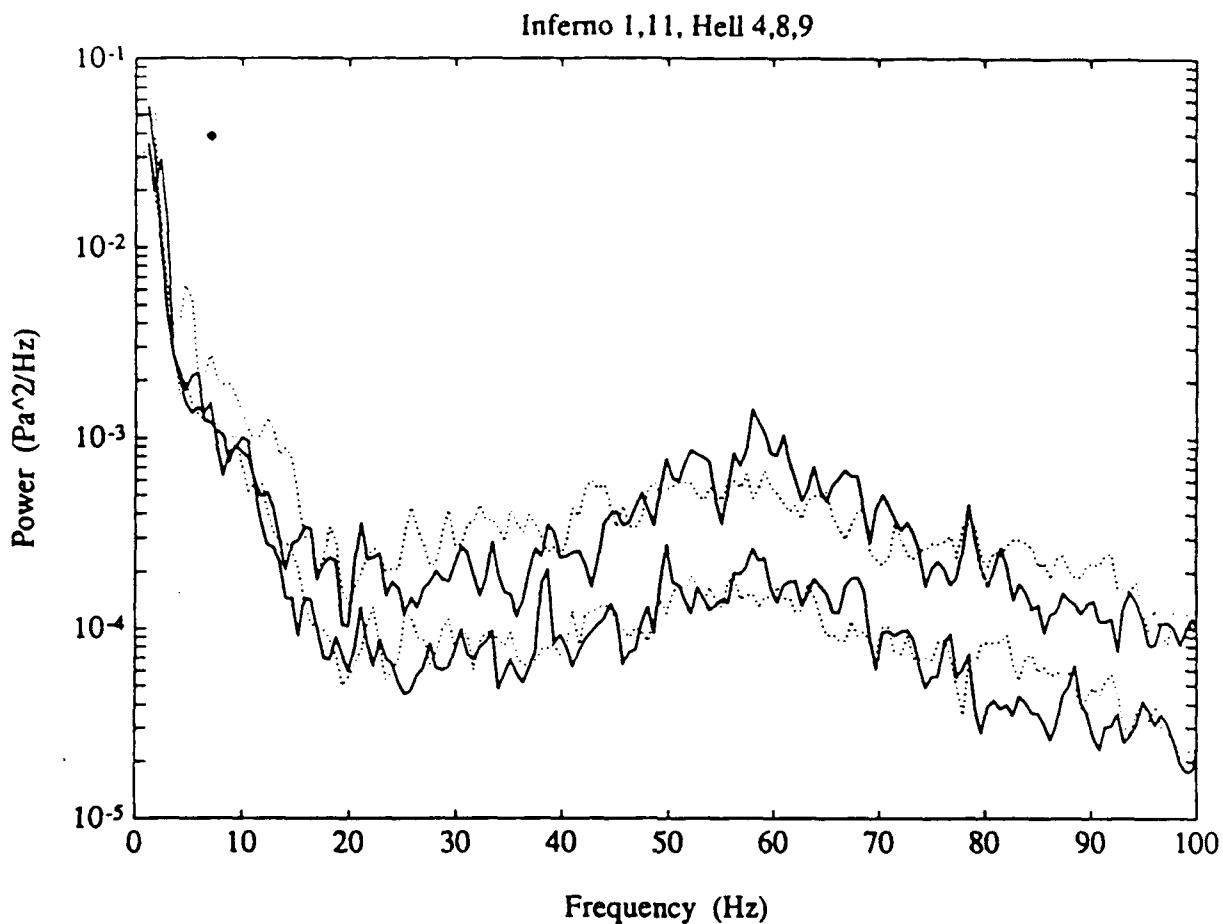
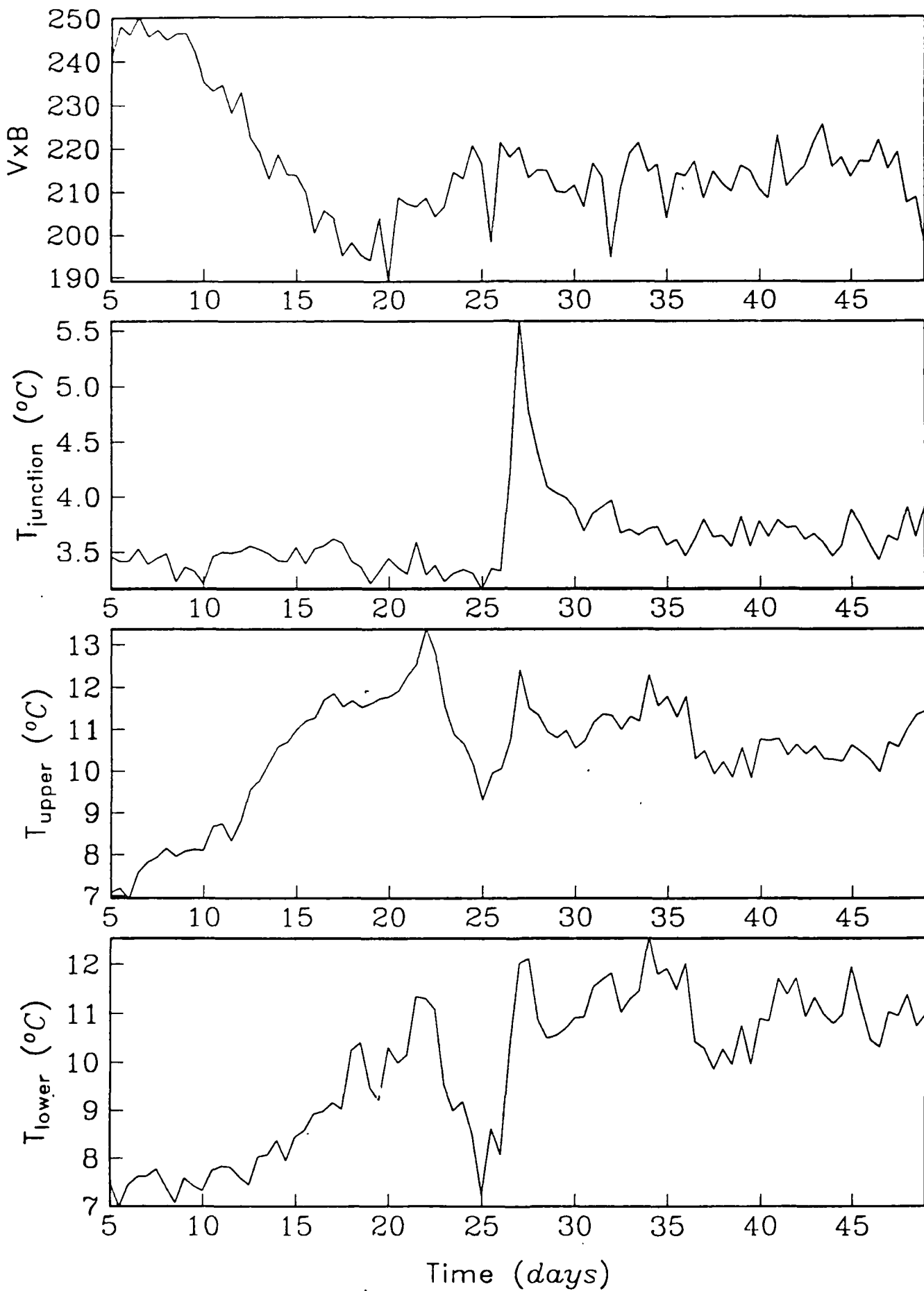


Figure C25

Low noise records used for calculation of signal detection index, average of Inferno records 1 and 11, (upper solid line from upper hydrophone, lower solid line from lower hydrophone) and Hell records 4, 8, and 9 (upper dotted line from upper hydrophone, lower dotted line from lower hydrophone). Note small peak in lower Inferno power spectral density near 38 Hz; this is attributed to jet noise from Inferno vent.

OBVxB Boxcar averages of 134 values (approx. 12 hours)



● Polymetallic Sulfide Edifice Hydrothermal Fluid Temperatures / Velocity, Endeavour Segment

APPENDIX A

ULF/VLF Workshop Talks,
29 November 1988

Pg #	Name	Talks
A-2	R. Jacobson, M. Orr, ONR	ULF/VLF Investigations (ONR-ARI)
A-9	D. Bibee, NORDA	Related NORDA Projects
A-10	R. Butler, IRIS	Oceans and the IRIS Global Seismic Network
A-12	H. Iyer, USGS	International Studies of Oceanic Microseisms and Future Research Directions
A-15	C. McCreery, F. Duennebier, HIG	Ambient Deep Ocean Noise Characteristics, 0.5 to 30 Hz, From the Wake Island Array and the Ocean Sub-bottom Seismometer
A-24	C. Cox, SIO	Cartesian Diver Observations of Double Frequency Pressure Fluctuations in the Upper Levels of the Ocean
A-27	R. Cessaro, W. Chan, Teledyne Geotech	Wide-Angle Triangulation Array Study of Simultaneous Low-Frequency Microseism Sources
A-31	G. Sutton, RAI (Barstow, Carter)	Long-Period Seismic Measurements on the Ocean Floor
A-40	S. Webb, SIO	ULF Seismology and Ocean Surface Waves
A-43	T. Yamamoto, UM-RSMAS (Turgut, Trevorrow, Goodman)	Buried OBS Array Measurements of ULF (.005-1 Hz) Ocean Noise Field
A-50	L. Dorman, SIO (Schreiner, Bibee, Hildebrand)	Deep-Water Array Observations of Sea-Floor Noise
A-69	J. Orcutt, P. Shearer, SIO	The 1989 Low Frequency Acoustic Seismic Experiment
A-77	F. Duennebier, HIG	Coupling
A-79	M. Purdy, WHOI (Dorman, Schultz, Solomon)	An Ocean Bottom Seismometer System for the Office of Naval Research
A-87	S. Solomon, T. Barash, MIT	The Sensor Package for the ONR VLF OBS
A-90	S. Webb, C. Cox, SIO	Electric Field and Pressure Gradient Measurements
A-92	H. Schmidt, W. Kuperman, MIT	Estimation of Surface Noise Source Level From Low-Frequency Seismo-Acoustic Measurements
A-96	M. Dougherty, R. Stephen, WHOI	Finite Difference Modeling of Scattering
A-97	W. Kuperman, J. Perkins, NRL	Three-Dimensional Noise Fields in Complex Ocean Environments
A-99	J. Smith, MPL-SIO	Open Ocean Surface Wave Directional Spectra From Doppler Acoustic Measurements
A-108	R. Beal, JHU-APL	Measuring Ocean Waves From Space: A Summary of the Last Decade
A-118	D. Tilley, JHU-APL	Radar Response to Extreme Sea Slopes and Breaking Waves
A-125	D. Farmer, IOS	Sound Generated by Breaking Waves
A-136	R. Holman, J. Oltman-Shay, OSU	Low Frequency Energy in the Nearshore
A-139	B. Lewis, A. Schultz, UW	VLF Noise in the Arctic Ocean

ULF/VLF INVESTIGATIONS

By R. Jacobson and M. Orr, ONR

Summary: To understand the space-time variability of the acoustic/seismic ambient noise field in the frequency band from 0.001 to 50 Hertz; to identify the noise sources and their noise generating mechanisms and to determine the propagation modes and characteristics for the ambient noise field.

Scientific Objectives: The main scientific questions which are to be addressed are:

- How does the noise field vary with time? What is the three-dimensional directionality of the noise field? What are the temporal and spatial coherence of the noise field? What is the dependence of the noise field on geography?
- What are the sources of the ambient noise? How important are meteorological effects; waves and turbulence in the ocean? How significant are geophysical events, such as earthquakes and hydrothermal venting?
- Along what paths does ambient noise propagate? How does the energy couple between waterborne, subbottom, and interface modes of propagation? How is the energy dissipated?

Although significant progress has been made on these questions, particularly in the VLF (1 to 50 Hz) band and in the "microseism" (0.1 to 1 Hz) band, there are still many critical questions remaining. An informal workshop involving Navy and academic investigators delineated the following unresolved problems in the ULF/VLF arena:

- How important is flow-induced instrument noise, and can it be reduced or eliminated? Can shallow (approximately 1 meter) placement of sensors into a soft bottom significantly reduce this flow noise? How important to observed noise spectra are intrinsic oceanic turbulence and the effect of current speed?
- What are the important VLF noise sources, excluding shipping and biological factors? What is the relationship between wind speed and VLF noise levels?
- What is the contribution of locally-induced "microseisms" [a misnomer - nonlinear wave interaction has been theorized to be the source mechanism] due to scattering from inhomogeneities to the overall microseism noise field? How do these microseisms propagate in the marine environment?
- Current theories state that the frequency of the microseism peak is exactly twice that of surface gravity waves, and that the single frequency component should not be seen as a pressure signal in the deep ocean. Yet these single frequency components are occasionally seen, and their mechanism of generation needs explanation.

Naval Significance: CNR documentation for POM 89 Priority Research Program Topics states that "...acoustic ASW becomes more difficult as the ratio of boundary area to ocean volume increase, and less predictable boundary effects assume greater importance on the acoustic signals. Of prime interest in this area is understanding better the effects of the ocean bottom on sound transmission, especially in the VLF band...". Similarly, the National Research Council's Research Opportunity in Underwater Acoustics states "...decoherence and probabilistic acoustic - in which the stochastic part of the acoustic field that limits the ultimate performance of acoustic systems, especially at low frequencies and with respect to large arrays of two and three dimensions, is quantified." Additionally, the draft Ocean Studies Board recommendations for Marine Geology and Geophysics includes recommendations that "...ONR support research in the VLF area since the experiments designed to understand transmission and interaction

problems for the Navy could yield very important scientific information about earth structure and processes...a number of important gaps in our knowledge at very low frequencies [exist]..."

Approach: In any acoustic ASW program, the objective is to increase signal to noise ratios. Most, if not all, existing programs within the Navy have focused upon increasing the strength of the signal, primarily by using various configurations of arrays to coherently combine signals. Our fundamental knowledge of the source mechanisms, propagation paths and spectral level variability of ambient noise is very limited, yet the ambient noise field is the fundamental limiting factor in all ASW work. This is even more true, as the Soviet submarine quieting program has become more effective. Understanding the ambient noise field will lead, at the minimum, to a determination of the ultimate limits on Naval system performance, or, hopefully if the noise field can be shown to differ from the signal field in some way, to new adaptive processing techniques which will further increase signal to noise ratios.

Our approach to answer the remaining critical scientific problems outlined above consist of three interrelated efforts. First, theoretical modeling of source generation using existing hypotheses is needed, as are development of new theories. Secondly, propagation modeling, although well understood for range-independent conditions, must be improved, particularly regarding rough boundaries, attenuation, shear wave velocities and modal conversions. Third, a major field effort is required involving arrays of instruments to sample the three-dimensional noise field so that measurements of the directionality, coherence and correlation lengths, can constrain the theoretical work.

As a result of recent advances in the development of ultra-low frequency hydrophones and long-period seismometers, we are now capable of making measurements of low frequency noise using three-dimensional arrays. Until very recently, our ability to measure ambient noise was limited to frequencies above 1 Hertz, and particle motions were only measured using geophones on the sea floor. Recent advances in swallow floats now permit low-frequency particle motion measurements to be made within the water column uncontaminated by flow effects. Long period geophone packages can now be built for deployment on the sea floor and possibly down a borehole. The advantages of three component particle motion sensors permit the determination of three-dimensional bearing angles from a single instrument package, whereas several hydrophones are needed to beamform to equivalent angles. Other approaches to observing the ambient noise field include deep borehole instrumentation and shallow penetrating sensors, if funding and technology are available.

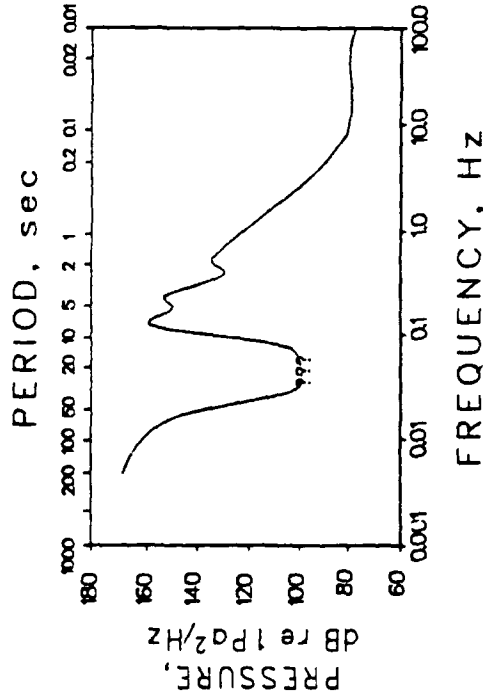
By using arrays of instruments, the directionality, spatial coherence, and spectral variability can be measured and quantified. The temporal variability will be determined by long-term (more than one month) and by multiple deployments. Theoretical work on the generation of ambient noise by meteorological, oceanographic, and geophysical processes and on its propagation will be validated using the array data.

Funding Profile (in thousands of dollars) for the ARI, shared between ONR Codes 1125OA and 1125GG:

<u>FY-89</u>	<u>FY-90</u>	<u>FY-91</u>	<u>FY-92</u>	<u>FY-93</u>
1885	2589	2220	1520	531



ULF/VLF INVESTIGATIONS



OBJECTIVES

- Understand noise field properties (spectral levels, coherence, directionality)
- Identify noise sources
- Understand noise generation
- Understand noise propagation paths

APPROACH

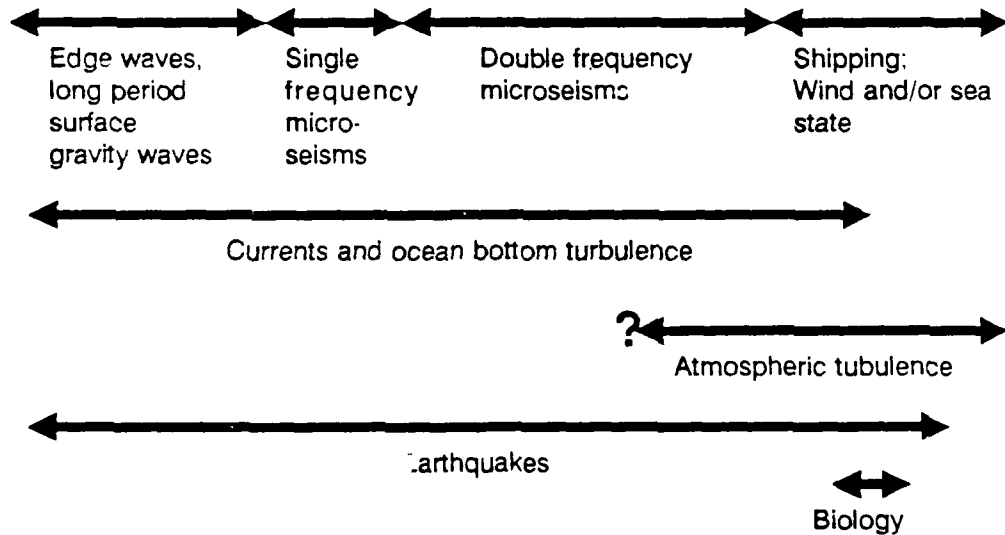
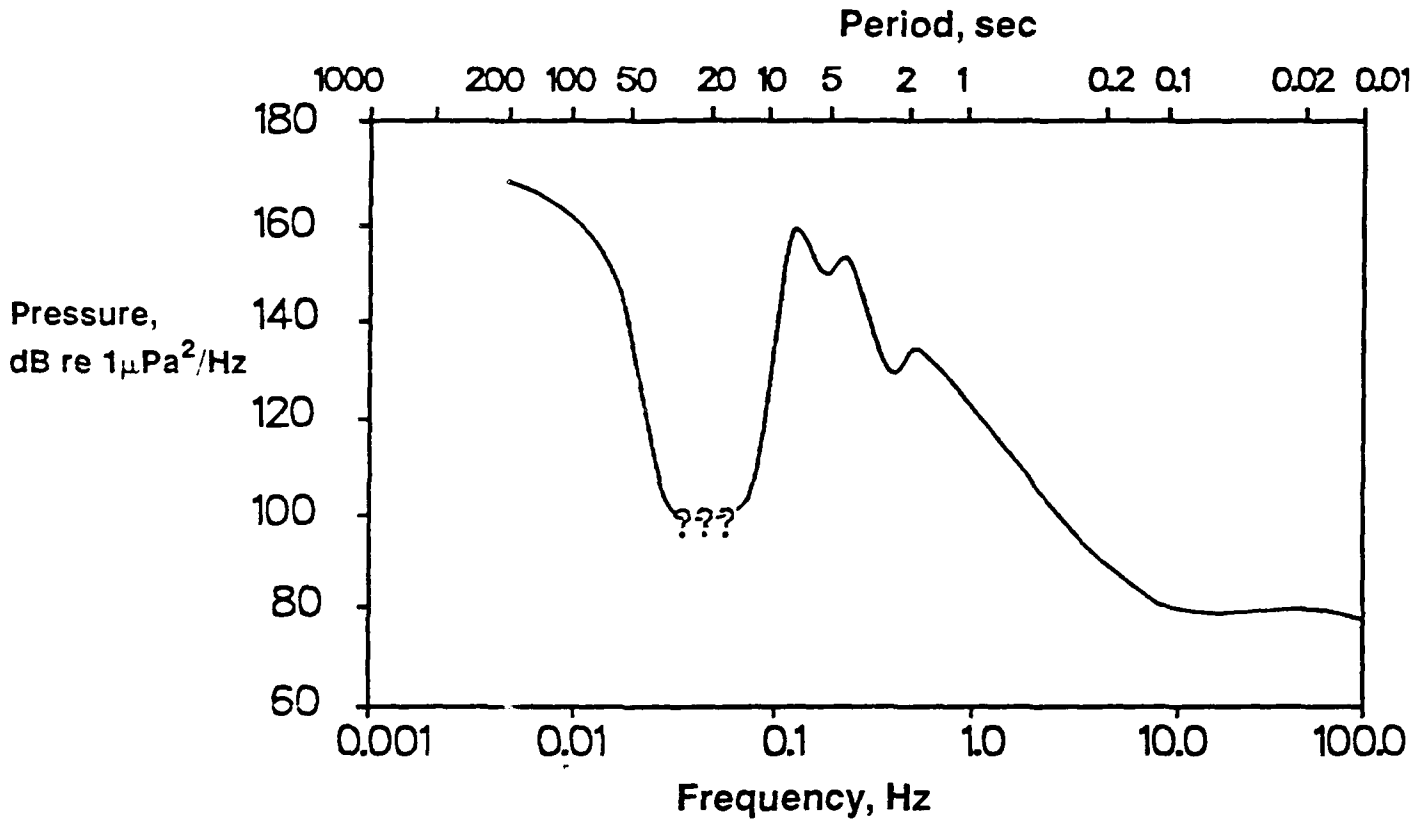
- Improve/develop theory
- Utilize recently developed sensors
- Record noise by 2-D arrays for coherence, directionality
- Concurrent measurements of environment

MAJOR THRUSTS

- Long-term deployments (2-12 months)
- Wide frequency bandwidth
- Joint acoustics/geophysics program
- Ambient noise only



ULF/VLF INVESTIGATIONS



Sources of Ambient Noise



ULF/VLF Investigations

FY 87-89 ONR-SONSORED EFFORTS IN ULF/VLF NOISE

Columbia OBSS ULF data analysis	<i>Sutton, Roundout</i>
VLF array measurements: scattered Scholte waves	<i>Dorman, Hildebrand, et al., MPL</i>
ULF hydrophone/seismometer measurements in shallow water	<i>Webb, MPL; Yamamoto, UMiami</i>
VLF study of hydrothermal vent noise	<i>Purdy, Little, WHOI</i>
ULF hydrophone/electric array measurements	<i>Cox, Webb, Orcutt, SIO</i>
Land ULF noise study of microseisms	<i>Chan, Cessaro, Teledyne</i>

INSTRUMENT CONSTRUCTION

- 5 electric antenna/ULF hydrophone
- 5 ULF hydrophone/directional hydrophone
- 2 prototype OBS in FY88
- 28 OBS in FY89
- 2 atmospheric pressure sensors

Other instruments are available:

- SIO/NORDA OBS
- UMiami OBS/hydrophone

FUNDING PROFILE

<u>FY89</u>	<u>FY90</u>	<u>FY91</u>	<u>FY92</u>	<u>FY93</u>
1885	2589	2220	1520	531



ULF/VLF Investigations

PURPOSE OF WORKSHOP

To address ambient noise scientific issues in interdisciplinary fashion

- acoustics
- geophysics
- physical oceanography
- meteorology
- remote sensing

To identify and prioritize scientific issues

Recommend experiments to address issues

Publish and distribute a white paper to ONR and to the academic community



ULF/VLF INVESTIGATIONS

ONR STRAWMAN EXPERIMENTAL PLAN

FY90:

Gulf of Alaska

- Double and single frequency microseisms
- Location of noise source(s)
- Temporal and spatial variability

Ocean/Continent Boundary

- Edge waves and single frequency microseisms
- Noise source generation and propagation
- Onshore/coastal/offshore components

FY91:

Arctic Environment

- Definition of ULF and VLF ambient noise under ice

North Atlantic

- High energy storm environment – VLF noise generation

FY92:

Juan de Fuca Ridge

- Examine differences between hard and soft bottoms for ULF and VLF noise

Related NORDA Projects - Dr. L. Dale Bibee

The Naval Ocean Research and Development Activity (NORDA) has several ongoing and planned efforts that directly relate to the ONR/CRD ULF/VLF Seafloor Noise ARI. With close cooperation, these programs may enhance and be enhanced by the ARI.

The Seismic Sensor Project is a 6.2 ONT funded effort to explore the use of subbottom seismic sensors in Navy applications. It is a component of the LFASE program which also includes DARPA and OP-21 sponsorship, and participation of Johns Hopkins APL, SAIC, Woods Hole, and Scripps. The cornerstone of the program for FY89 is a deep water experiment involving reentry of DSDP site 418 with an array of seismometers, and supporting OBS and hydrophone array instrumentation. Further work involving penetrator emplaced seismometers is envisioned for FY91.

The Interface Waves Project is a planned FY90 start for a 6.1 core program. The objective of the project is to study the generation mechanisms of interface waves (seismic surface waves) in the frequency range from 0.02 to 10 Hz. We view this program as complementary to the Seafloor Noise ARI in that we are concerned with active sources as opposed to strictly noise, but many of the mechanisms will operate with signal or noise. The primary field effort will be carried out in FY91, and we would welcome joint field efforts with ONR/CRD contractors.

Other programs at NORDA are less directly related. An ARI in VLF propagation is just being completed that dealt with quantitative measurements of energy partitioning in continental margin environments. Work is ongoing in the development of range dependent elastic models using a finite element approach. A 6.2 effort to provide performance prediction models for sonobuoys is underway. Finally we have acquired ULF measurements in conjunction with Navy contractors for 6.2 and 6.3 efforts in the ULF frequency bands.

Oceans and the IRIS Global Seismic Network

By Rhett Butler, IRIS

Introduction

The determination of the structure of the Earth is limited by the locations of seismic observatories, which presently are sited on the continents and a few islands. Thus the oceans, which comprise about two thirds of the planet, are dramatically under sampled. This very non-uniform sampling of the planet is manifested as spatial-aliasing in earth studies. To some extent seismologists have worked around these problems by studying waves which propagate across the oceans (i.e., surface waves) or remotely sample the oceans (i.e., multiply reflected body waves like ScS_n and SS) but such extrapolations from the continents into the ocean interiors cannot replace *in situ* measurements. Siting seismic stations on islands improves the global coverage. However, islands are not in general representative samples of the oceans and vast areas of the oceans are without islands. For the seismic study of the Earth on a planetary scale, there is no substitute for deep-ocean seismic observations.

Beyond the fundamental question of the earth's structure, the non-uniform seismic coverage of the planet poses problems in the study of seismic sources, both natural and man-made. The present continental and island siting of seismic stations leads to large gaps in coverage, and the lack of adequate coverage often introduces substantial uncertainty in the source mechanism of events. In California, for example, earthquakes are well covered over about 180° of azimuth with seismic stations from Alaska to Central America. However, only sparse coverage in the oceans is available to sample the seismic energy radiating to the south and west. As in the case with earth structure, the gaps in coverage are a fundamental limitation, and there is no acceptable recourse short of deep-ocean seismic observations.

For both earth structure and seismic source studies high-quality, 3-component, broad-band seismic data from the deep oceans are important. Borehole deployment in the marine basement is ideal in terms of platform stability, low-noise, and good coupling. Long-period (1 hr. to 10 sec.) gravity and broad-band pressure sensors can complement the seismic instrumentation. High-frequency (10 to 50 Hz.) instrumentation is important in the study of wave propagation within the oceanic lithosphere.

High-resolution seismic studies of the Earth would be advanced by ocean bottom or sub-bottom observatory sites throughout the oceans, and these advances would affect many fields in the geosciences. The seismic structure of the Earth's interior relates to problems of geodynamics and mantle convection. Detailed tomography of mantle plumes has ties to mantle geochemistry and petrology. The structure of the core-mantle and the inner-outer core boundaries bear important relationships to geohydrodynamic models of the Earth's magnetic field. A detailed seismic structure of the oceanic lithosphere, both elastic and anelastic, will strongly influence models of the evolution of the lithosphere. Intrinsic attenuation of seismic energy is a thermally activated process and thus knowledge of seismic attenuation within the Earth provides one of the best constraints on temperature variation in the Earth.

Seismology gives the Earth scientist the capability of mapping the anelastic and elastic properties of the planet. It is one of the few sciences in which an active experiment can be applied to investigate a feature deep within the Earth. The capability to move into the oceans greatly broadens this inherent capability present in seismology today.

Present Course

The Incorporated Research Institutions for Seismology (IRIS) - a consortium of fifty-seven universities in the United States - has embarked upon the establishment of a new global digital seismographic network of 100 stations evenly distributed around the planet. IRIS is funded through the National Science Foundation Continental Lithosphere program within Earth Sciences, and works in partnership with the US Geological Survey in the undertaking.

The IRIS design goals for the network calls for very-broad-band borehole and vault seismometers supplemented by optional high-frequency and low-gain instrumentation where appropriate. Three component broad-band data is logged continuously with 24 bits of resolution, and optional channels are recorded at 16 bits resolution in a triggered mode of operation. Rapid data availability is important, and is implemented by use of telephone dial-up capabilities on the station processors and satellite telemetry access where cost effective. Satellite clocks are used for time reference. All equipment is state-of-the-art off-the-shelf technology.

The current IRIS program tackles the problem of oceanic coverage through siting seismometers on available islands. This direction is based upon expediency. Ocean-bottom installations require a significant research and development effort which cannot be justified within the current constraints imposed by NSF funding to the program. Island sites can be deployed with available technology. This is not to say that island sites are superior or inferior to ocean siting. The relative merits of oceanic versus island siting have not been well demonstrated across the broad frequency band of seismic interests and this is an area of research which would have a substantial impact upon future plans for the deployment of seismic instrumentation for the broad coverage of the planet.

For ONR Workshop Nov 29 - Dec 1, 1988. Austin. Texas

International Studies of Oceanic Microseisms
and Future Research Directions

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Introduction

The study of oceanic microseisms (OM) in the frequency range 0.1-1 Hz has been one of the largest and most exciting fields of seismological research since the beginning of instrumental seismology. In terms of participation by eminent scientists and the number of papers written, OM remains on par with topics such as *earth structure* and *earthquake prediction*. An additional attraction of OM is that it embraces two other major disciplines, namely oceanography and meteorology. In spite of these facts, interest in oceanic microseisms waned noticeably about two decades ago, mainly because, its chief application to storm and wave forecasting was superceded by the arrival of weather satellites and rapid strides in oceanography and meteorology. However, the fascination of OM remains in a few countries (USSR and some European countries) which seem to be able to afford the luxury of pure research. Unfortunately U.S.A. does not seem to be one of those countries. Therefore, I am very happy that this workshop has been convened by the ONR and hope we will get an opportunity to address some of the unanswred questions in OM.

Unanswered Questions

Clearly the most important unanswered question in OM is *do storms generate OM in the deep ocean directly beneath them?* The sparse work done in the world during the last two decades in OM seems to be mainly addressed to answering this question. Most of the studies use existing seismic networks to study amplitude/spectral variations in relation to storm and ocean wave activity. A few deep-ocean measurements have also been carried out in the USSR and U.S.A.

The second question relates to the *composition of microseisms*. It is clear that fundamental and higher mode Rayleigh waves constitute a significant part of OM. The question is *are body waves and Love waves also present and if so what are their mechanisms of generation?*

The last question relates to the interpretation of amplitude/spectra of OM simultaneously recorded on land and in under the ocean. Specifically, *what exactly is generated in the deep ocean and near the coast, how does it propagate, particularly across the continental barrier, and how much of it gets through to land?*

A Future Solution

As a member of the IASPEI (IUGG) Member of the Commission on Microseisms, I have participated in two workshops on microseisms during the past eight years. I find that scientists in USSR and Europe are making a brave effort to address the above questions in OM using available seismic networks and state-of-the-art analytical techniques. However, in my judgement, no conclusive results have emerged because the studies represent isolated pieces of individual research rather than mission-oriented programs. The solution therefore lies in a co-ordinated international effort involving seismologists and oceanographers to tackle the problem of generation, composition, and propagation of microseisms, using modern instrumentation and analytical techniques. The following progressively complex steps are recommended.

1. LITERATURE SEARCH to produce a state-of-the-art status report on available data and results.

2. USE OF EXISTING INSTRUMENTATION. In a few areas of the world where seismic arrays and ocean-wave recorders are available, carry out a systematic but intensive study of Oceanic Microseisms for two or three years. An obvious effort is the use of the extensive instrumentation available in the U.S.A.
3. INTERIM EXPERIMENTS. In one or two areas where fast-moving cyclones occur and can be tracked accurately and the associated ocean-wave fields can be monitored, operate seismic arrays (at least tripartite arrays) to "catch" several storms.
4. FINAL EXPERIMENTS. Based on the experience from (2) and (3) carry out one or more high-resolution experiments in locations where fast moving cyclones occur, by operating large-aperture, dense, portable seismic arrays with supporting ocean-bottom seismometers and ocean-wave recorders.

Ambient Deep Ocean Noise Characteristics, 0.5 to 30 Hz, from
the Wake Island Array and the Ocean Sub-bottom Seismometer

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Introduction.

Data from an array of hydrophones near Wake Island in the northwestern Pacific (Fig. 1) and from an ocean subbottom seismometer (OSSIV) deployed down a DSDP drillhole off the Kuril Islands (Fig. 2) display similar ambient noise level dynamics. At Wake, three-minute-long ambient noise samples were taken once every six hours over a one-year period from each of four hydrophones. These data were compared with cotermporal surface windspeed measurements made at Wake Island by the National Weather Service. In a similar effort, the ambient noise over a sixty-four-day period measured by OSSIV was compared with the surface atmospheric pressure differential (an indicator of windspeed) measured from satellite weather maps. Both studies show the same, distinctive patterns of noise level variation with windspeed, and the nature of these patterns suggests that the noise is more directly related to ocean surface wind-driven waves.

0.5 to 6 Hz.

Between about 0.5 and 6 Hz, noise levels increase with increasing ocean-surface windspeed until a well-defined saturation level is reached (Figs. 3 and 4). This saturation level appears to be related to a corresponding saturation of wind-driven waves on the ocean's surface (Fig. 5). The spectral slope of the saturated noise, $f^{-4.5}$, is approximately twice that of the saturated waves, $f^{-2.5}$; and the frequency range of the noise, 0.5-6 Hz, is approximately twice that of the waves, 0.2-2.5 Hz. These features are characteristic of noise generated by non-linear wave-wave interactions. This mechanism for ocean bottom noise requires two opposing wavefields interacting coherently with each other over a region with dimensions comparable to the water depth, and the amplitude of the noise generated is proportional to the product of the heights of the opposing waves. Wavelengths of wind-driven waves corresponding to noise at these frequencies range from 40 m to less than 0.5 m. These wavelengths are very small compared to the water depth of the data, 5.5 km, and any coherent interaction of wavefields over these dimensions must be only a small fraction of the total amount of interaction. Also, very little variation of the noise level is seen at the different times of noise saturation (Fig. 6), indicating that the height of each set of opposing waves is either constant and

the same during all saturation periods, or that an averaging of interactions with a relatively constant result is taking place. Alternately, a different, and as yet unknown, mechanism may be responsible for the noise. Simultaneous measurements of the sea-surface directional wave spectrum at short wavelengths and ocean-bottom noise will be necessary to resolve these questions.

Noise levels at Wake between 2 and 6 Hz are saturated roughly two-thirds of the time. This probably corresponds to an equal time period of saturation of the wind-driven waves. Since corresponding wind-waves are very short (< 3 m) at these noise frequencies, they are easily saturated in even a moderate wind. The mean windspeed at Wake is 6.3 m/s (14 mph), a typical value for the tradewind belt. We believe that this saturated noise spectrum is observable most of the time in all of the world's deep oceans and we call it the "holu" spectrum after the Hawaiian word for deep ocean. The constant holu spectrum could be very useful for future ocean experiments, as a reference for in situ calibration of instruments, and possibly as an energy source for studying properties of the bottom through propagation effects. In addition, this spectrum may provide a simple way to measure the general state of the ocean's surface.

4 to 30 Hz.

From 4 Hz to at least 30 Hz (the upper frequency limit of our measurements), noise levels vary little with windspeed until windspeed exceeds about 7 m/s (14 mph) (Figs. 3 and 4). For windspeeds above this threshold, noise levels increase regularly and equally across the entire frequency band, and the spectral shape of the noise is white. It is distinguished from the noise at lower frequencies by its markedly different spectral slope, and by its non-saturation. It has been proposed that this noise is caused by the action of breaking open-ocean waves. Measurements of the onset and amount of breaking waves and simultaneous measurements of deep ocean noise at these frequencies should be made to verify this hypothesis.

Propagation.

Both types of deep ocean noise are affected to some extent by propagation effects. Factors such as water depth, sediment thickness and character, bottom roughness, and water-column velocity profile may all play a roll in determining the absolute levels and spectral shape of the noise. Also, the sensor location within the water-sediment column is an important factor. Additional measurements of wind-waves and noise should be made in a variety of environments, and theoretical modelling of the propagation effects should be made once more is known about the character of the noise source.

FIGURE CAPTIONS

Figure 1. This map shows hydrophone locations of the Wake Array relative to belts of shallow seismicity in the northwestern Pacific. Hydrophones 71-76 are located on flat ocean bottom at about 5.5-km depth. Hydrophones 10, 11, 20, 21, and 40 are at about 850 m depth, the approximate depth of the Sound Fixing and Ranging (SOFAR) channel.

Figure 2. This map shows the location of the Ocean Sub-bottom Seismometer (OSSIV) that was deployed down Deep Sea Drilling Project (DSDP) drillhole site 581. The water depth at the site is approximately 5.5 km, and the sensors were locked in a position 350 m into the sediments, just above basement basalt.

Figure 3. Average spectra from Wake hydrophone 74 for eight windspeed ranges. The saturation slope, with a value of about $f^{-4.5}$, is marked by the arrowhead.

Figure 4. Horizontal geophone noise levels from OSSIV during the onset of a storm. Each line is the noise spectrum measured at six hour intervals, with windspeed increasing from spectrum 1 to spectrum 5. The saturation slope has a value of $f^{-5.5}$ in particle amplitude, equivalent to $f^{-4.5}$ in particle velocity or pressure.

Figure 5. The saturation level of ocean wind-driven wave heights based on several groups of data compiled by O. M. Phillips.

Figure 6. Each of the 1460 noise level measurements at 2.34 Hz made over one year at Wake from hydrophone 74 plotted as a function of the windspeed at Wake. The saturation level is clearly visible at around -5 dB relative to 1 microbar per root Hz.

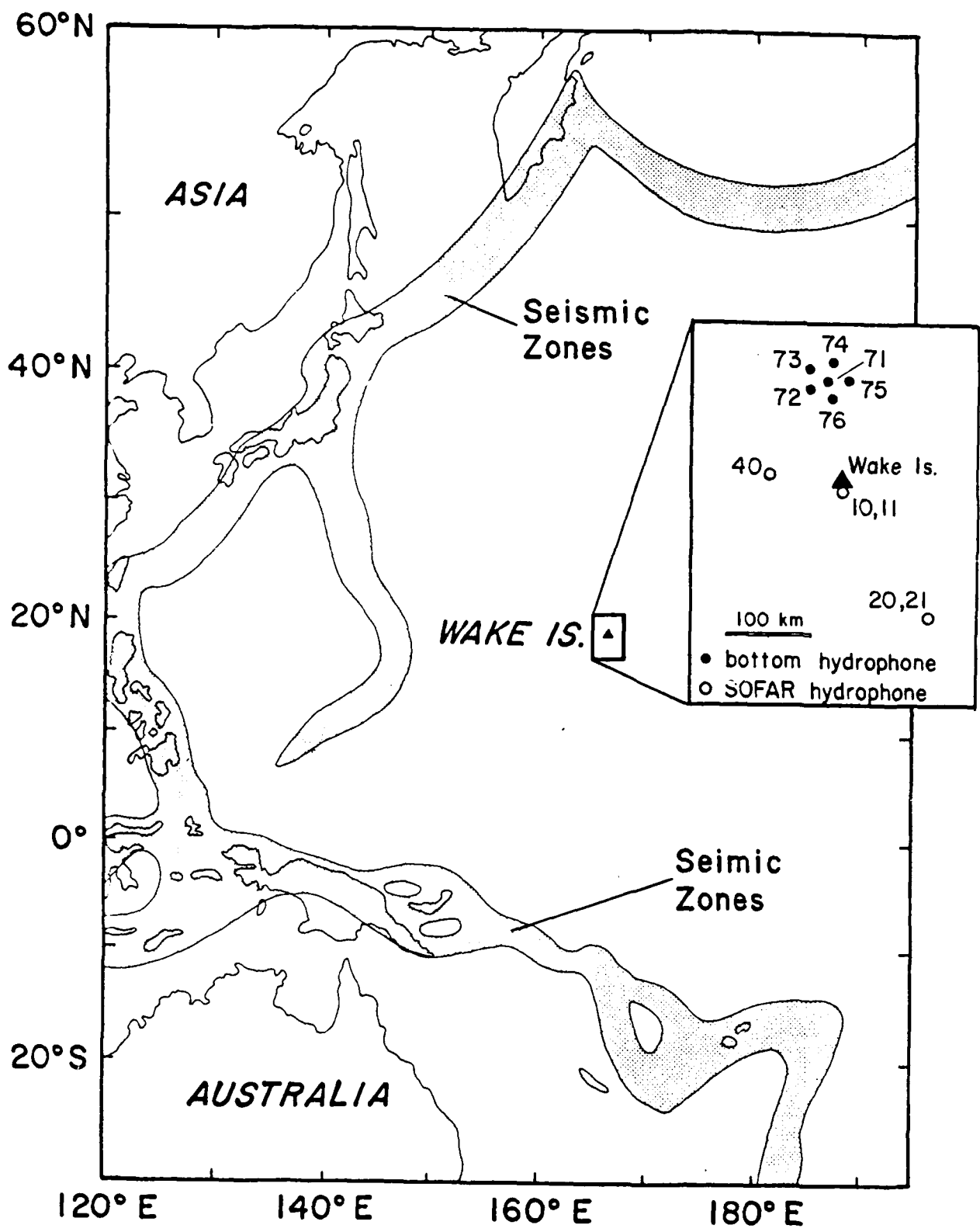
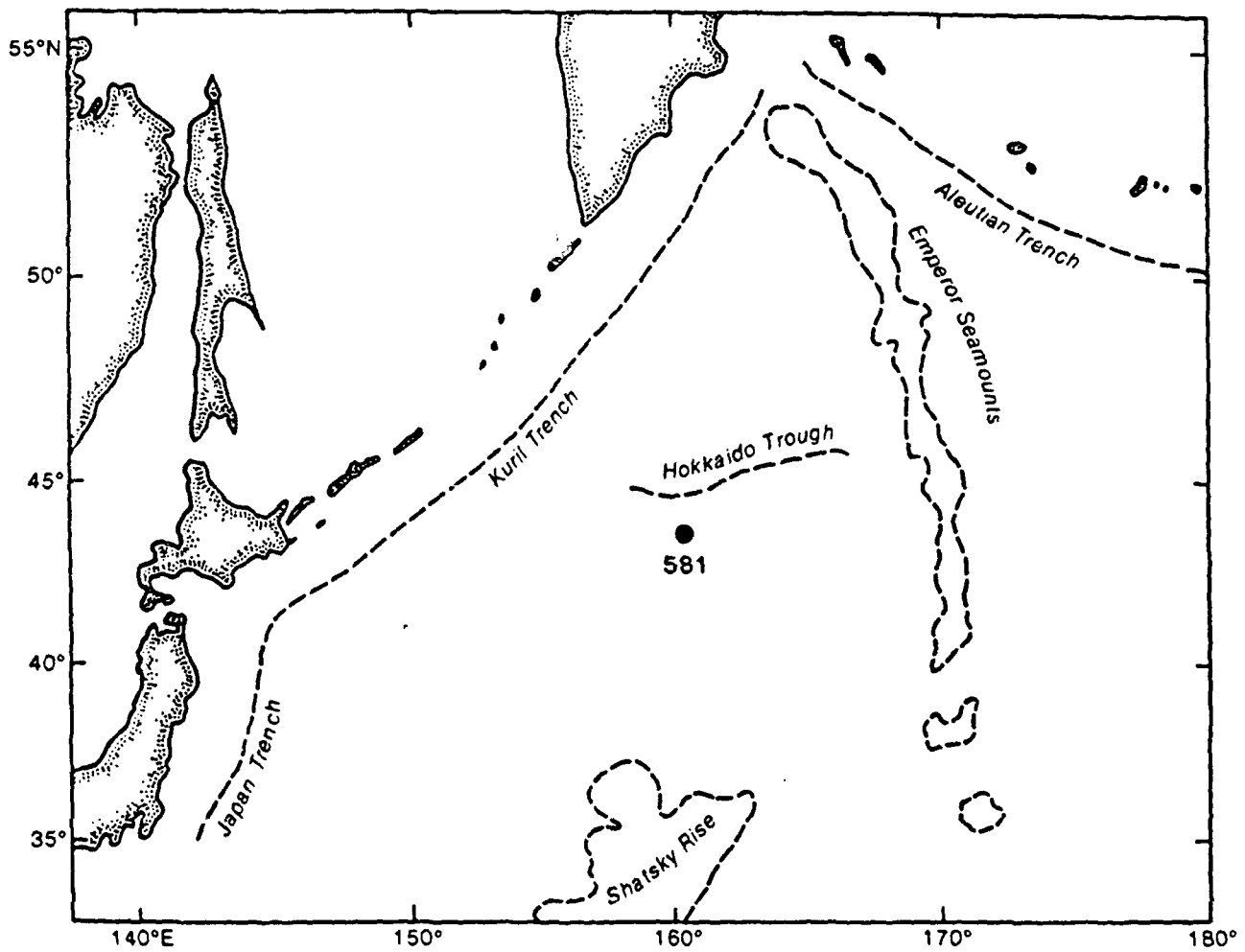
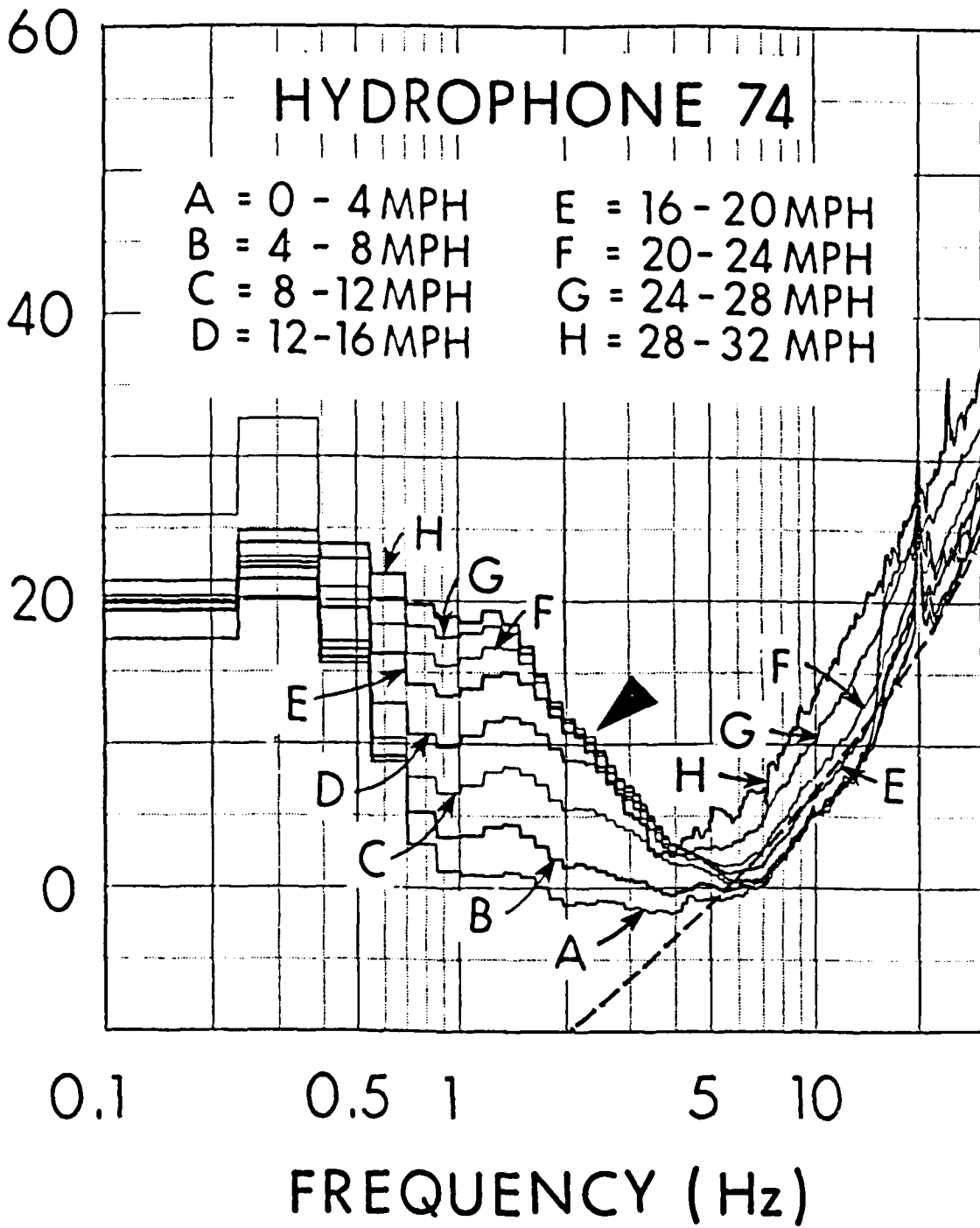
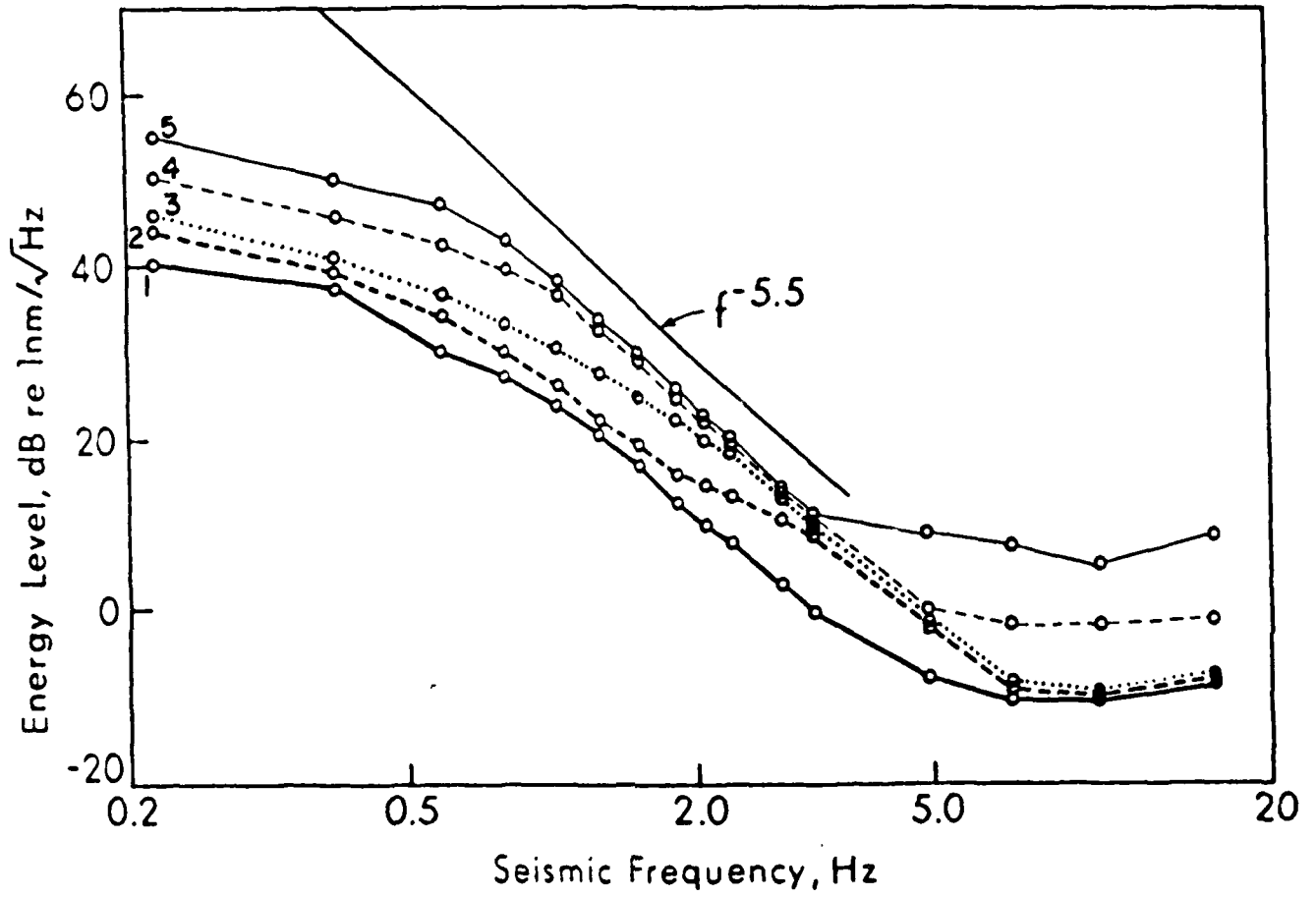


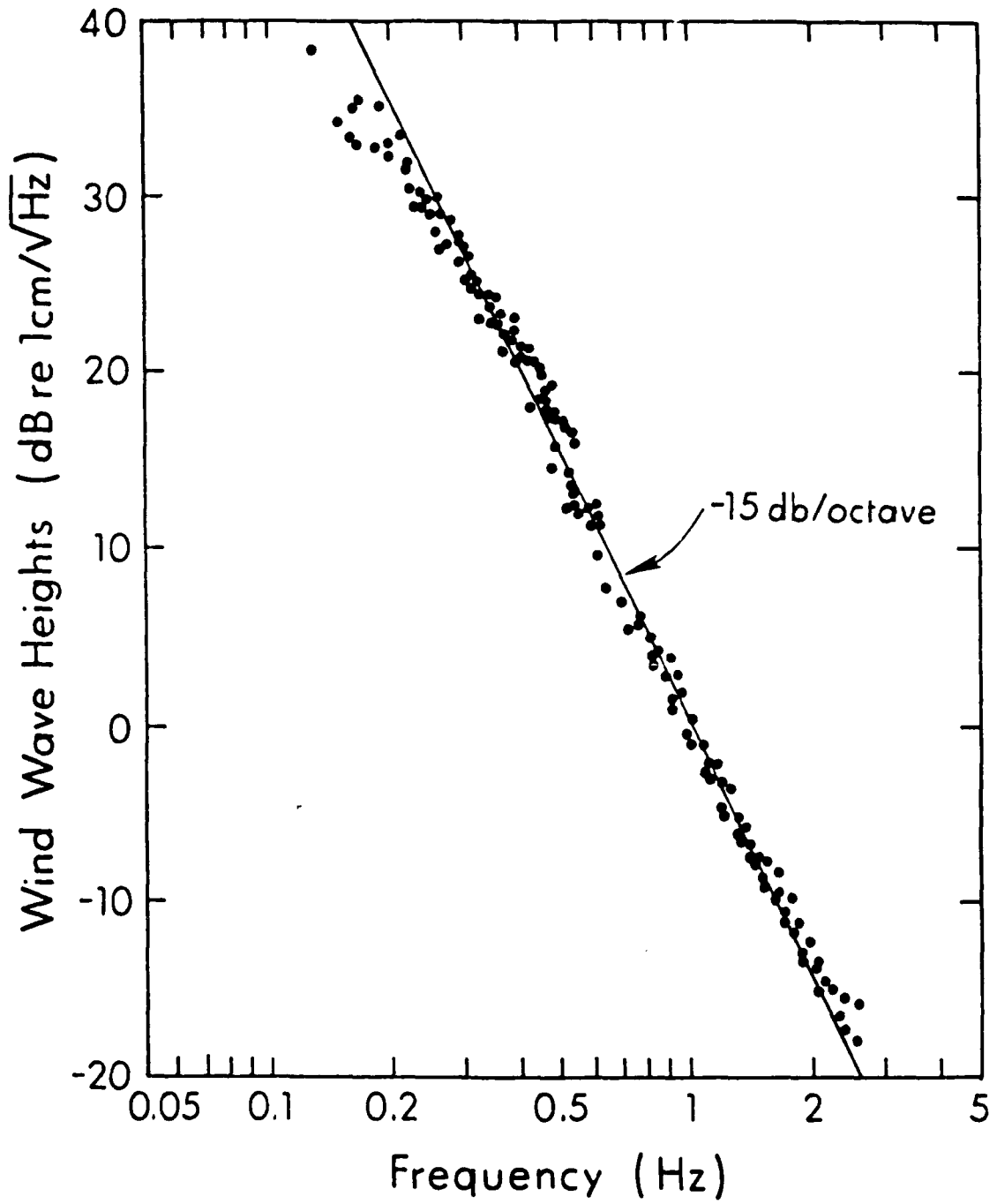
Fig. 1



dB re 1 MICROBAR (0.156 Hz BANDWIDTH)
DATA ROTATED COUNTERCLOCKWISE ABOUT 1 Hz BY 18dB/OCTAVE

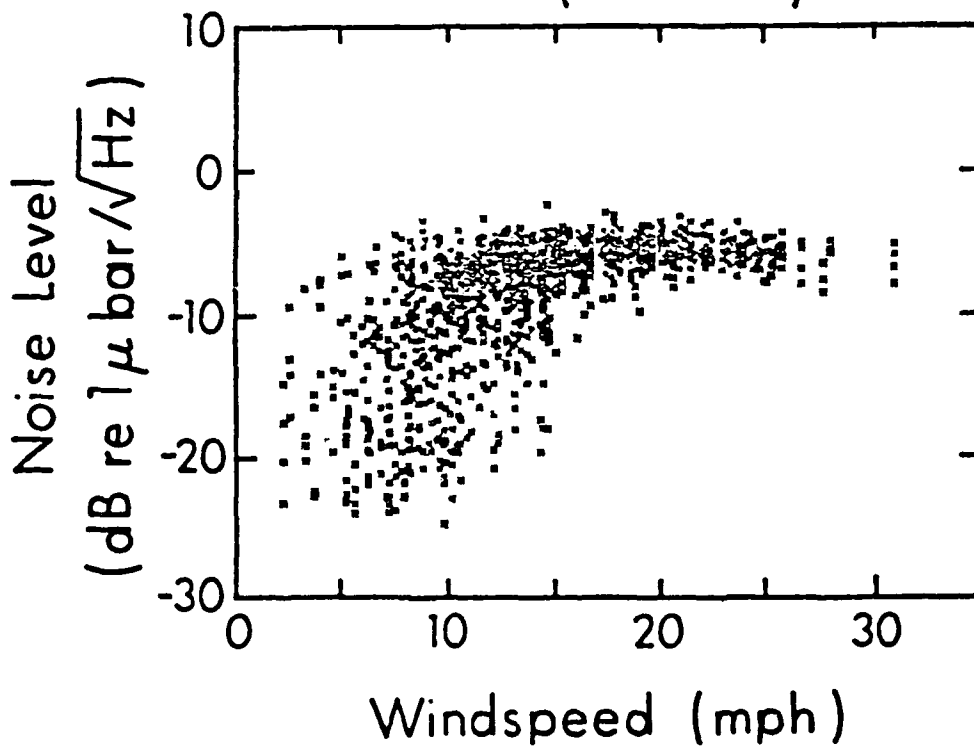






Hydrophone 74

16 (2.34 Hz)



CARTESIAN DIVER OBSERVATIONS OF DOUBLE FREQUENCY PRESSURE
FLUCTUATIONS IN THE UPPER LEVELS OF THE OCEAN

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Some recently analyzed data recorded by the Cartesian diver in May, 1968 demonstrate the existence of pressure fluctuations at double the frequency of the surface wind waves. The observations were made at depths ranging from 100 to 300 m, essentially in the near field of the sources according to theories of the generation of microseisms by Longuet-Higgins (1950) and Hasselmann (1963). In this depth range the variation of the pressure intensity with depth has a particularly simple form that allows one to measure the input of energy from wave-wave interactions directly above the diver.

The diver moves alternately up and down under control of adjustable buoyancy through a depth range from the mixed layer to 300m. It moves amazingly smoothly, essentially at its terminal velocity of 0.14 m/s relative to the surrounding water at all times except when changing direction. It has a device that records the rate of change of pressure in order to measure the vertical velocity of the diver in an absolute frame of reference and thereby infer the vertical component of the orbital velocity of internal waves. Owing to the smoothness of motion it has proven possible to extract from the pressure records evidence of minute pressure fluctuations of much higher frequency than those caused by the internal waves. The spectrum of pressure rate-of-change (figure 1) shows, at low frequencies, the fluctuations brought about by the motion of the diver in response to the vertical orbital motions of internal waves. In addition the spectrum has a sharp peak near $f_0 = .16$ Hz rising well above instrumental noise level. I believe this is the pressure signature of the wave-wave interference process. Dave Jacobs, who made the observations, has verified that it is not an artifact of some vibratory motion of the diver by examining data from a very sensitive angular accelerometer capable of indicating any irregular motions of the diver. None were detectable.

The spectral peak varies in intensity both with depth and with time. The depth variation follows the theoretical prediction for a near surface source having a nearly "white" intensity in wavenumber space. The simplest prescription is that the source spectrum is constant and isotropic up to a maximum wavenumber, k_m , and zero above. According to the theory of non-linear interference between opposed wave trains, the effective value of k_m is a function of the angular pattern of the waves and of the wavenumber of waves of frequency $f_0/2$. It cannot be greater than twice the latter, $2(\pi f_0)^{1/2}/g$, or in the present conditions, .066/m. The spectrum of pressure as a function of frequency alone is found by integrating over all contributing wavenumbers. Below the surface one must allow for exponential decay of the each wavenumber component. In the near-field region then, the pressure frequency spectrum may be approximated by

$$S_p = C \int_0^{k_m} \int_0^{k_m} \exp(-2kz) P \, dk_1 \, dk_2,$$

where C is the value of the pressure wavenumber-frequency spectrum just below the sea surface, and z is the depth where the pressure frequency spectrum is measured. With this form, S_p can be evaluated as

$$S_p = \tau C / 2d^2 [1 - (1 + u) \exp(-u)] \quad (1)$$

where $u = 2kd$. At depths such that $u \gg 1$ (but still in the near field of the sources, that is down to a kilometer or more) the pressure spectrum varies inversely as depth squared as a consequence of the gradual loss of shorter wavelength components of the pressure with increasing depth: $S_p = \tau C / 2d^2$. Below the near field region no further loss of pressure fluctuations occurs, but there are interference effects from bottom reflection of the acoustic pressure waves. The influence of the reflection coefficient on the pressure spectrum is shown in figure 2. Shown here are theoretical models that take into account the finite speed of sound in the water and its variation with depth, and assume the generating area is homogeneous to a great distance in all directions.

The important feature of these curves is that there is almost no effect of bottom reflectivity in the near-field region (figure 2, left). There is a large variation near the seafloor (right).

The observations fit reasonably well to the theoretical curve (figure 3). Here the square points represent mean square dp/dt found by integrating over the "microseism" peak and averaging over all data in 20m depth windows centered at 110m, 130m, . . . , 290m depths. In short, we have a measure of the local wave-wave interference intensity in a depth region (the near field) where the unknown parameter of bottom reflection is unimportant. The normalization factor C for the theoretical curve provides a measure of the local generation.

The time variation of the pressure spectrum is shown by successive eight hour averages (averaging over all depths between 100 and 300m, in figure 4. Numerals 1 through 6 indicate the eight hour periods beginning at 1600 UT on 5 May. During this time the windspeed generally decreased over a large area of ocean as shown by wind observations on shipboard and on the "Harvest" drilling platform about 94 miles away off Point Conception. The eight hour intervals mentioned above are indicated by numerals.

It is probable that the wind-waves and swell also decreased during this time, thus explaining the decreasing wave-wave interaction and therefore decreasing pressure intensity. A plot of the mean square pressure intensity (for each eight hour period, and found by integrating over the microseism peaks at all observational depths is shown in figure 5 for comparison with the windstress and windstress squared both estimated from the shipboard wind speed. This comparison is not expected to be close because the wave-wave interference can very well be associated with swell reflected from the coast or coastal islands, and the swell seems to have been generated by a storm center several hundred kilometers north west of the Cartesian Diver location. The local wind is part of the same storm system and is thus a proxy for the actual wave generation system. I plot windstress squared because the wave intensity may be proportional to windstress, while the second order interference is proportional to the wave spectrum squared.

Measurement of pressure fluctuations in mid-water is an excellent tool for unravelling the sources of microseismic motion on the seabed. This type of measurement may be the only way to get at the spectral intensity of opposed wave trains at the sea surface. In general it is very difficult to make adequately sensitive surface measurements to get the wave directional spectrum, especially as the opposed wave trains may be relatively weak compared to the predominant waves that move in the direction of the wind.

8hr avg, 100-300m

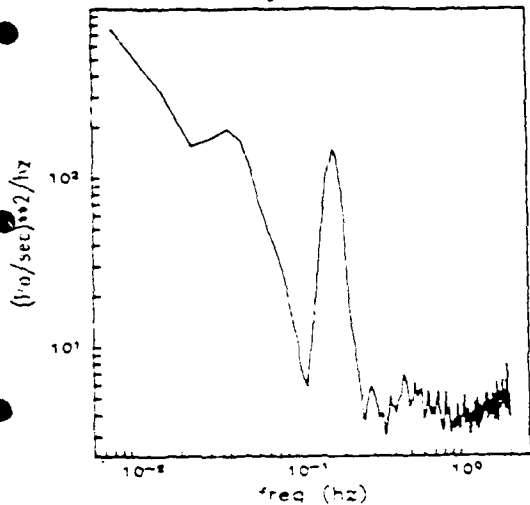


Fig 1. A spectrum of rate of change of pressure from the Cartesian Diver

Pressure spectrum at 0.17 Hz

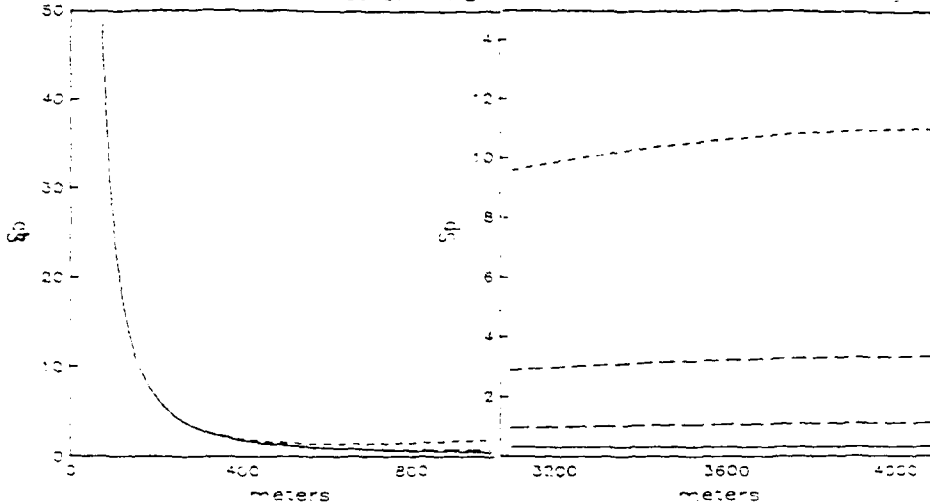


Fig 2. Variation of surface generated 0.17 Hz pressure fluctuations with depth. Water depth 4096 m. Bottom reflection coefficients as follows: 0 (solid line), .7 (long dash), .9 (medium dash), .97 (short dash).

Mean square rate of change of pressure

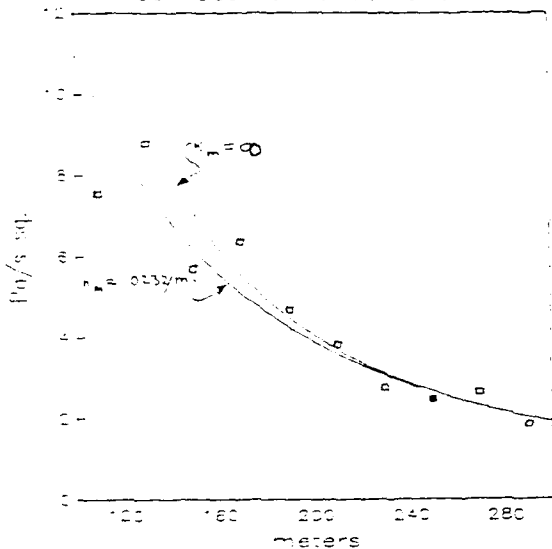


Fig 3. Mean square rate of change of pressure compared with equation 1. The maximum wavenumber k_m is noted.

Eight hour averaged spectra

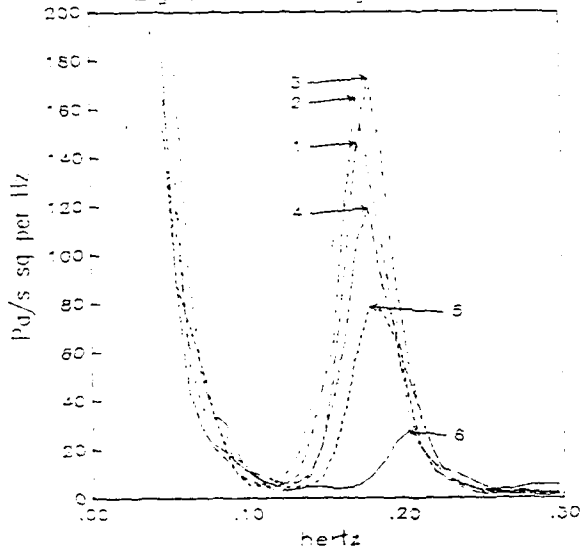


Fig 4. Variation of mean square rate of change of pressure with time during a 48 hour period

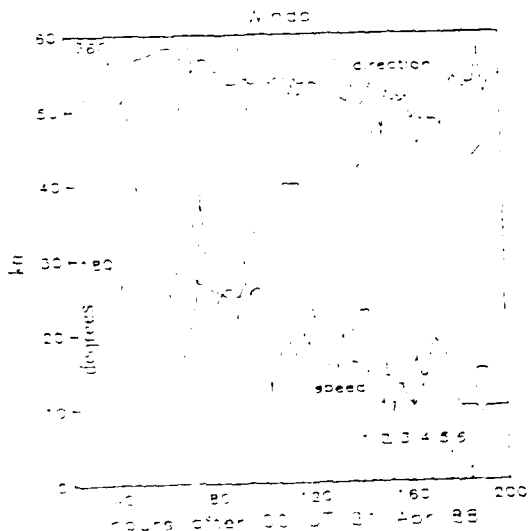
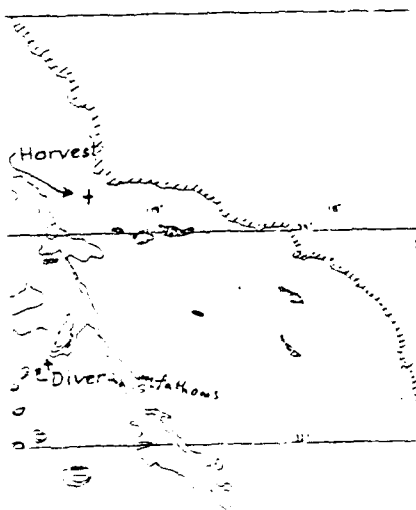


Fig 5. Windspeed at two locations. Harvest Platform, solid line; Cartesian Diver, dashed line.



Map shows locations Harvest Platform, solid line; Cartesian Diver, dashed line.

Eight hour averages

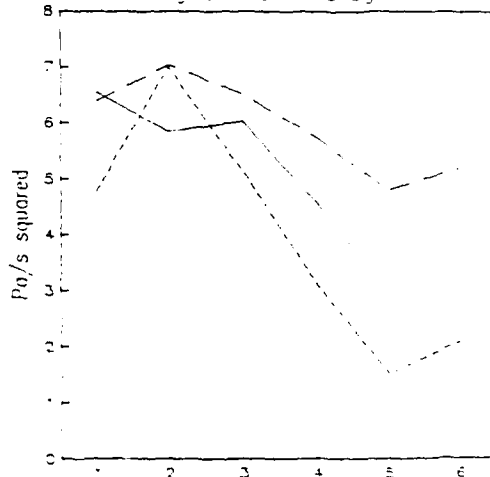


Fig 6. Mean square rate of change of pressure (continuous line) compared with windstress (long dashed line) and the square of the windstress (short dashed line). Normalization is arbitrary.

Wide-Angle Triangulation Array Study of Simultaneous Low-Frequency Microseism Sources

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INTRODUCTION

In this summary we highlight results of a study of microseisms at approximately 50 mHz, generated by separate major oceanic storms located in the North Atlantic and North Pacific oceans. These microseismic signals were recorded on the Alaskan Long Period Array (ALPA) and the Large Aperture Seismic Array (LASA) in Montana during November 26-28, 1973. In this study, we apply high-resolution frequency wavenumber (FK) analysis and beam forming techniques to determine direction of approach for microseisms arriving as surface waves from pelagic storms recorded at each array. We perform a wide aperture triangulation from simultaneous observations made at the ALPA and LASA arrays, adding to the work of previous array studies (Toksöz and Lacoss 1968; Capon, 1970) by improving the resolution of distance and azimuth between sources and receivers. The information on distance is most important since we may then address the question of near-coast or near-storm origin of the ambient noise field. We believe this is the first simultaneous determination of multiple microseismic source locations with this method.

Using data from large aperture seismic arrays offers potentially higher resolution in the study of microseism directionality and source location than obtainable from single station or network data. But for various reasons, array data are not easy to acquire and have not been fully utilized. Numerous investigations of microseisms have yielded valuable information directed towards an understanding of their generation and propagation. Yet, contradictions persist regarding the generation of microseisms observed on land. Whether they are near-coast ts (e.g. Haubrich and McCamy, 1969) or due to distant source (e.g. Iyer, 195) remains unresolved. The effects of path propagation of microseisms are still little understood and the generation of Love wave energy in microseism awaits further investigations (Rind and Donn, 1979).

RESULTS AND DISCUSSION

Primary microseism spectral power is observed to vary by more than an order of magnitude over a time scale of minutes. When signals from multiple microseismic sources are recorded by an array, their separate source amplitude fluctuations are observed as the modulation of their corresponding FK power peaks. Since the microseismic signal amplitudes from the two storms of this study were nearly equal, a moving time window analysis provided representative azimuth determinations. The relative stability of individually determined approach azimuths is displayed as a histogram in Figure 1. A similar distribution is also exhibited in phase velocities, ranging

from 3.1 to 3.4 km/sec, that being characteristic of Rayleigh-mode propagation over continental paths. Although the FK peak observations accurately reflect the significant microseismic sources detected during each window, we wish to emphasize that the number of observations made for a given azimuth are skewed in the sense that stable secondary peaks associated with the temporarily weaker of two microseismic sources are not well represented in our current analysis scheme.

Figure 2 shows the microseism source locations determined by triangulation of the approach azimuths. The distribution of approach azimuths determined for different time windows over the same 4000-point data sample provides an estimate of azimuth stability and error. For the Atlantic storm, the azimuths obtained from LASA differ by 4° over a 48-hour period, while the approach azimuths determined from ALPA data are nearly constant. The approach azimuth variation for the Pacific storm is similar: LASA azimuths differ by 16° while the azimuths from ALPA are within 3° over the same time period. An example distribution for both ALPA and LASA approach azimuths is shown in Figure 1 for the time period 1200 to 1300 UT, Nov. 26, 1973. Examining the clustering of azimuths determined for the entire time period studied suggests a stable location for the dominant FK power peaks.

When both storm microseism source amplitudes were low, other FK power maxima emerged, displaying more or less random azimuthal distribution. These may be ascribed to local or distant sources but could not be triangulated because of lower FK peak power at either array. The more persistent of the remaining azimuths along with their distribution are mapped onto the central triangulation position for the microseism sources. In doing so, it appears that the source locations for the Pacific storm are likely to be away from the coast, while the location for the Atlantic storm appears to be near coastal as shown in Figure 2. For the Pacific storm however, the actual storm position is well to the northwest. Also, the triangulation microseism source locations determined for time periods associated with earlier storm positions do not track the fast-moving storm in the Pacific, but rather remain more nearly stable near the position shown in Figure 2. The persistent location of the microseismic source suggests that it is not a direct function of the storm position. We speculate, instead, that it is associated with a stable near-coastal "bright" spot that acts to enhance the coherent storm-generated microseismic energy received at the two arrays.

Although pelagic storms provide the source of microseismic wave energy, it is the interplay between the storm parameters, the resulting storm waves, the direction of storm wave propagation, and the near-shore processes that acts to enhance or inhibit the production of coherent microseisms. From the perspective of a seismic array, over a specific time interval, only the most energetic and coherent portion of the noise field is detected in the FK domain. FK analysis is sensitive to, and therefore reflects the position of, only a fraction of the total storm-related microseismic noise field.

CONCLUSION

The results of this study demonstrate the capability of FK analysis to identify two or more sources of coherent microseism energy. The stability and robustness shown by the method permit simultaneous determination of phase velocities and approach

direction of the microseismic wave propagation. Triangulation by two arrays reveals the locations of two microseism sources which appear to be clearly associated with storm systems but not as a close function of their locations.

Some interesting questions have emerged from the results of this study. In particular: what makes a particular shoreline or near-shore location persistently bright in an FK power sense? We speculate that it may be the result of some combination of local coastal resonance modes involving coastal sea-bottom morphology, storm wave spectral power distribution and storm wave approach directions. Do similar storms excite the same local modes? Are there particular near-shore environments that preferentially tend to generate FK peaks regardless of the storm location within a given oceanic basin? Does the excited near-shore region vary with storm speed, peak frequency, or tracking direction? A related, but still unanswered, question regards the source location of *double-frequency* microseisms. Does its location exhibit similar stability? We expect to report on these and related questions in the course of our ongoing research.

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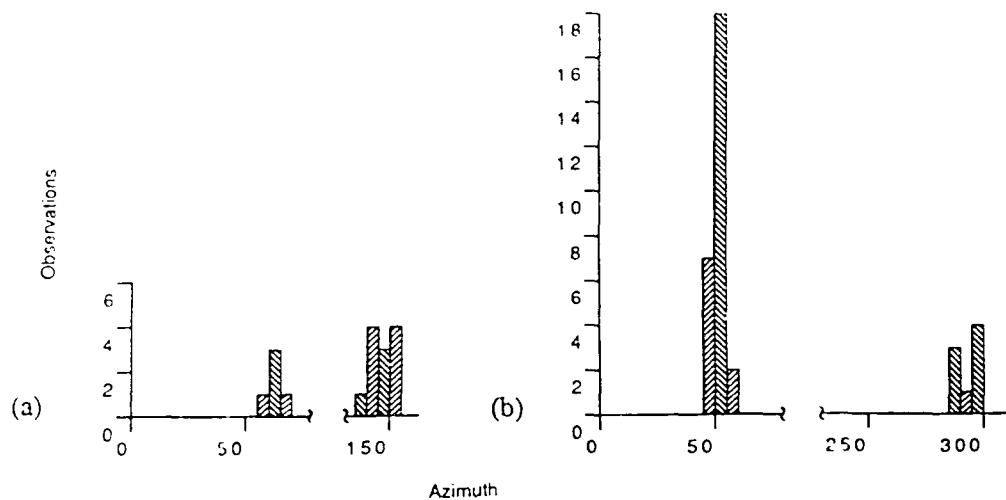


Figure 1. Histogram showing an example distribution of consistent approach azimuths determined from FK analyses of (a.) ALPA and (b.) LASA microseism data. Anomalous azimuth and phase velocity determinations, based on the whole time period analyzed, have been deleted for clarity. The two azimuth peaks shown in each graph are associated with the two separate microseism sources discussed in the text. The azimuth distributions are used to estimate the likely microseismic source locations shown in Figure 2.

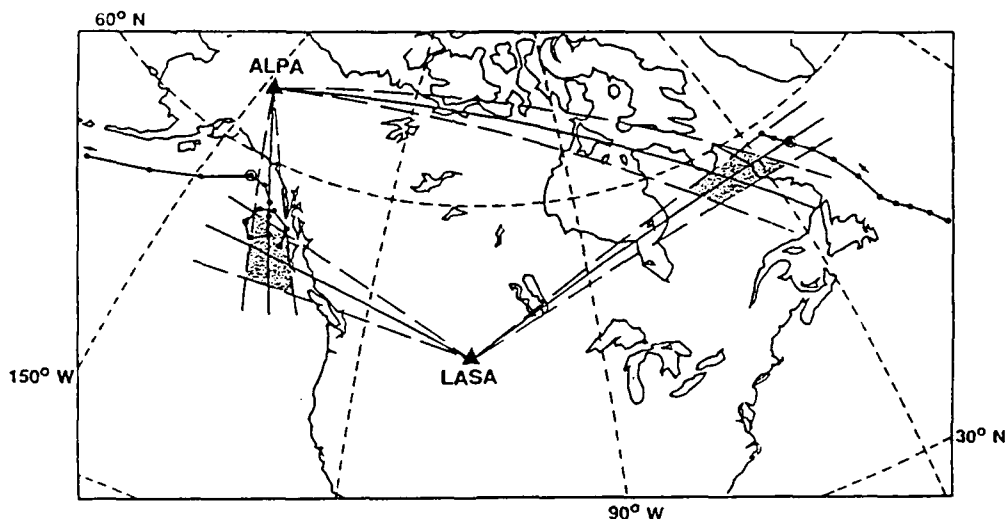


Figure 2. Map showing locations and tracks of the centers of the two major storms believed to be the source of the microseisms analyzed in this study (After Mariners Weather Log, 1973). The storm center locations are indicated by solid circles at 6 hour intervals. Two small reference arrows are placed at the reported storm positions for Nov. 25 1800UT and indicate the storms' directions. The two solid dots enclosed by open circles show the storms' positions on 26 Nov. 1973, at 1200UT. The azimuths determined from multiple windowed FK analyses for this time period and their estimated standard deviations are indicated by solid and dashed lines respectively. The shaded areas give an estimate of the areas most likely to contain the microseism sources. The perimeters are drawn from the FK-determined azimuth distribution over the length of the time period (4000 sec). Locations of ALPA and LASA arrays are denoted by solid triangles. This map is an azimuthal equidistant projection centered on the midpoint of a great circle connecting the two arrays.

Particle Motion and Pressure Relationships of Ocean
Bottom Noise at 3900 M Depth: 0.003 to 5 Hz

by

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Introduction

Possibly the most extensive set of ocean bottom data on long period seismic background noise and signals was obtained from the Columbia-Point Arena Ocean Bottom Seismic Station (OBSS). OBSS operated for over six years and a number of papers (e.g., Auld *et al.*, 1969; Latham and Nowroozi, 1968; Piermattei and Nowroozi, 1969; Sutton *et al.*, 1965; Nowroozi *et al.*, 1968 and 1969) have been published using OBSS data on, e.g., gravity and pressure tides and tidal currents; ocean bottom microseisms; Rayleigh waves from pure oceanic paths and across the continental margin; and local and regional earthquakes.

The Columbia-Point Arena OBSS was installed on 18 May, 1966 at 38° 09.2'N, 124° 54.4'W about 200 km west of San Francisco at a depth of 3903 meters. It was in continuous operation for more than six years, until 11 September 1972. The OBSS included a Lamont long-period (LP) triaxial seismometer (15 sec natural period, originally developed for lunar use); three-component short-period (SP) system (1 sec natural period); long-period (crystal) hydrophone; short-period (coil-magnet) hydrophone; ultralong-period (Vibratron) pressure transducer; thermometer; current amplitude sensor; and a current direction sensor.

Data

The data, recorded on two 7-channel instrumentation-quality FM tape recorders and on seismograph-type drum recorders and strip chart recorders, are currently stored at Lamont-Doherty Geological Observatory (L-DGO). Short- and long-period response curves are published in Sutton *et al.*, 1965. The seismometers, both long- and short-period, as well as the amplifiers and recorders were well calibrated (this is admirably documented in S. N. Thanos, 'OBS Calibration Manual,' L-DGO Technical Report, 1966) and calibration pulses were recorded daily on the tapes.

A Sangamo (TM) analog system and calibrated reproduction discriminators were used to play back the old FM tapes and the data were digitized by an A-to-D system built by D. Lentricia at L-DGO. The long-period data are digitized at 8 sps and the short-period data at 80 sps. A four-pole Bessel anti-alias filter was applied at a frequency equal to one-fifth the sample rate. For each sample of signal or noise digitized, we also digitized the calibration pulses for that date. When, later, we removed the system responses from simultaneous noise samples of long- and short-period data, for each of the three components of ground motion the LP and SP noise spectra were perfectly

matched between about 0.1 and 0.5 Hz (system noise limits the spectra above 0.5 Hz for LP data and below 0.1 Hz for SP data).

To further check the accuracy of the newly-digitized, twenty-year-old data, we digitized a few earthquakes to verify polarities for all eight components, to compare m_b magnitude determinations with those published, and finally to compare pressure-vertical (P/Z) ratios for both compressional and Rayleigh waves with theoretical values. The only potential problem we encountered was with the pressure data. Unlike the seismometers, the hydrophone channels have no daily calibration pulses. Comparison of observed and theoretical P/Z amplitude and phase relationships for digitized earthquake P-waves and Rayleigh waves has convinced us that the 6 dB/octave high-pass corner at 3 Hz, included in the original calibration curve for the coil (SP) hydrophone, does not exist and should be replaced with a 0.3 Hz, 6dB/octave high-pass corner. The original 3 Hz corner represents the effect of an hydraulic relief system required for installation in deep water; perhaps it doesn't work the way it theoretically should. Comparison of corrected background pressure spectra from the LP and SP (with the 3 Hz corner shifted to 0.3 Hz) hydrophones shows that the shapes of the spectra, both amplitude and phase, match closely between about 0.1 and 0.5 Hz (the overlapping frequency band), but that the SP hydrophone data is about 6 dB higher than the LP hydrophone data. So far, we haven't found the cause of the offset, but we are trusting LP hydrophone data. Be aware, however, that there may be a +6dB uncertainty in the LP pressure power spectra presented in this paper. This may explain why we obtain slightly lower values of LP pressure-vertical displacement ratios for earthquake Rayleigh waves and microseisms than those published by early investigators of the OBSS data (Piermattei and Nowroozi, 1969).

A final comment about the data: though one channel on both the LP and SP FM tapes measured compensation for wow and flutter during initial recording, we did not have the equipment to electronically compensate during playback. Instead, we subtracted the digitized compensation channel from each of the data channels and thereby improved data quality. Overall system noise limits background data above about 5 Hz and below about 0.003 Hz. (Except that the long-period horizontals record tilts down to at least 0.001 Hz.)

Results

Pressure and vertical velocity spectra, corrected for instrument response, are shown in Figure 1A. The pressure spectra agree with those obtained by Cox *et al.* (1984), showing a strong minimum between about 0.03 and 0.1 Hz. At times, the "single frequency" microseismic noise, at about .06 Hz in this sample, is nearly absent, giving a stronger minimum. The pressure and vertical velocity spectra are quite similar except that the P/Z ratio increases significantly below about .02 Hz. These spectra are calculated for a one hour period during which the bottom current velocity did not exceed 2 cm/sec. H. Bradner and M. Reichle (personal communication) have obtained comparable results for the P and Z components of the OBSS between about 0.05 and 0.2 Hz for a different time period.

In Figure 1B the power spectrum from Figure 1A including SP vertical data is compared with seismic noise on continents. Curves for continents represent noisy and quiet conditions on hard rock. Between .06 and 5 Hz, the ocean bottom noise level pictured here is generally less, but close to the level of noisy continental data: below .04 Hz it is well above continental noise. The greatest "real" differential occurs near .01 Hz where the OBSS noise is over 32 dB above the "noisy" continental curve and it is possible that the "real" difference continues to decrease below .004 Hz. How much improvement would be obtained by better bottom-package design; shallow or deep burial within the bottom-sediment; or rigid coupling within the basalt of layer two are important questions to consider for the design of future OBS systems (e.g., Sutton and Duennebie, 1987).

Power spectra and coherency among three components - pressure (P), vertical (Z), and horizontal motion perpendicular to shore ($H \perp$) - are calculated for two noise samples, 21 June and 4 July 1966 (Figures 2 and 3). The two samples illustrate time-variable features of microseismic noise. Strong coherency in Figure 2 near .06 Hz ("single frequency" microseisms) coincides with a peak in the power spectra of all three components. Between vertical displacement and pressure, coherency is strong from about .06 to .14 Hz and also at .30 Hz. Corrected phase relations and amplitude ratios for the spectral and coherency peaks near .06, .14, and .30 Hz are appropriate for fundamental mode Rayleigh waves. Theoretical results for appropriate velocity structures (Latham and Nowrozi, 1968) indicate that at .14 Hz fundamental mode Rayleigh waves are near the cross-over from retrograde (at the longer periods) to prograde particle motion. Thus, the observed phase relationship between Z and $H \perp$ indicates propagation toward shore at the OBSS location (about 160 km offshore) for the "single frequency" microseisms. Although Z - $H \perp$ amplitude coherency is not strong for the .14 and .30 Hz peaks, a stable phase coherency at .14 Hz indicates predominant $\pi/2$ phase difference, opposite the sign of the .06 peak. If the .14 Hz microseisms are prograde, their propagation is shoreward. The phase difference between Z and $H \perp$ around .30 Hz, though messy, looks more like $\pi/2$ than $-\pi/2$, again suggesting shoreward propagation of prograde fundamental Rayleigh waves. In contrast, the phase difference between Z and $H \perp$ in Figure 3 appears to be $-\pi/2$ which would indicate seaward propagation. The relatively poor coherency between Z and $H \perp$ for both the .14 and .30 Hz microseisms suggests variable propagation directions. Note also in Figure 3 that the "single frequency" microseism is not well developed.

In Figure 4 a P-Z coherency maximum near .01 Hz coincides with a spectral maximum most clearly observed from the hydrophone. The ratio of pressure to vertical velocity is much too high for fundamental mode Rayleigh waves. The amplitudes of these disturbances do not correlate in time with tidal currents but do correlate with ocean wave heights observed along the nearby California coast. An example is shown in Figure 5. The pressure signal is produced either by long (shallow water) waves non-linearly generated near shore from the ocean swell, or by differences in pressure beneath the high-amplitude vs. low-amplitude swell as varying amplitude wave-sets pass over OBSS, or a combination of the two. In the former case shoaling water is required, whereas in the latter case the disturbance would be observed wherever water depth is not large compared to the "wavelength" of the wave sets. In either case the velocities and wave

lengths involved are much smaller than for seismic signals of the same period and the bottom moves as a forced deformation.

Summary

Results discussed in this paper are:

1. Coherent spectral peaks are observed on vertical seismometers and hydrophones near .01, .06, .14, and .3 Hz;
2. The peak at .06 Hz also shows strong coherency with horizontal motion perpendicular to the coast;
3. The peaks at .06, .14, and .30 Hz have comparable pressure-vertical velocity ratios and $\pi/2$ phase difference appropriate for free-running boundary waves. Amplitude and phase relations, including horizontal components, suggest the .06 and .14 Hz peaks are predominantly fundamental mode Rayleigh waves propagating toward shore; the .30 Hz peak is not as clear;
4. A peak at .01 Hz has 5-10 times greater pressure-vertical velocity ratio than shorter period peaks;
5. The amplitude of the .01 Hz pressure peak correlates with ocean wave heights observed at nearby coastal stations.

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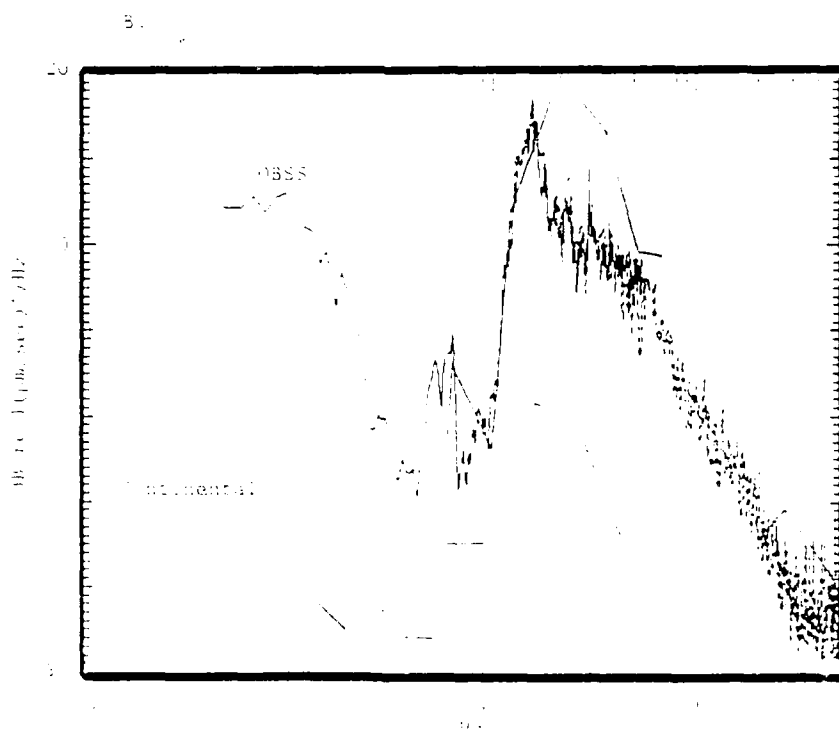
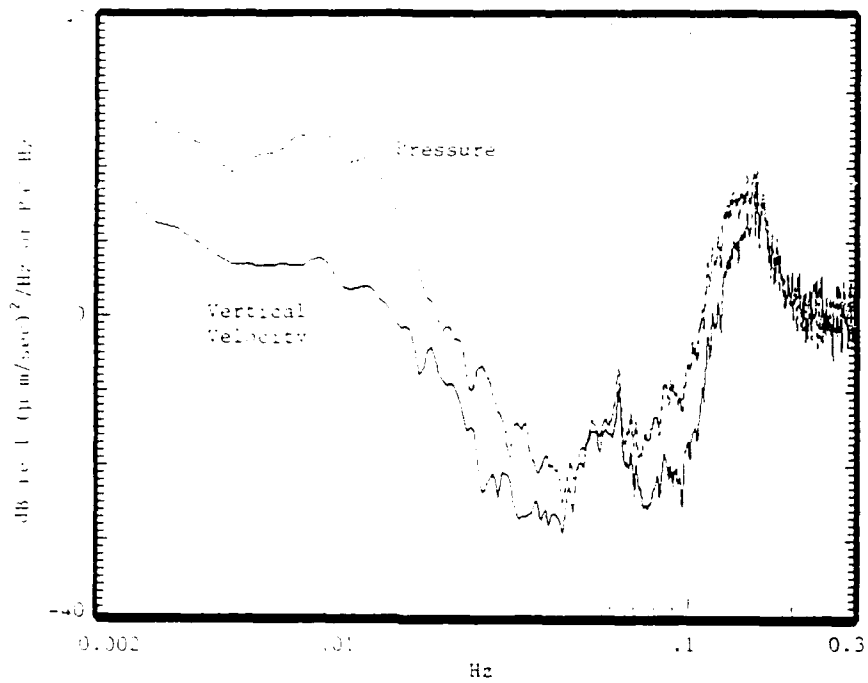


Figure 1. (A) Instrument-corrected noise power spectra of pressure and vertical velocity. There is an uncertainty up to +6dB for the pressure data. (B) Vertical velocity power spectra for OBSS compared with spectra for continental stations under quiet and noisy conditions (Jon Peterson, USGS, unpublished data). The OBSS data are simultaneous samples from LP (solid line) and SP (dashed line) seismometers.

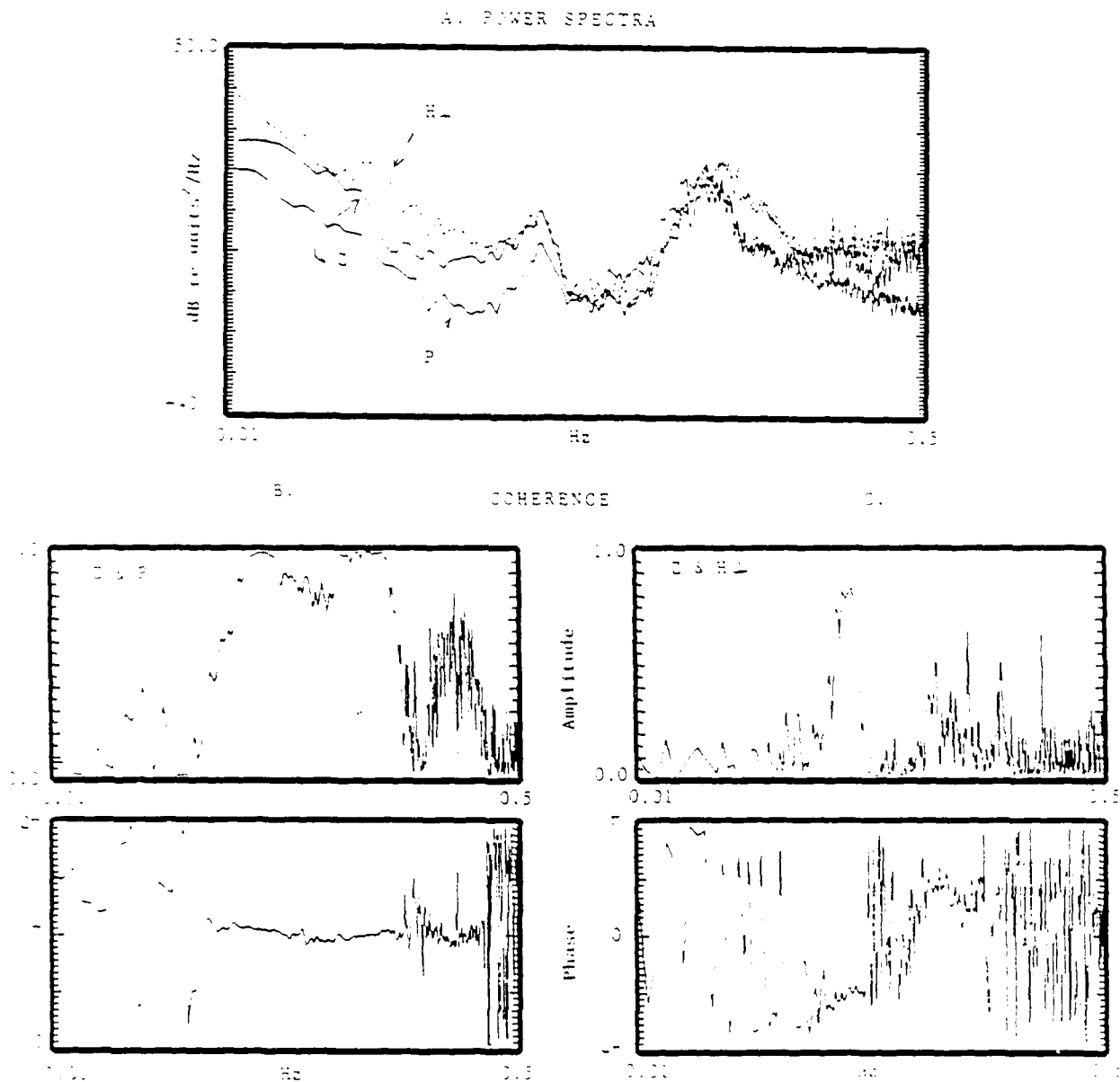


Figure 2. (A) Instrument-corrected noise power spectra of pressure (P), vertical (Z) and horizontal (H_{\perp}) motion. Units are Pa for P and μm for Z and H_{\perp} . Power spectra are calculated for 3655 secs of noise processed with 8 Hanning windows, 50% overlap. (B) Coherency spectra, Z and P components, amplitude and phase of the 3655 secs in A.: 16 windows, 62% overlap, 552 sec per window. Phase convention: P leads Z. (C) Same as 2B. but for Z and H_{\perp} components: $+H_{\perp}$ is azimuth 246° . Phase convention: H_{\perp} leads Z.

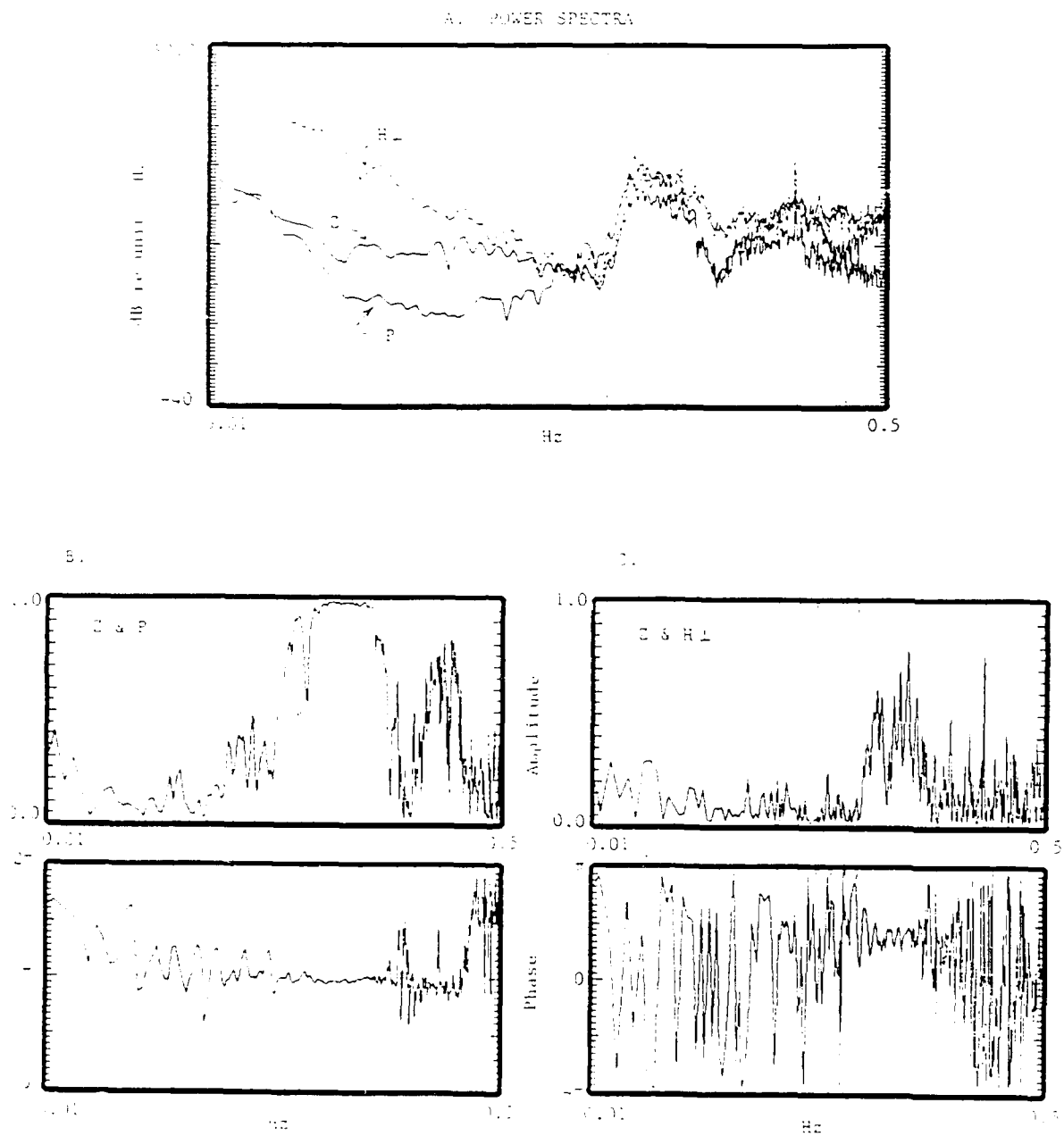


Figure 3. (A) Instrument-corrected noise power spectra of pressure (P), vertical (Z) and horizontal (H_{\perp}) motion. Units are Pa for P and μm for Z and H_{\perp} . Power spectra are calculated for 3660 secs of noise processed with 8 Hanning windows, 50% overlap. (B) Coherency spectra, Z and P components, amplitude and phase of data in A: 16 windows, 62% overlap, 552 sec per window. (C) Same as B. but for Z and H_{\perp} components. Phase conventions same as Figure 2.

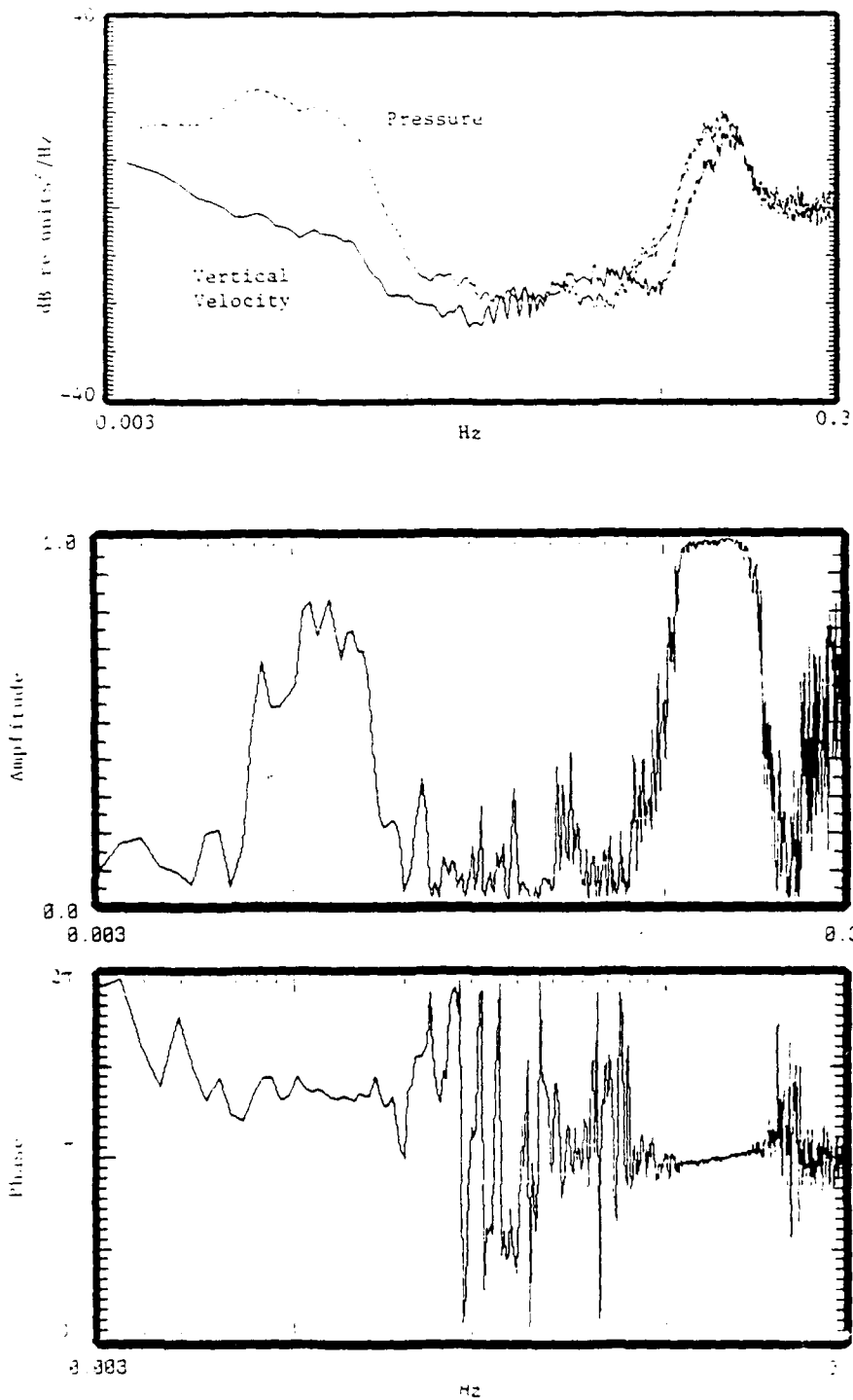


Figure 4. Instrument-corrected noise power spectra and squared coherency spectra for Z and P components. Units are Pa for P and $\mu\text{m}/\text{sec}$ for Z. Power spectra are calculated for three hours of noise processed with 4 Hanning windows, 50% overlap. Coherency spectra of the same data (vertical displacement rather than velocity) are calculated with 16 windows, 62% overlap, 1630 sec per window. Phase convention: P leads Z.

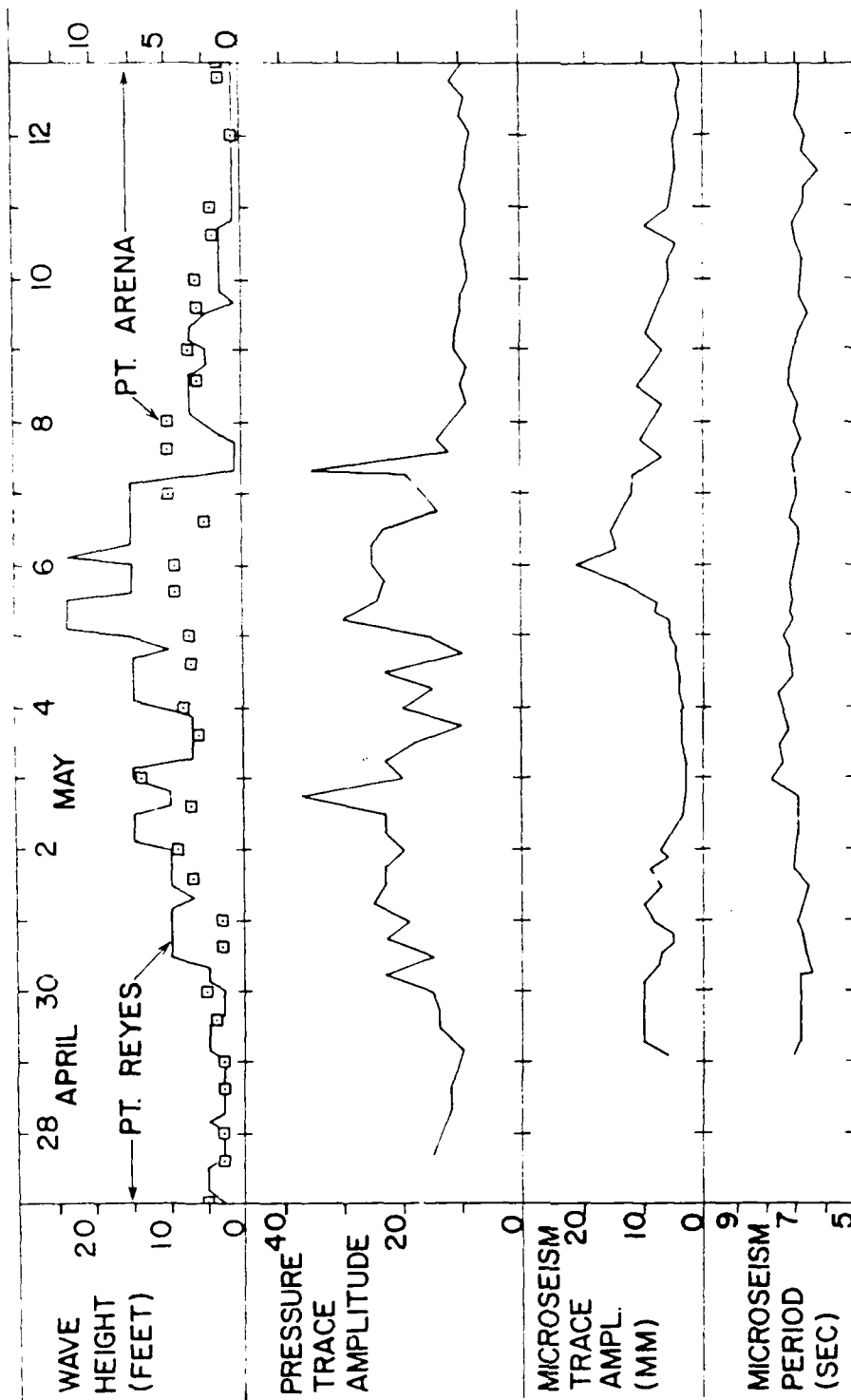


Figure 5. Comparison between visual wave height observations at Pt. Reyes and Pt. Arena and observations from OBSS. Amplitudes of pressure disturbances are generally for the period range 60-180 sec; the sharp maxima on the pressure record are related to bursts of longer period (240-300 sec) energy (figure from Sutton *et al*, 1965).

ULF Seismology and Ocean Surface Waves

S.C. Webb, Scripps Inst. Oceanography

Ocean surface waves are most energetic at periods shorter than 15-20 seconds, but the spectrum extends down to the frequency of the tides both on the shelf and in deep water. The waves are true surface waves and pressure sensors at the seafloor will detect only those waves which are of sufficiently low frequency so that kh (wavenumber times water depth) is of order one. On the shelf the high frequency cutoff will occur at frequencies as high as .1 Hz (200m), in deep water (4-5 km) only waves at frequencies below .03 Hz generate detectable signals at the seafloor. The propagating pressure disturbances associated with these waves drive vertical displacements of the seafloor. The natural background of seismic noise is very low at frequencies below the microseism peak (<.1Hz) and at frequencies below the surface wave cutoff frequency the displacements and pressure signals associated with the surface waves usually predominate.

Most estimates of the pressure spectrum from stations in the NE Pacific show the spectral level of pressure fluctuations near .01 Hz usually lies between 10^3 and 10^4 Pa²/Hz. Low frequency ocean waves are generated from short wavelength, wind driven waves by nonlinear processes in shallow water. Most of this low frequency energy is trapped as edge or shelf waves on the continental shelves. A little energy leaks off the shelves to become freely propagating waves in deep water. Except on the shelves, frictional attenuation is negligible and the waves lose energy only by repeated reflection off of opposite coastlines. The deep ocean low frequency energy depends on some average of the short wave energy hitting the coasts around the entire ocean basin and thus in a complicated fashion on the average wind stress across the basin.

A four month long record of low frequency pressure fluctuations from the Fall of 1985 from a site in the NW Atlantic show much more variability in spectral levels in the surface gravity wave band than I normally associate with measurements from the Pacific. These waves are also consistently less energetic (10-100 Pa²/Hz). The record shows large variations clearly associated with large storms. A nice example is that of a hurricane off the Florida coast

which day by day generates larger and larger low frequency waves (and therefore must be generating bigger wind waves and swell) until the storm disappears into the Gulf of Mexico, after which the long waves quickly abate with a e-folding time of less than six hours. Tsunamis in the Pacific lose energy with an e-folding time of about twelve hours, roughly corresponding to the time required for a long wave to propagate across the Pacific. Thus the much shorter decay time observed in the Atlantic seems very reasonable.

At other times the low frequency wave energy correlates poorly with particular storms but well with the area averaged wind stress over the North Atlantic. There is a phase lag of about two days, presumably associated with growth of wind waves and the propagation of these short waves to the coasts.

T. Yamamoto has demonstrated that measurements of the seafloor displacements generated by surface waves can be used to determine the shear modulus of the shallow sediments on the continental shelf. The same technique is also viable in deep water. The wavelengths associated with surface waves in deep water are much larger so the displacements depend on the sediment and crustal structure at much greater depths below the seafloor than in shallow water.

Measurements from sites on Axial Volcano on the Juan deFuca Ridge provide useful constraints on the shear modulus to depths of a few kilometers. These measurements were obtained by J. Hildebrand and myself using a tethered LaCoste-Romberg gravimeter adapted for deep sea use combined with a pressure gauge. T. Yamamoto has developed a linearized inversion for the low shear velocity limit applicable for the shallow water problem. I believe the shear modulus can be shown to be the most important parameter in the inversion for the deep water case also. For a preliminary attempt at an inversion of the pressure and acceleration transfer function the density and compressional velocity have been fixed to values determined from seismic refraction work and only the shear velocity is varied. The inversion is based on a propagator matrix forward model routine and a linearized inversion scheme called Occam's razor developed by S. Constable. The transfer function measurements appear to require a very low shear velocity in the upper 100m of the crust. Below this layer a constant Poisson's ratio of about .27 well describes the model. The inversion minimizes the rms slope of the model, and puts a constant shear velocity layer below 2 km depth. Other modeling efforts suggest the measured transfer function

appears to preclude a significant low shear velocity layer (magma chamber) to depths of a few kilometers.

The technique appears promising, but awaits longer deployments (probably with an autonomous vehicle) to provide better statistical certainty and lower instrumental noise.

Buried OBS Array Measurements of ULF (.005-1 Hz) Ocean Noise Field

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Introduction

Over the last five years, quantitative measurements of the seismo-acoustic noise fields in the shallow oceans have been made using buried OBS arrays by the authors. This was mainly motivated by the invention of an entirely new geophysical remote sensing technique using gravity wave induced bottom motion (Yamamoto and Torii, 1986). The Geo-Acoustics Laboratory OBS sensor package contains a PME differential pressure transducer, three orthogonal medium period seismometers (Teledyne S-750), and two tiltmeters. Data from up to nine OBSs may be acquired either through analog cables, a digital radio telemetry system, or a self contained digital recording system. The hardwired system is limited to 200 m water depths; the radio telemetry system to 2000 m; and the self contained OBS to 7000 m.

To measure the seabed motion accurately, OBSs were buried about 1 m below the sediment water interface (Trevorrow et.al.,1988). This is necessary to avoid hydrodynamic forces on the OBS due to moving water and to insure good OBS/sediment coupling. (Note that the bed rigidity increases with the square root of the depth of burial). The BSMP method is capable of remote sensing the shear modulus of the seabed down to 1 km with a resolution of several meters (Yamamoto et.al.,1989).

ULF Noise Field in Shallow Water

We have made measurements of the ULF seismo-acoustic noise field in shallow water at Bahama Bank, northeastern continental shelf of the United States, and the continental shelf of Japan using buried OBS arrays.

Power spectra of pressure and three components of seabed acceleration measured at COST-G1 Borehole site on George's Bank (50 m water depth) are given in Figure 1. The pressure data show the wind wave energy (6-20 s) and the long gravity wave energy (20-200 s). Seismometer data shows the micro-seism energy (1-4 s) and the wind wave energy (6-20 s). The long gravity wave energy was not detected by the seismometers due to their limited sensitivity at frequencies lower than 0.05 Hz.

The seabed motion amplitude is inversely proportional to the shear modulus of the seabed (Yamamoto et. al. JFM 1978). George's Bank is made of a thick layer of dense sand. Therefore, the ULF seismic noise at George's Bank is about 50 dB quieter compared to muddy sites (like AMCOR 6011 site off New Jersey or Mississippi River Delta site).

Spacial coherence of pressure and vertical and horizontal ground acceleration measured by two buried OBSs separated by 270 m are shown in Figure 2. High coherence exist in the wind wave band and the long gravity wave band of pressure signal. Also showing good coherence is the microseismic band and the wind wave band from the seismic data. Burial of the OBSs revealed for the first time, that horizontal ground motion is quite coherent.

Figure 3 shows the admittance spectra with their associated coherences for a) Pressure and Vertical motion, b) Pressure and Horizontal motion at Atlantic Generating Station site (12 m water depth). Admittance is defined as the ratio of amplitude of seabed displacement to the water wave height.

ULF Noise Field in Deep Water

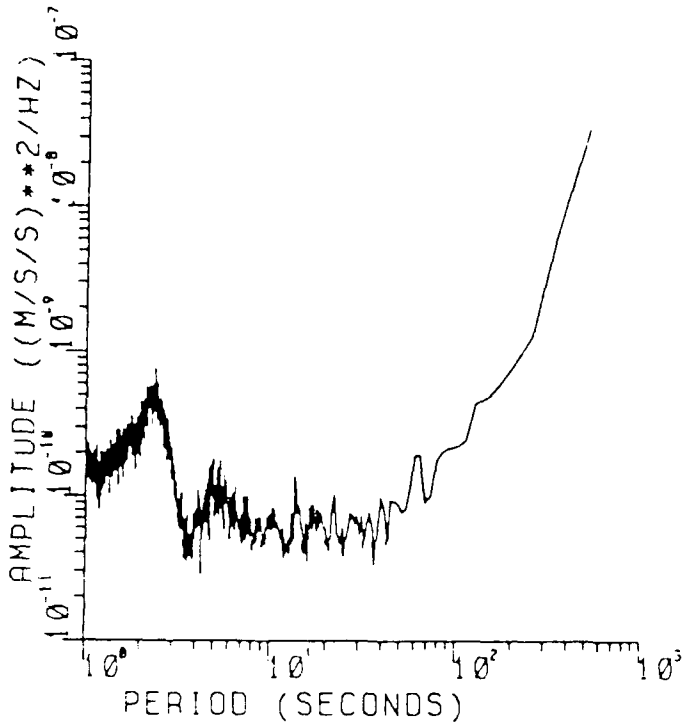
Our quantitative measurements of ULF noise field in shallow water revealed that the ULF seismic noise is keenly influenced by the bottom shear modulus structure. It is expected that this is also true in deep water. The only difference would be the effect of water depth on the noise generation. Gravity wave pressure decreases exponentially with the depth, so that only very long waves can excite the deep ocean bottom. In Figure 4, we show the predicted ULF seismic noise power spectra of the sandy site and the muddy site at 4 km water depths. In the calculation, bottom pressure spectra from Webb and Cox (1986) is used. Note that the ULF seismic noise is about 30 dB quieter at the sandy site compared to the muddy site.

This result strongly suggests that the ULF signal detection should be made by an OBS array buried in a sandy bottom in deep ocean, such as the continental slope of George's Bank.

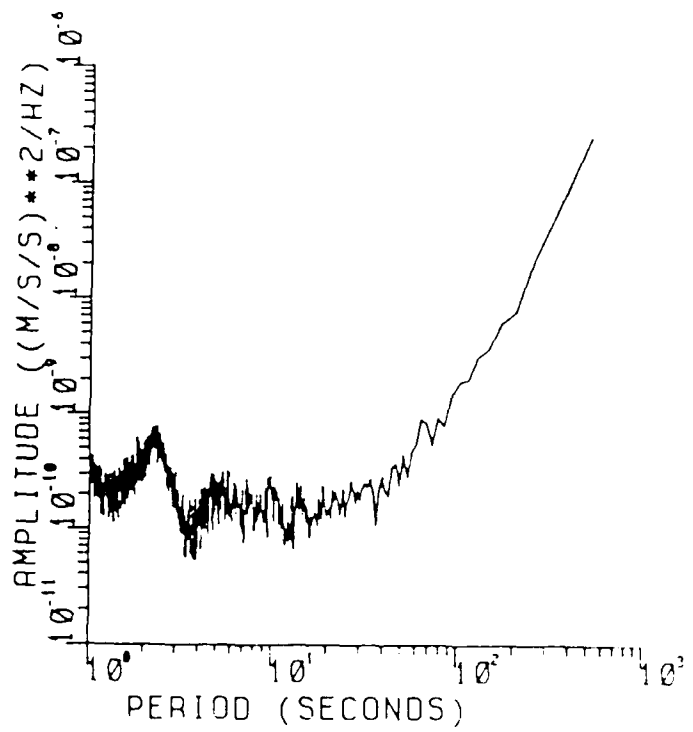
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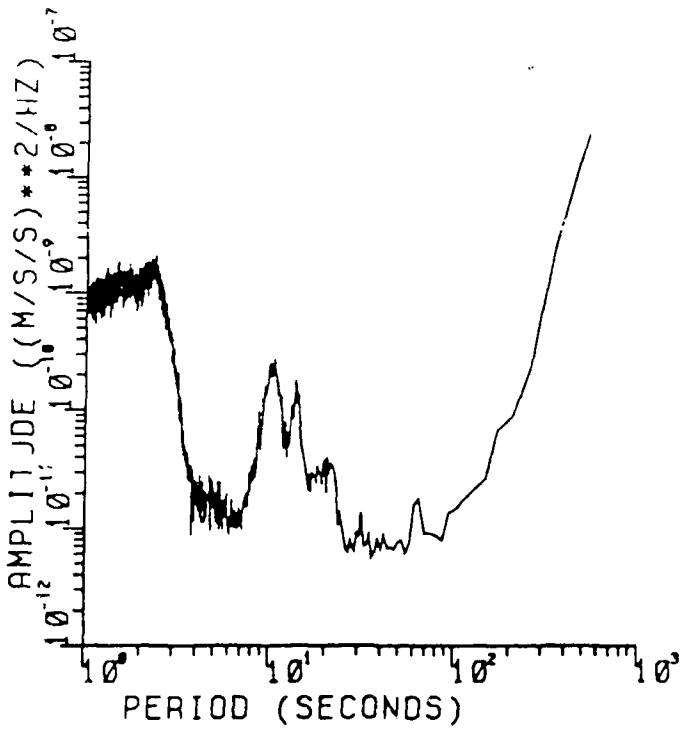
RADIAL SEISMOMETER



TRANSVERSE SEISMOMETER



VERTICAL SEISMOMETER



PRESSURE TRANSDUCER

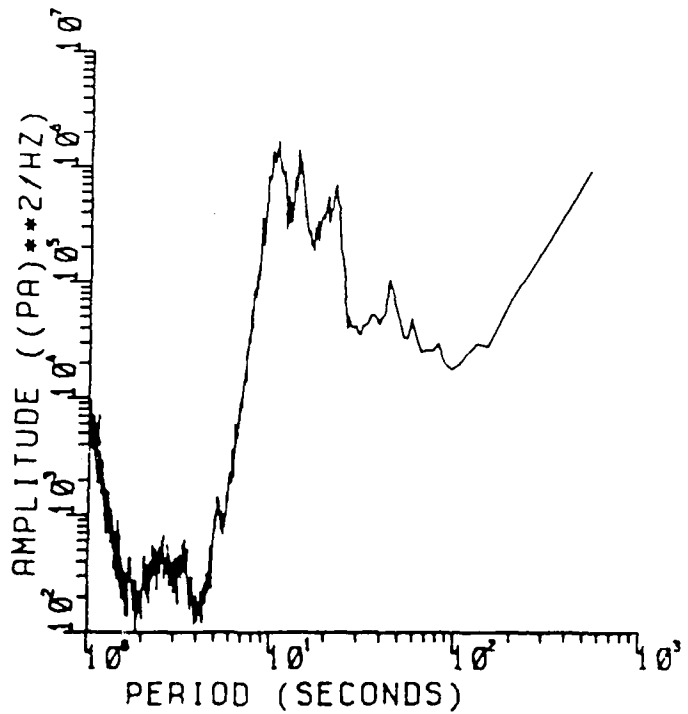
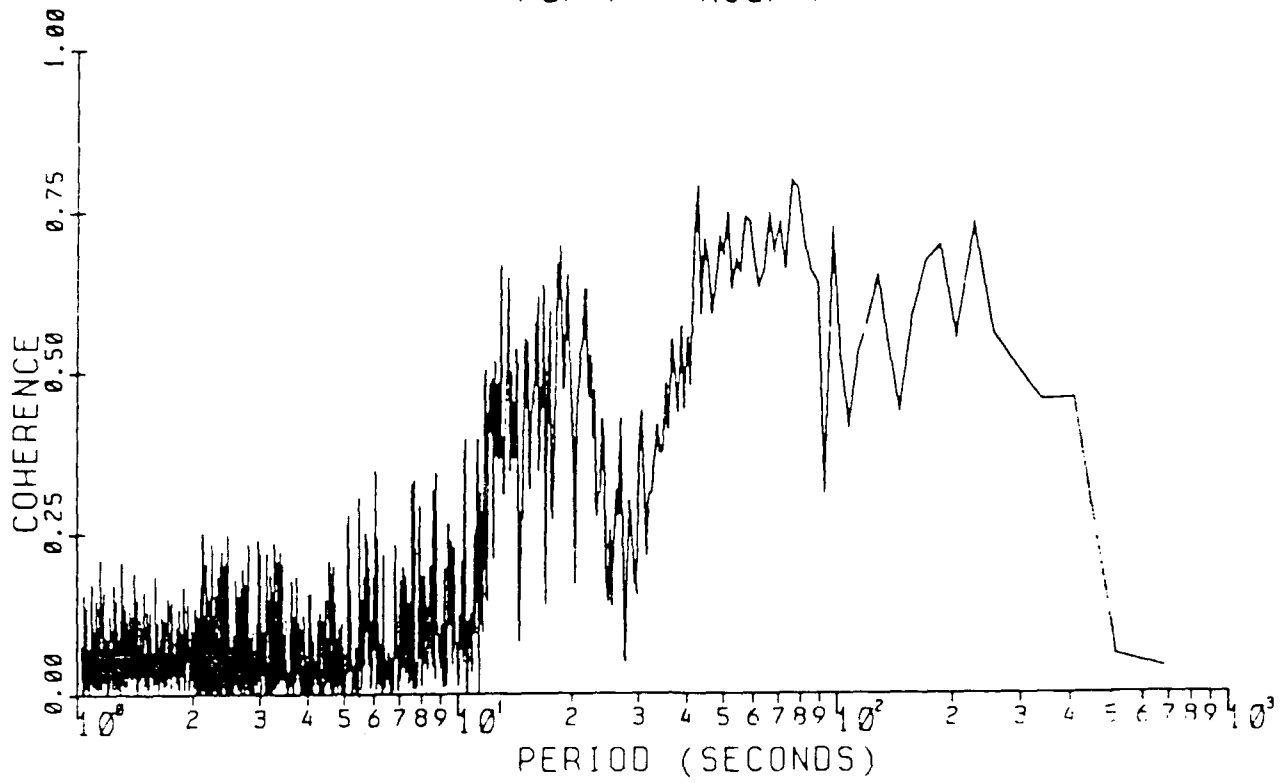


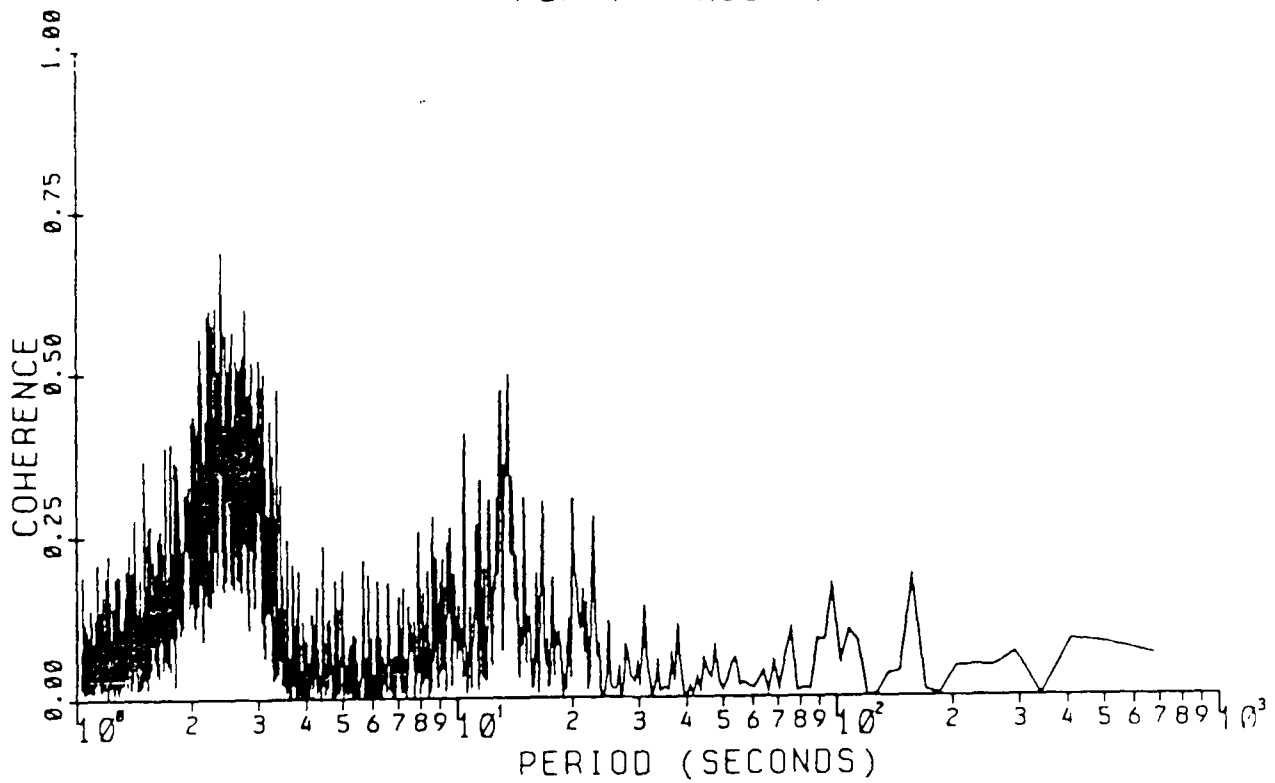
Figure 1. Power spectra of pressure and three components of seabed acceleration measured at COST-G1 Borehole site on George's Bank.

For P - Near P



(a)

For V - Near V



(b)

Figure 2. Spatial coherence of a) pressure, b) vertical seabed acceleration measured by two buried OISs separated by 270 m at George's Bank.

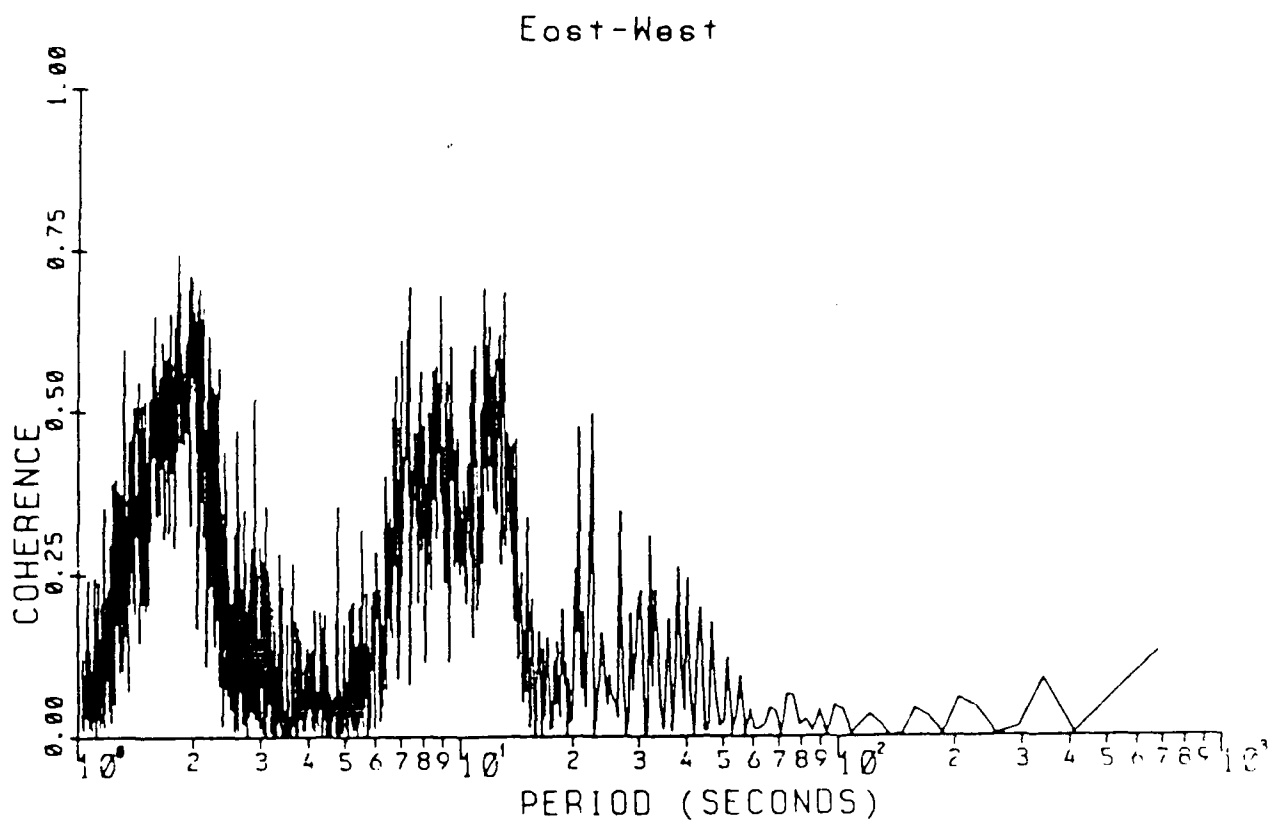
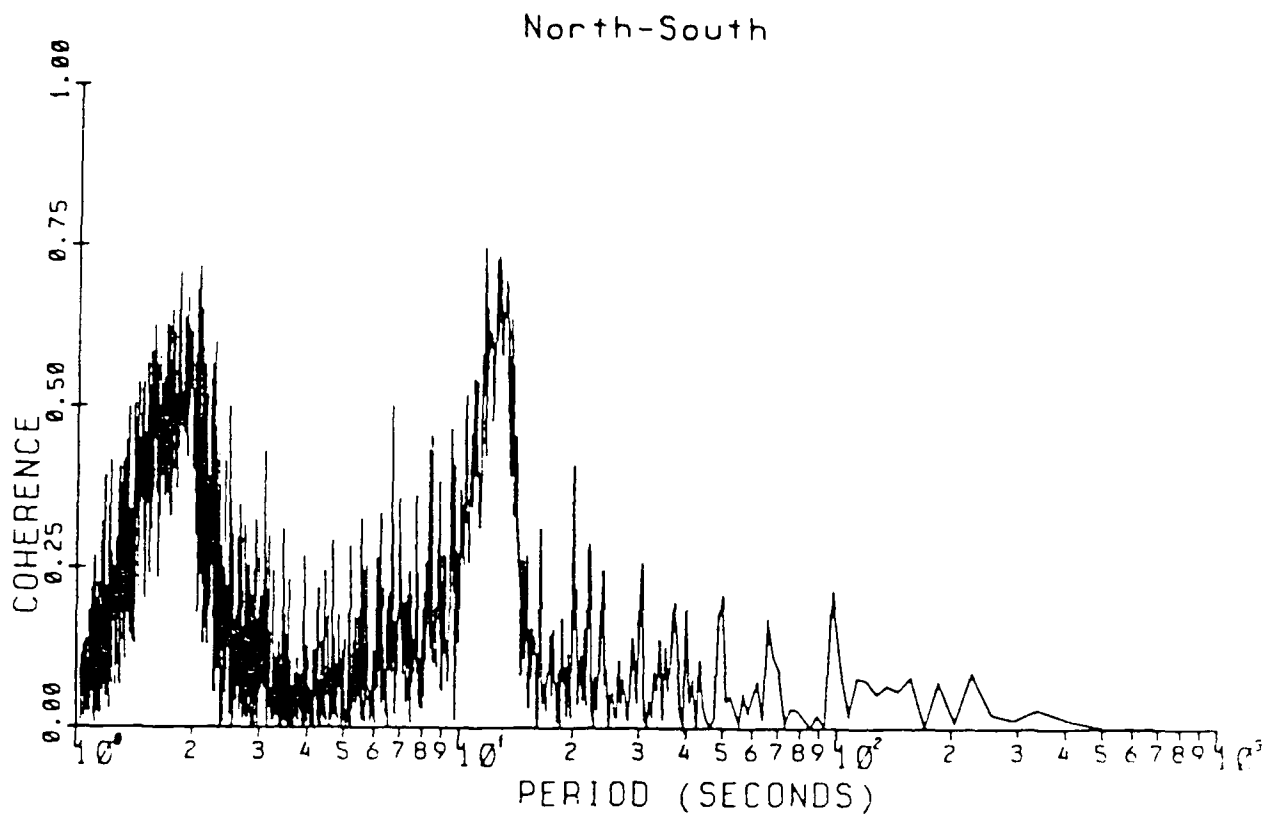


Figure 2.c,d Spatial coherence of horizontal seabed acceleration measured by two buried OBSs separated by 270 m at George's Bank.

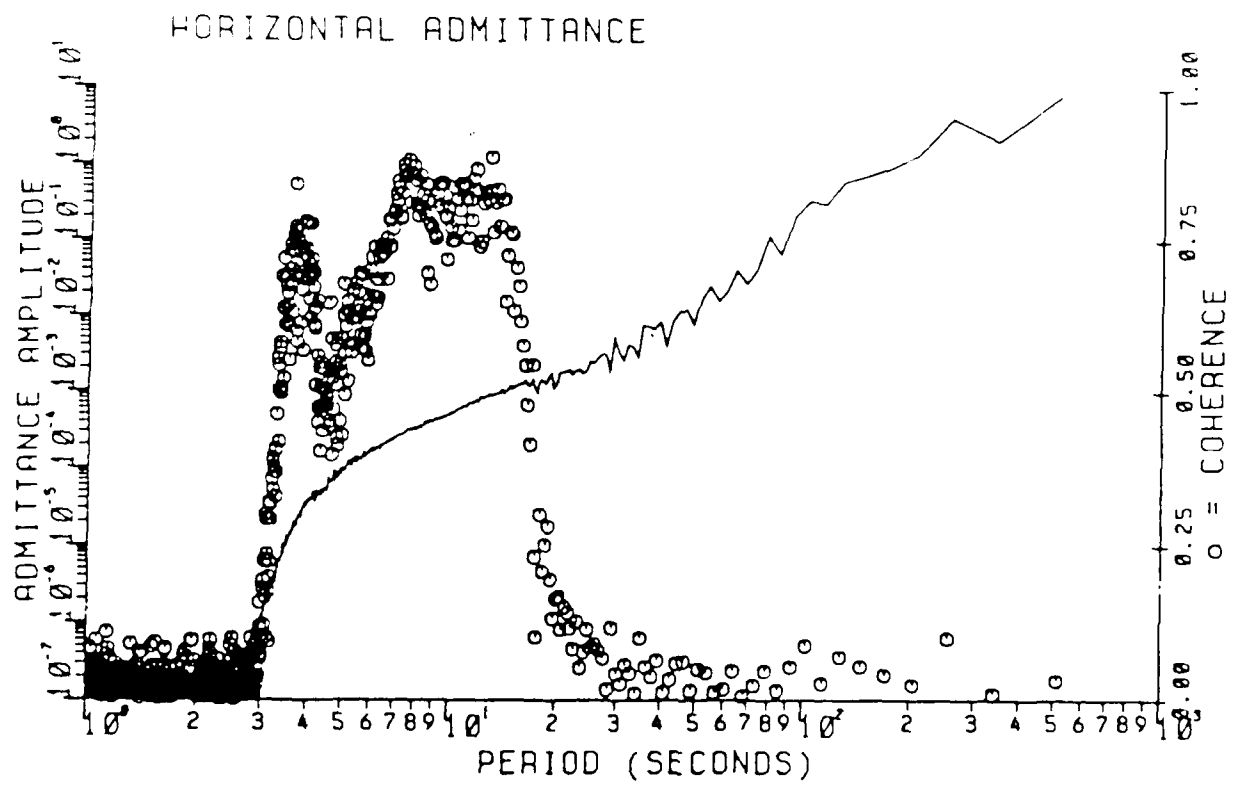
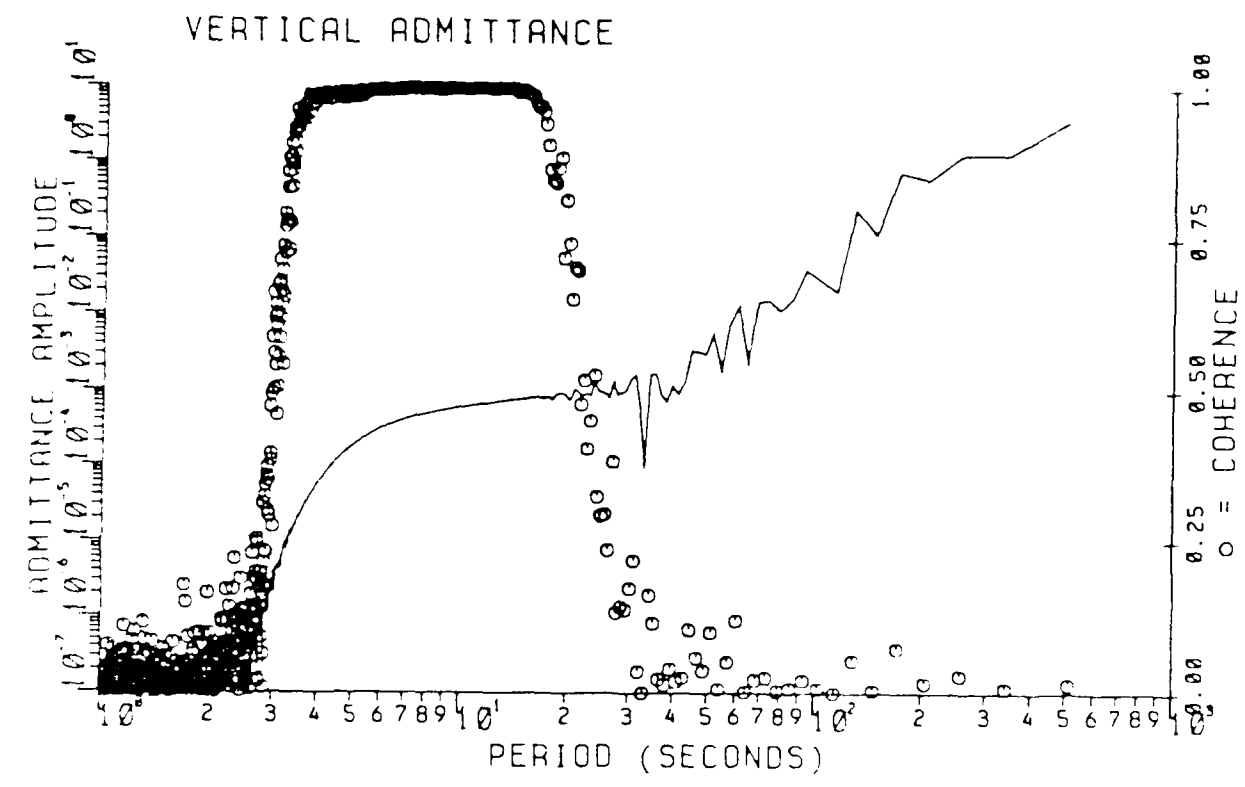


Figure 3. Vertical and Horizontal admittance spectra and coherence from a single OBS u at Atlantic Generating Station site.

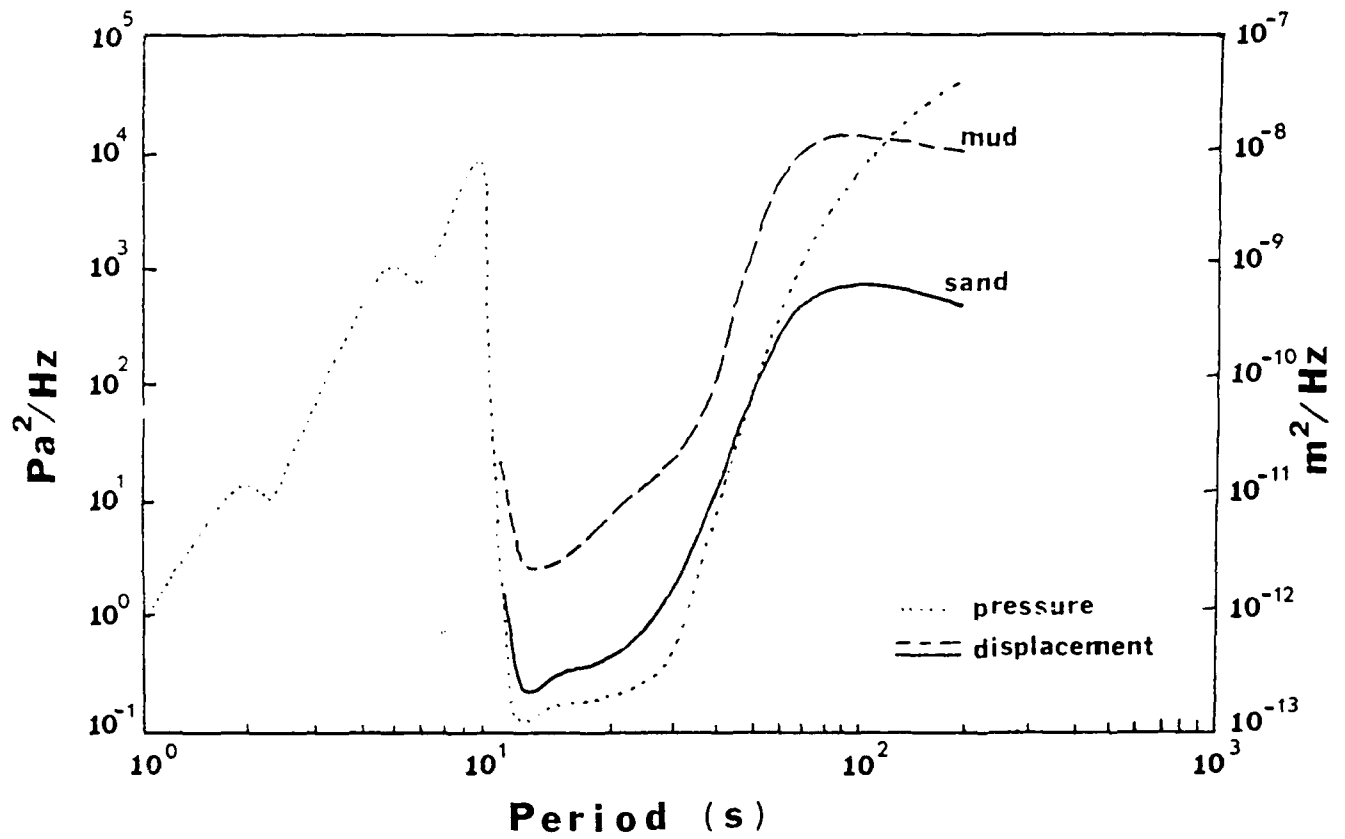


Figure 4. Predicted ULF seismic noise power spectra for sandy and muddy bottom at 4 km water depth.

Deep-Water Array Observations of Sea-Floor Noise

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ABSTRACT

We report initial analyses of data from observations of sea-floor noise (pressure and three-component motion) made in deep water off the California and Washington coasts using a closely-spaced two-dimensional minimally redundant array of sensors. We present coherences as functions of distance and frequency in the 10-156 m and 0.05-10 Hz range. The coherence below 0.4 Hz is independent of sensor spacing while above that frequency it falls off with distance at the 100 meter scale. Below 0.4 Hz the noise level correlates most strongly in amplitude and frequency with swell and above that frequency, with wind. The energy below 0.4 Hz probably travels as oceanic Rayleigh waves and above 0.4 Hz as Stoneley (interface) waves although the evidence supporting the latter statement is not conclusive.

The scientific questions vital to understanding sea-floor noise are:

- 1 What are the mechanisms of noise generation in the several frequency ranges?
- 2 What are the propagation mechanisms?
- 3 How does energy traveling by the various propagation mechanisms manifest itself on any particular sensor system?
- 4 What is the noise coherence as a function of space and frequency?

Two-dimensional arrays of Ocean Bottom Seismographs (OBS) can provide observations bearing directly on the last three questions and indirectly on the first. In pursuit of answers to these two questions, SIO and NORFAL have conducted two array experiments, one off San Diego in 3800 meters of water during April 1987, and the other off the Straits of Juan de Fuca in 300 meters water depth during July 1988. See Figure 1 for locations. The goals of these experiments were to establish the correlation lengths of the noise field and to see whether Stoneley waves (interface waves concentrated at the sea floor) are a significant part of the noise field. We present here preliminary results, primarily from the 1987 experiment named "CIRCUS."

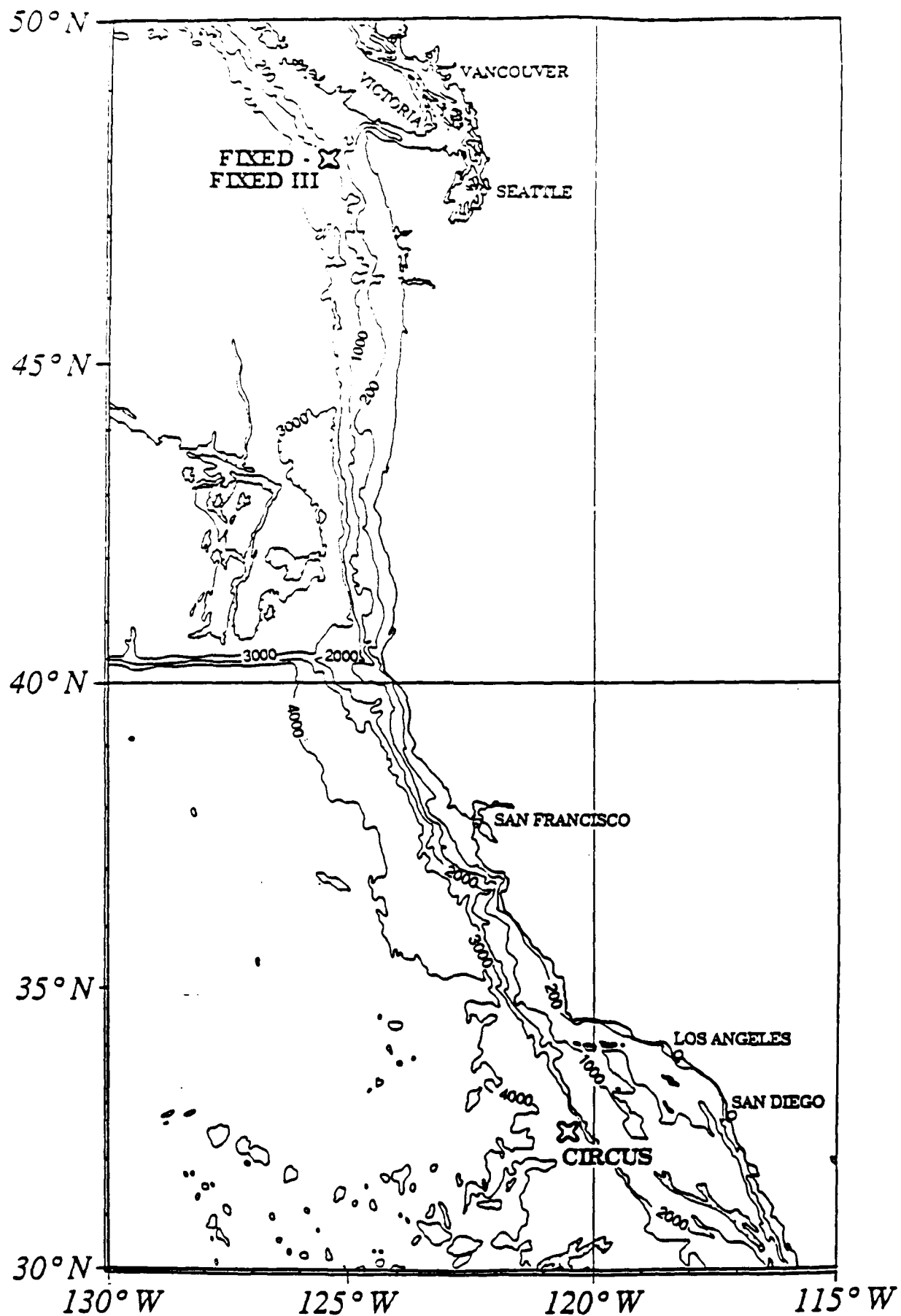


Figure 1. The locations of the two experiments discussed. The water depth of the CIRCUS experiment is 3800 meters while the instrument depths in the FIXED-FIXED III experiment ranged from 267 to 1143 Meters.

For many years the effectiveness of both active and passive sensing systems has been enhanced by using sensors in arrays rather than in isolation. The advantages deriving from the use of arrays stem from the ability of arrays to partition signal and noise by angle of arrival or, equivalently, spatial frequency, and from redundant sampling to allow averaging to reduce some types of noise. For arrays constructed on ships, the interaction between the sensors and the structure supporting is, at least in principal, calculable, and is, in any event, amenable to measurement and prediction. For arrays on or near the sea floor, the array performance is strongly dependent on the acoustic properties of the sea floor and these vary greatly from place to place.

The performance of an array, expressed as array gain, depends on the the type of processing, usually expressible in terms of the weighting of the individual sensors, and the statistics of the noise and signal fields, usually expressed through their correlations or coherences. See, for example Urick (1983). For simple delay-sum processing the array gain improves with increasing signal coherence and degrades with increasing noise coherence. If adaptive signal processing is used, noise coherence can be exploited by calculating array weights that cancel out noise but not signal. No matter how the data are processed, it is important to measure, understand, and predict these statistics and to use them in array design.

Figure 2 shows a more detailed view of the bathymetry of the CIRCUS site along with a reflection profile taken as part of the site surveying for the drilling of DSDP Hole 469. The hole location is within the array.

Earlier we (Sauter, Dorman and Schreiner, 1986) determined the seismic structure of the sediments at this site using sea-floor explosions as sources and Ocean Bottom Seismographs (OBSs) as sensors. These instruments (Moore et al., 1982 and Figure 3) record water pressure as well as three components of ground motion in the frequency range 0.05-32 Hz. The sea-floor shots generate interface waves (Stoneley/Scholte waves) whose dispersion is controlled by the shear velocity structure of the waveguide formed by the low-velocity sea-floor sediments. Figure 4 shows a seismogram including the well-dispersed interface modes. Note that the propagation velocity is very slow. Figure 5 is a time-frequency decomposition (sonogram) of one seismogram, showing the dispersion curves of two modes, the fundamental (mode 0) and an overtone (mode 2). A seismic velocity structure producing the observed dispersion is shown in Figure 6. The slow propagation velocities are caused by the low shear velocities in the upper sediments. These low phase velocities mean that the wavelengths of the associated modes are very short. This small scale, of course, has a profound effect on the design of array experiments since the sampling theorem must be satisfied at the shortest wavelengths present for each frequency of interest or spatial aliasing can occur.

Our array design is shown in Figure 7. When the number of sensors is limited and the desired ratio of array aperture to minimum interelement spacing is large, an array of minimally redundant design is appropriate (Haubrich, 1968). The design philosophy is that the set of vector interelement spacings should be reasonably uniform and that only one sample pair should exist with a given vector offset. This allows the maximum number of vector offsets to be generated with a given number of instruments. Use of such a configuration relies on the space-stationarity of the wave-field for its effectiveness. These arrays have proven effective on land for instruments used to detect and locate earthquakes and to monitor nuclear tests, but the validity of the assumption of space-stationarity of sea-floor noise is yet to be established. The site of the 1987 "CIRCUS" experiment was a sediment pond with uniform structure in the sediments so structural contributions to spatial non-stationarity should be minimal (at least for energy confined to the upper sediments.)

The instruments were emplaced using the apparatus sketched in Figure 8. The thruster device in the center is a product of F. N. Spiess and T. Boegeman at the Marine Physical Laboratory of the Scripps Institution of Oceanography. In operation, the OBS was suspended about 30 meters below the thruster

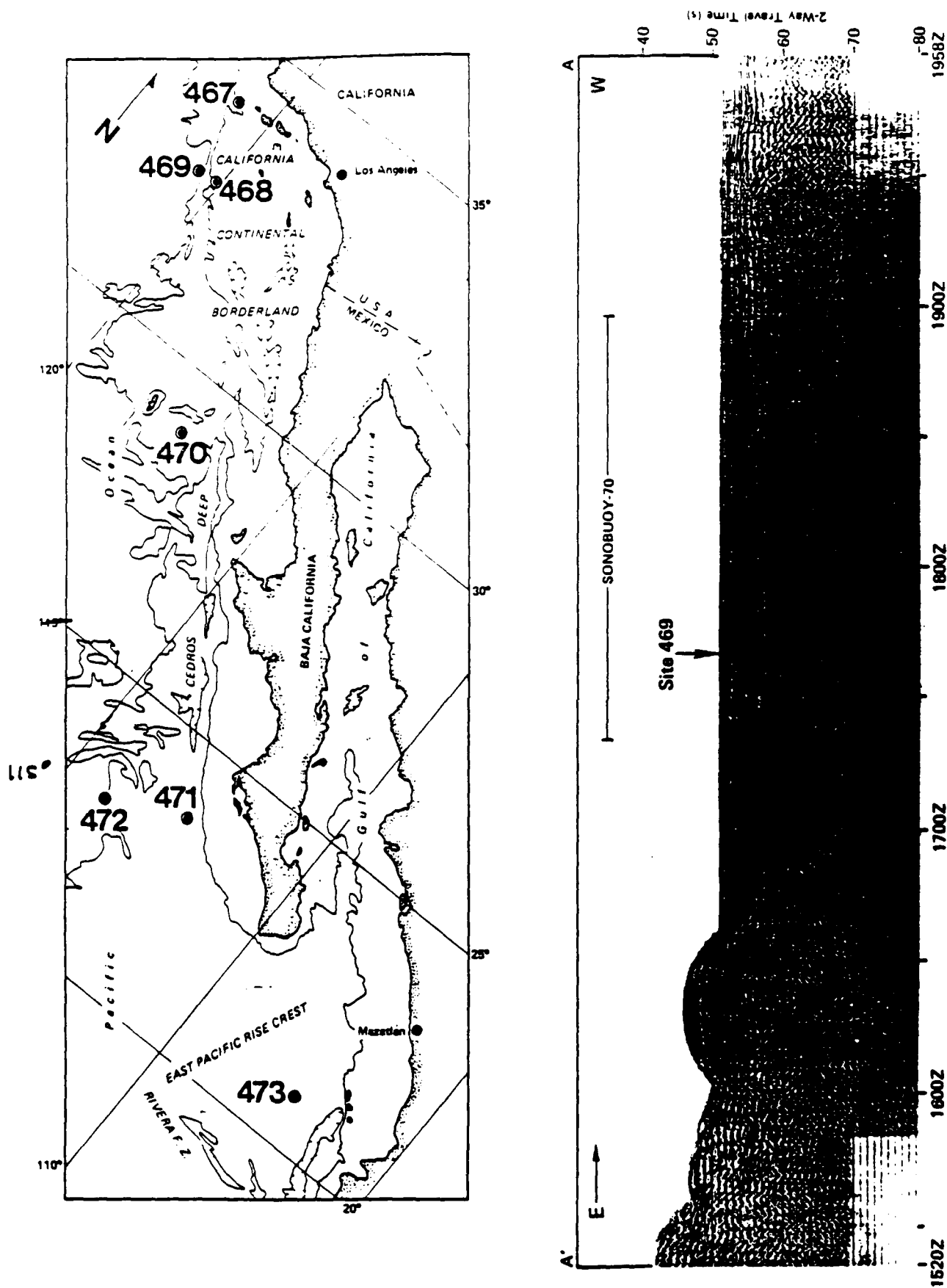


Figure 2. Regional bathymetry near and a seismic reflection profile over DSDP Site 469, the location for the CIRCUS experiment.

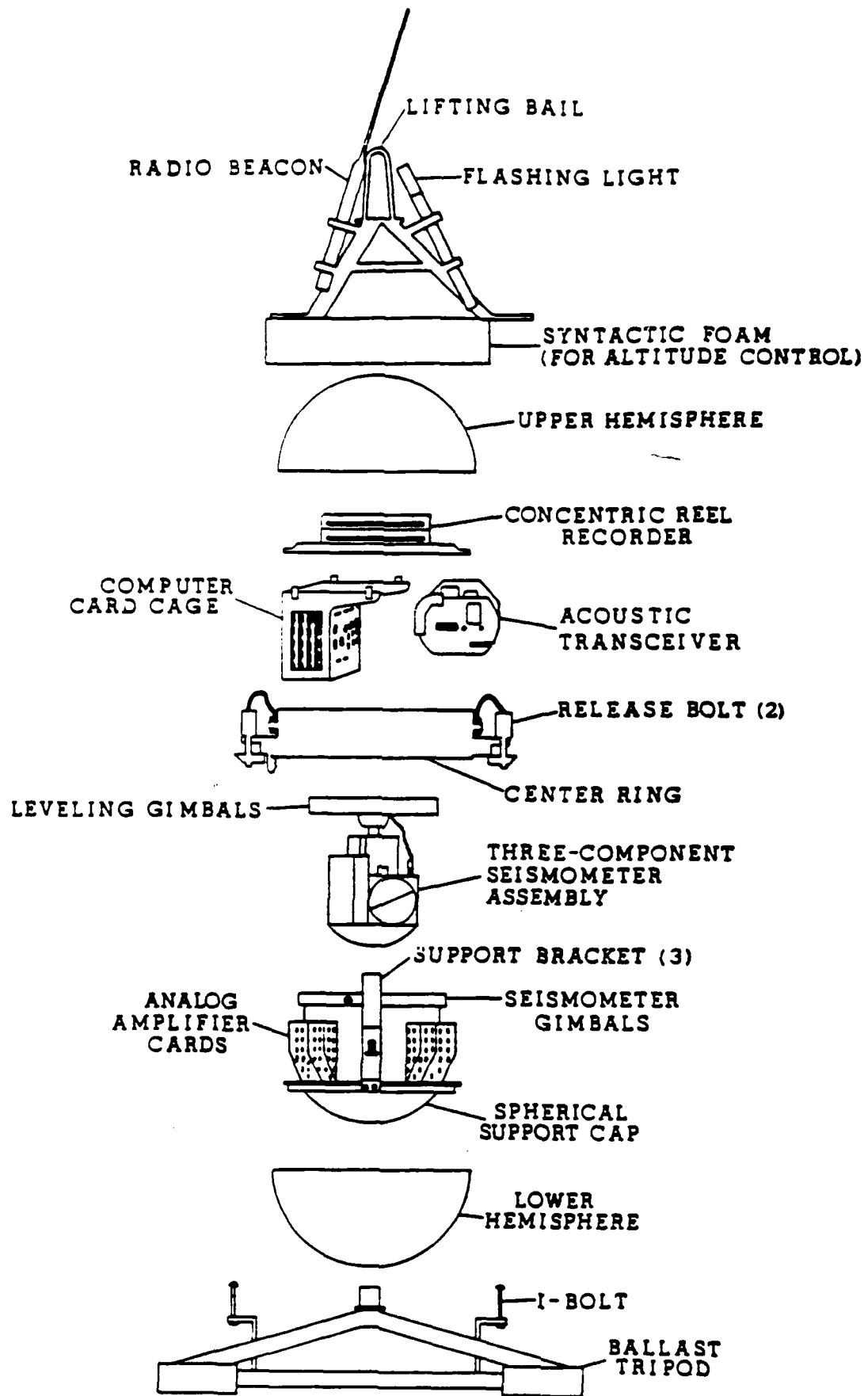
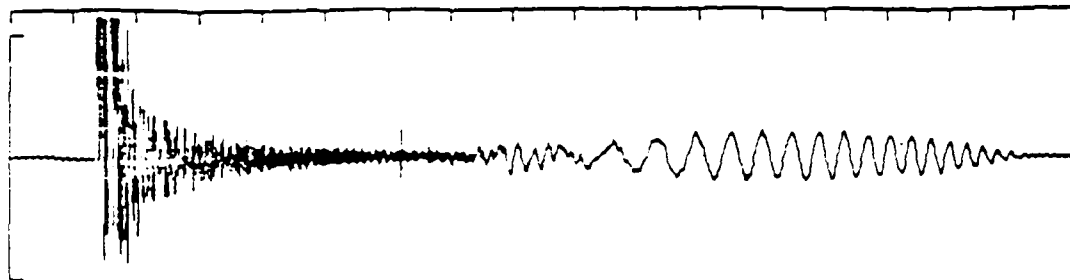
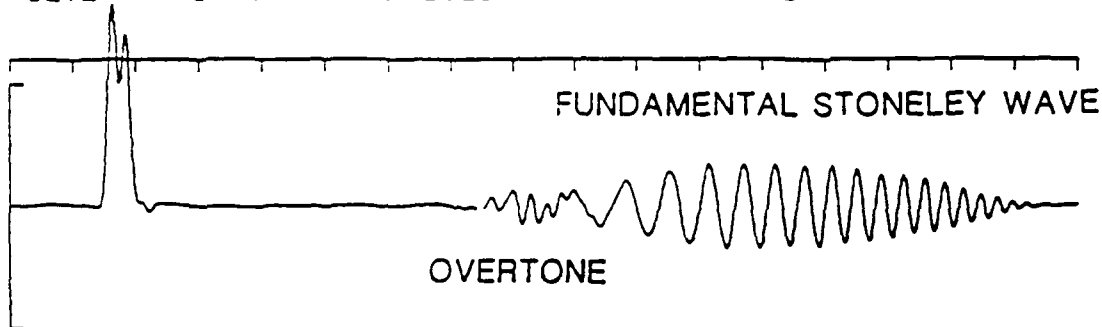


Figure 3. The SIO Ocean-bottom seismograph in a blow-up view.

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GAIN: 24



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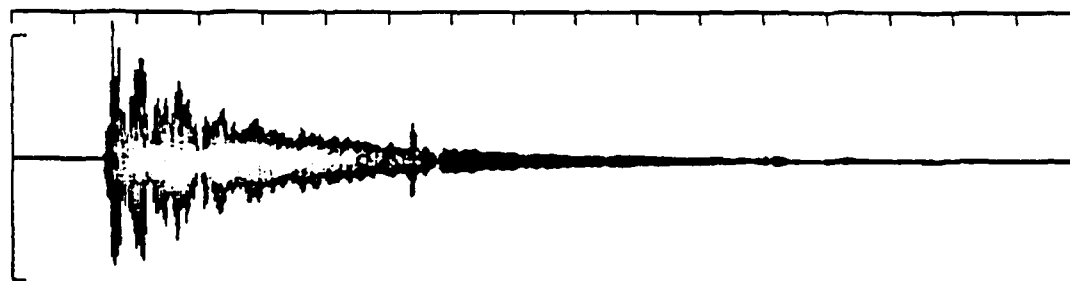
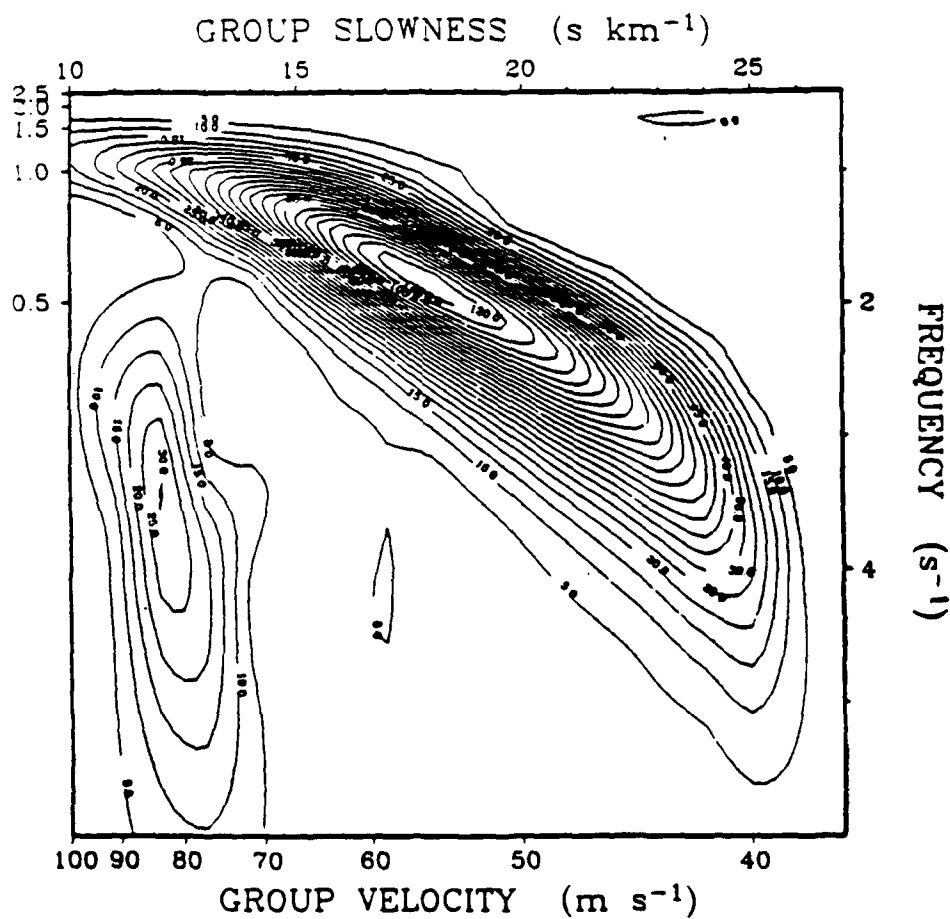
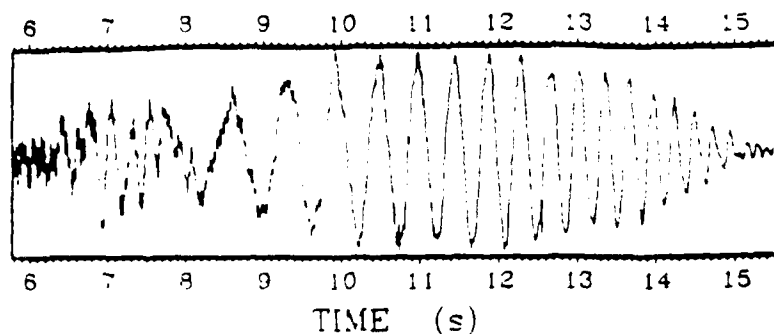


Figure 4. A seismogram of a Stoneley wave from a small sea-floor shot. The top trace is the unfiltered seismogram while the three below are bandpass filtered at 0.25-2.5 Hz, 10-25 Hz, and 20-30 Hz respectively.

INSTRUMENT 1 EVENT 9 COMPONENT 1
RANGE 0.578 km



AMPLITUDE

CONTOUR INTERVAL 5.0E-02

FILTER PARAMETERS: ALPHA 1.00 RELATIVE BANDWIDTH 0.25

Figure 5. A Time-Frequency decomposition of a Stoneley wave train from a small sea-floor shot. The large ridge is the fundamental (mode 0) while the near-vertical ridge is mode 2. Mode 1 is not strongly excited.

Model 11.63.02
Velocity (km/s)

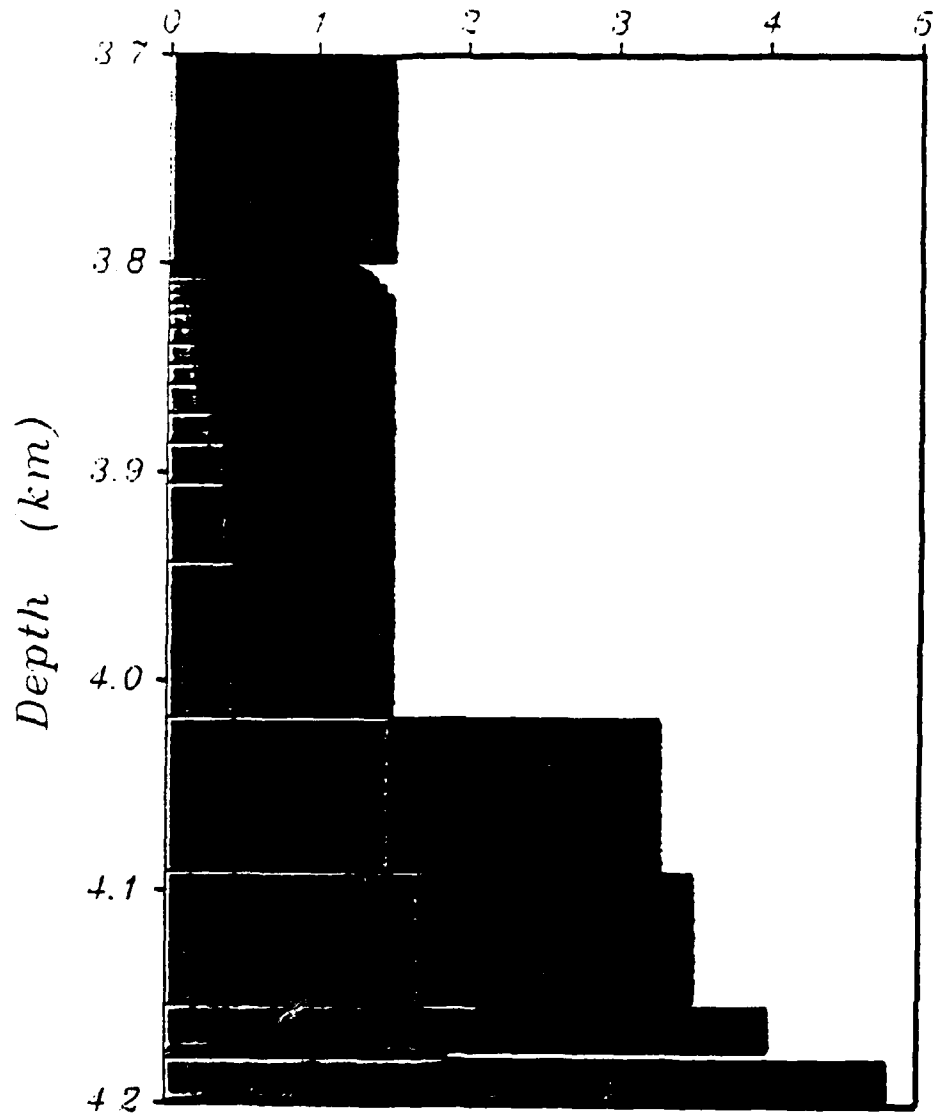


Figure 6. A velocity model derived from the Stoneley wave dispersion.

Circus Sensor Array

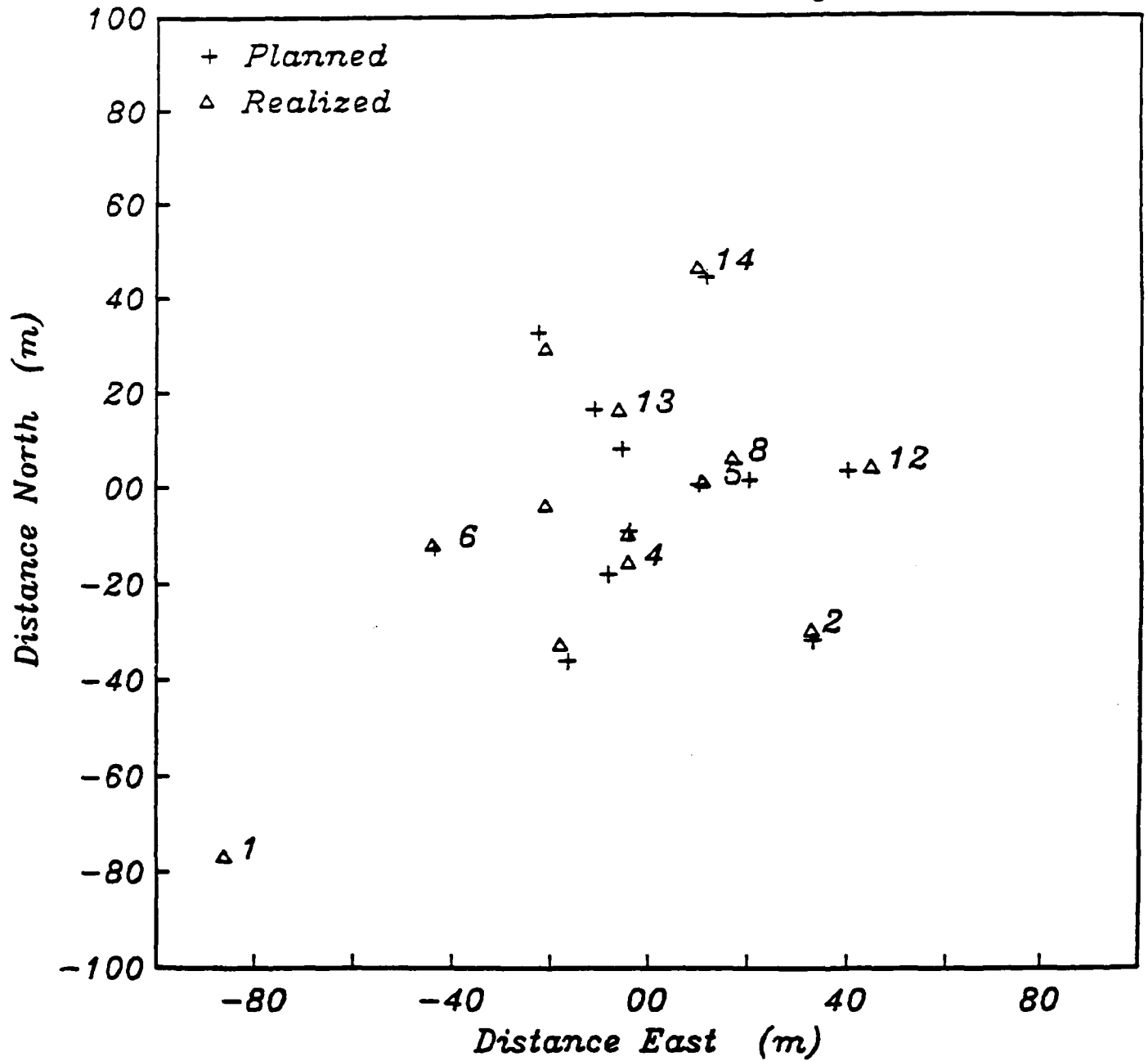


Figure 7. Instrument disposition for the CIRCUS experiment. The water depth is 3804 meters. Instrument 1, at the lower left, was dropped freely from the surface because of time limitations.

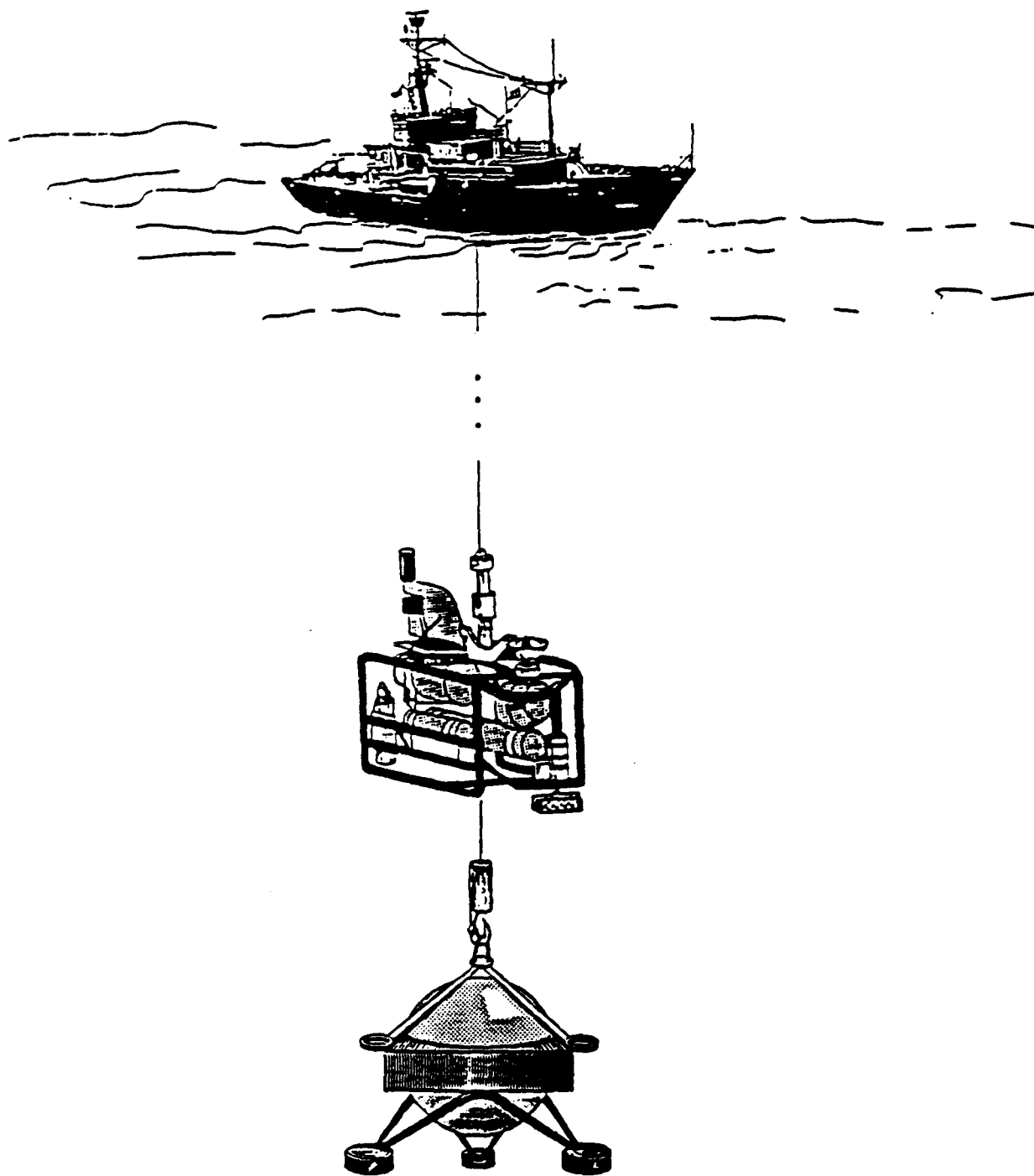


Figure 8. The Spiess-Boegeman thruster package set up for precision OBS deployment. The instrument, suspended from an electrically-operated release hook, is about 30 meters below the thruster package. The 0.680 coaxial cable is normally trailed from the A-Frame at the ship's stern rather than from the center, as shown in the cartoon.

and the assembly was lowered to within 100 meters of the sea floor. The thruster contains acoustic transponding equipment and its position is monitored from the ship with an accuracy of about a meter. The ship is maneuvered so that the OBS swings over the desired position and the OBS is released. In practice, two passes were usually required, and the total time for an emplacement was about eight hours per instrument.

The coherence as a function of frequency and interelement spacings is shown in Figure 9. Examining the coherence as a function of frequency, we see two distinct peaks. The lower peak extends from the lower limit of our instrument's frequency response to 0.4 Hz. There is no detectable decay with increasing sensor separation at the scales we observed. This peak is probably attributable to the oceanic Rayleigh waves and are the oceanic equivalent of the double frequency microseisms observed on land.

The higher frequency peak extends from 0.4 Hz to about 5 Hz. It begins to decay as a function of distance within 100 meters. Our initial postulate was that this peak is due to propagating Stoneley waves. We have performed wavenumber analyses on this energy with little in the way of consistent results thus far. We think this suggests that our time and/or position control is not adequate.

The instrument positions were determined by least-squares adjustment of the acoustic ranges observed between the ship and the instruments as well as between the thruster and the instruments. We estimate that the position accuracy is about 2 meters.

The time control for each instrument is a temperature compensated crystal oscillator (TCXO) whose drift is rated at a few parts in 10^7 . The uncertainty implied by this drift rate is certainly inadequate for beam forming at the high end of our frequency range but we have made time corrections using transients observed during the recording time (up to 8 hrs) over the deployment period of 31 days. The timing is, of course, more than adequate for the coherence calculations.

We had hoped to see evidence supporting interface wave propagation in the frequency-wavenumber spectra. That is, we had hoped to see, at some given frequency, propagation at the velocities corresponding to the several modes of the interface waves. In a laterally uniform medium, it is difficult to couple energy into a strong waveguide from an energy source; in this case from the sea surface, outside the waveguide. The real sea floor, however, contains inhomogeneities that can scatter high-velocity energy into the sea-floor waveguide. Although the sedimentary structure within the sediment pond is smooth, the pond is small and the irregular surface of the crystalline basement bounds the sediment pond from below and from the sides. Levander and Hill, 1985, have made finite-difference simulations of the effect of irregularities in the thickness of the boundary of a low-velocity layer and shown that the inhomogeneities cause energy from an incident high-velocity plane wave to be scattered into the waveguide and that the scattered energy travels at the mode velocities of the waveguide. Their calculations were based on a two-dimensional treatment of the terrestrial analog, the surficial weathered layer overlying the crystalline crust but the physics is similar to the sea-floor case.

There are several possible explanations for the conditions we see. It may be that we are using insufficient averaging so that the cross-spectral matrix we are generating is not sufficiently stable. We have stabilized our cross-spectral calculations by using averages of 20 to 40 blocks, numbers that are adequate when the noise field is random. It may be, however, that the basin is "ringing" with a long time constant and that the blocks of data we are using are not truly independent. Here some of the degrees of freedom we thought we had may be illusory.

A more dramatic possibility is that our interpretation of the physics is totally inappropriate. The noise source that is both energetic and nearby is the gravity ocean waves at the sea's surface. The short wavelengths of ocean waves at these frequencies causes their wavefunctions to decay rapidly with depth. The non-linear wave-wave interaction (Miche, 1944; Hasselmann, 1965; Longuet-Higgins, 1950), can

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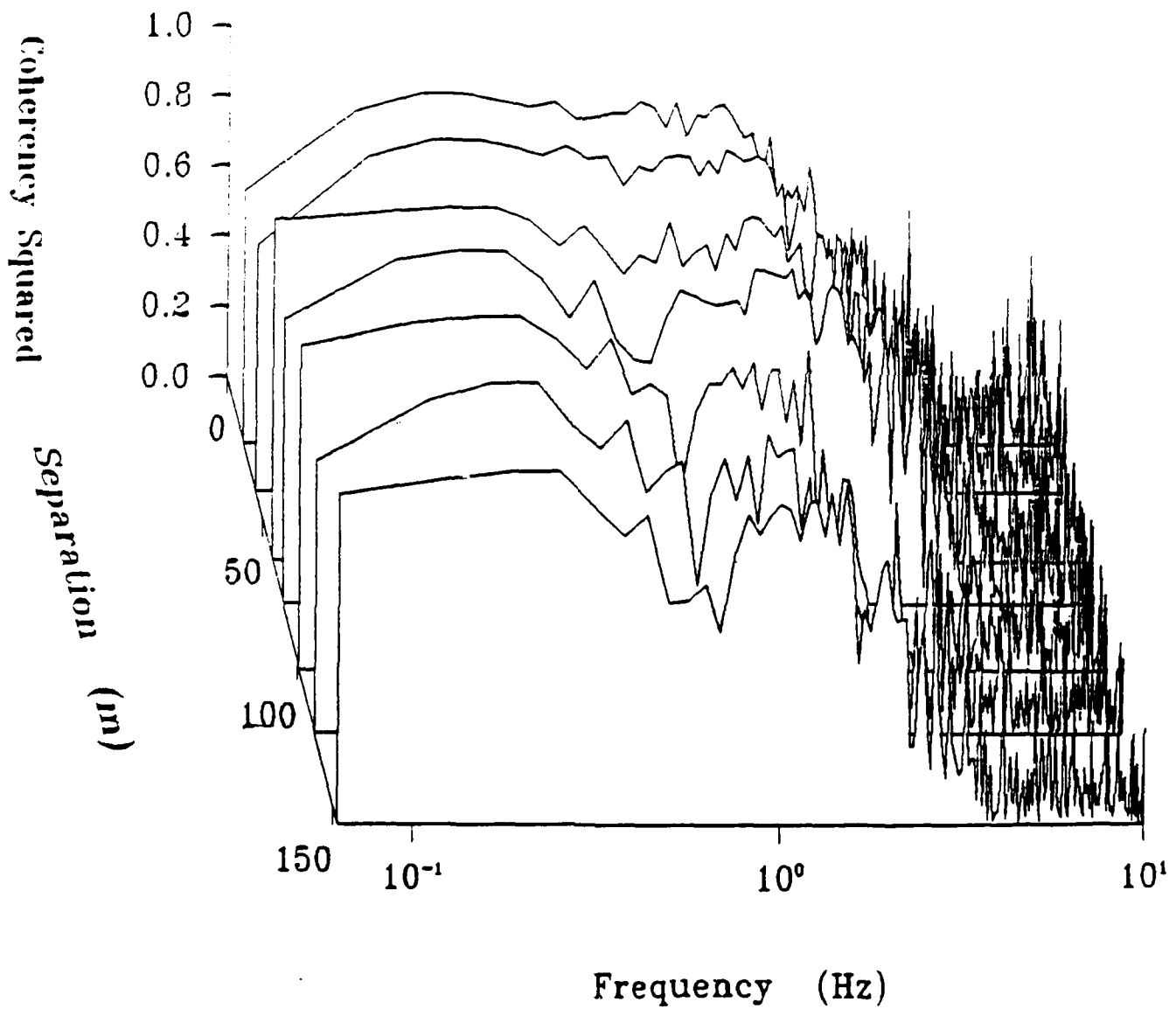


Figure 9. Coherence between vertical components as a function of frequency and sensor separation. For frequencies lower than 0.4 Hz the coherence is essentially distance-independent while above that frequency there is decay with distance with a length scale of a few hundred meters.

however, create a wavefield containing components that propagate in the water and that can reach the sea-floor with little attenuation. To our knowledge, the correlation function of this wavefield at the sea-floor is unknown. Indeed, spectra and coherence of the ocean waves themselves at these frequencies are poorly known. It may be that the coherence we see is dominated by the correlation function of the incident field from nonlinear wave-wave interactions at the sea surface other than by sea-floor propagation phenomenon.

Power spectra for vertical and horizontal accelerations and for pressure are shown in Figures 10, 11, and 12. Note that the ratio of vertical to horizontal motion is different above and below the frequency (0.4 Hz) of the notch in the coherence plot. This lends support to the idea that propagation mechanisms above and below that frequency are significantly different.

We have obtained hindcasts of wind and ocean swell for this location from the Fleet Numerical Oceanography Center and compared these with power spectra gathered over a month. There is evident correlation between the swell in the frequency range 0.05-0.2 Hz and sea-floor noise at twice that frequency, as demonstrated at the coast by Haubrich, Munk, and Snodgrass, 1963. At higher frequencies (above the 0.4 Hz notch) the noise is seems controlled by the local wind velocity. The effect of the saturation of the wind wave spectrum noted by McCreery (1989) is clearly evident in our data (see Figure 13).

Window 237 Vertical Acceleration

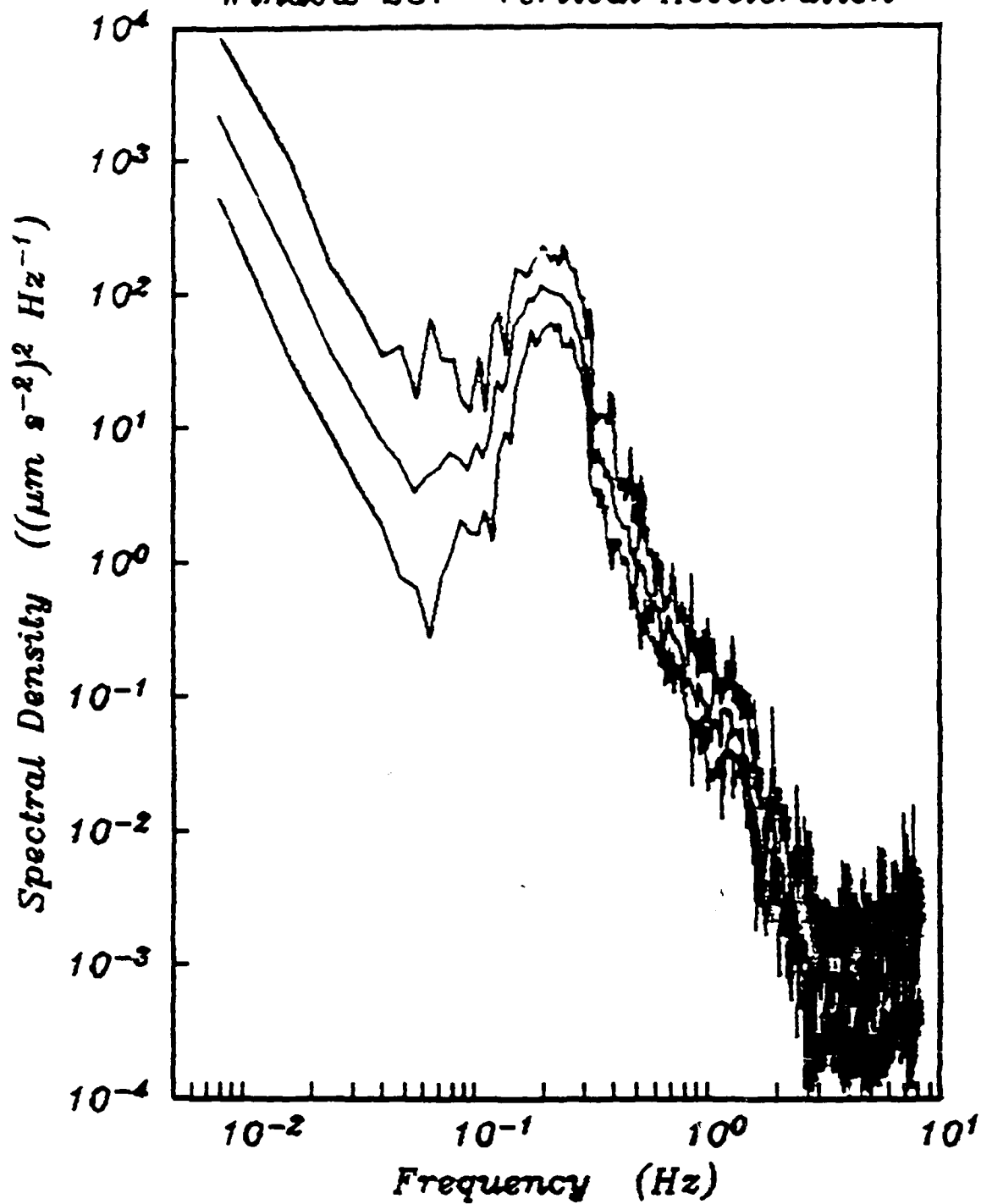


Figure 10a. Vertical acceleration for one 60 second window of data from the CIRCUS site. The spectrum was calculated by the multiple-window method using a time-bandwidth of 4. The error bounds are the 90% confidence levels calculated by the jackknife method.

Circus vs. Fixed-Fixed 3

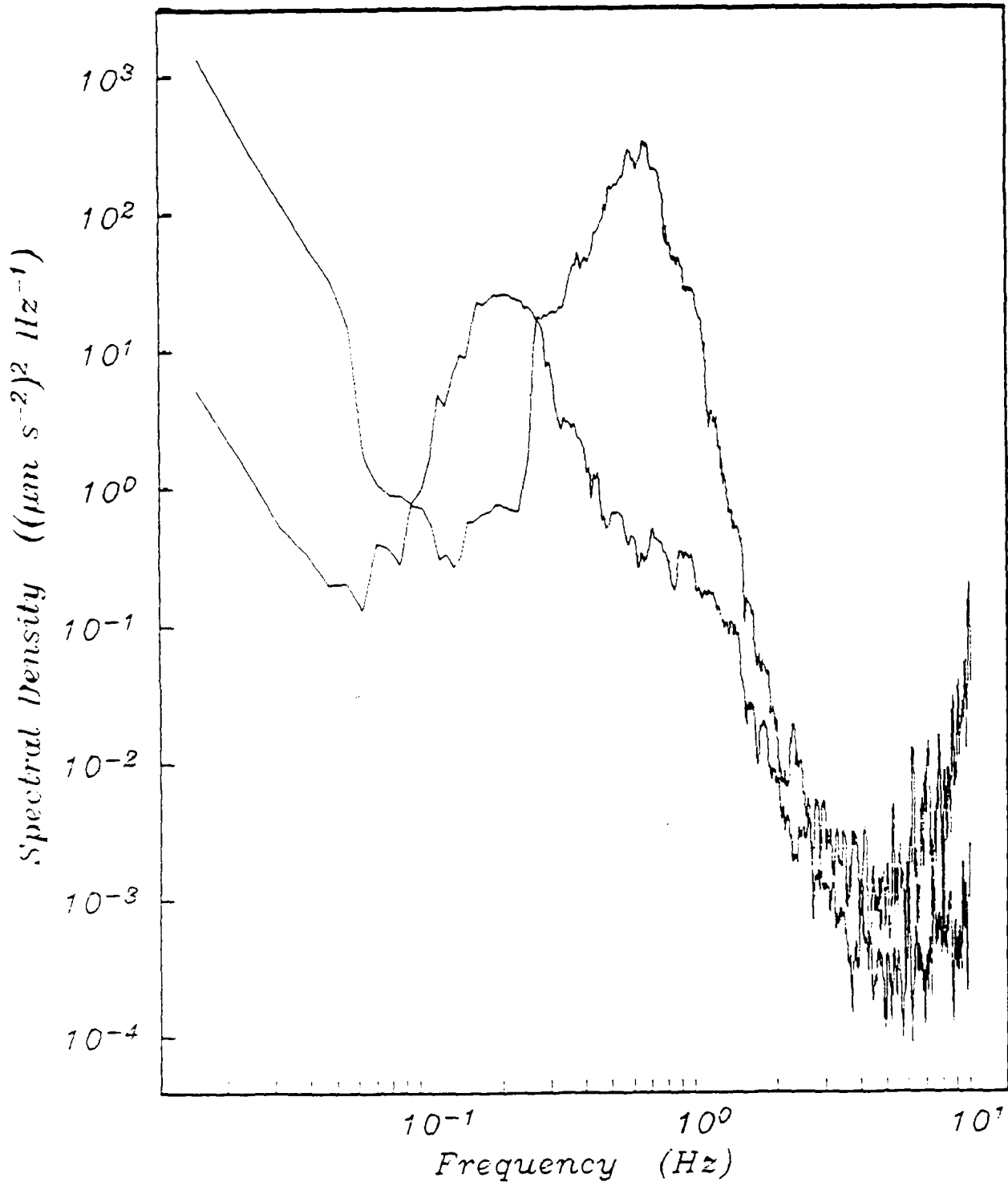


Figure 10b. Vertical component spectra from the CIRCUS and FIXED-FIXED III experiments. The spectral levels are similar at 0.1 Hz and in the 1.5-4.0 Hz range. Below 0.05 Hz the spectra are probably showing instrument noise. The low-frequency instrument noise for the F-F III site appears worse because the high spectral levels near 0.5 Hz force the instrument to reduce the preamplifier gain and thus degrade low-frequency noise performance.

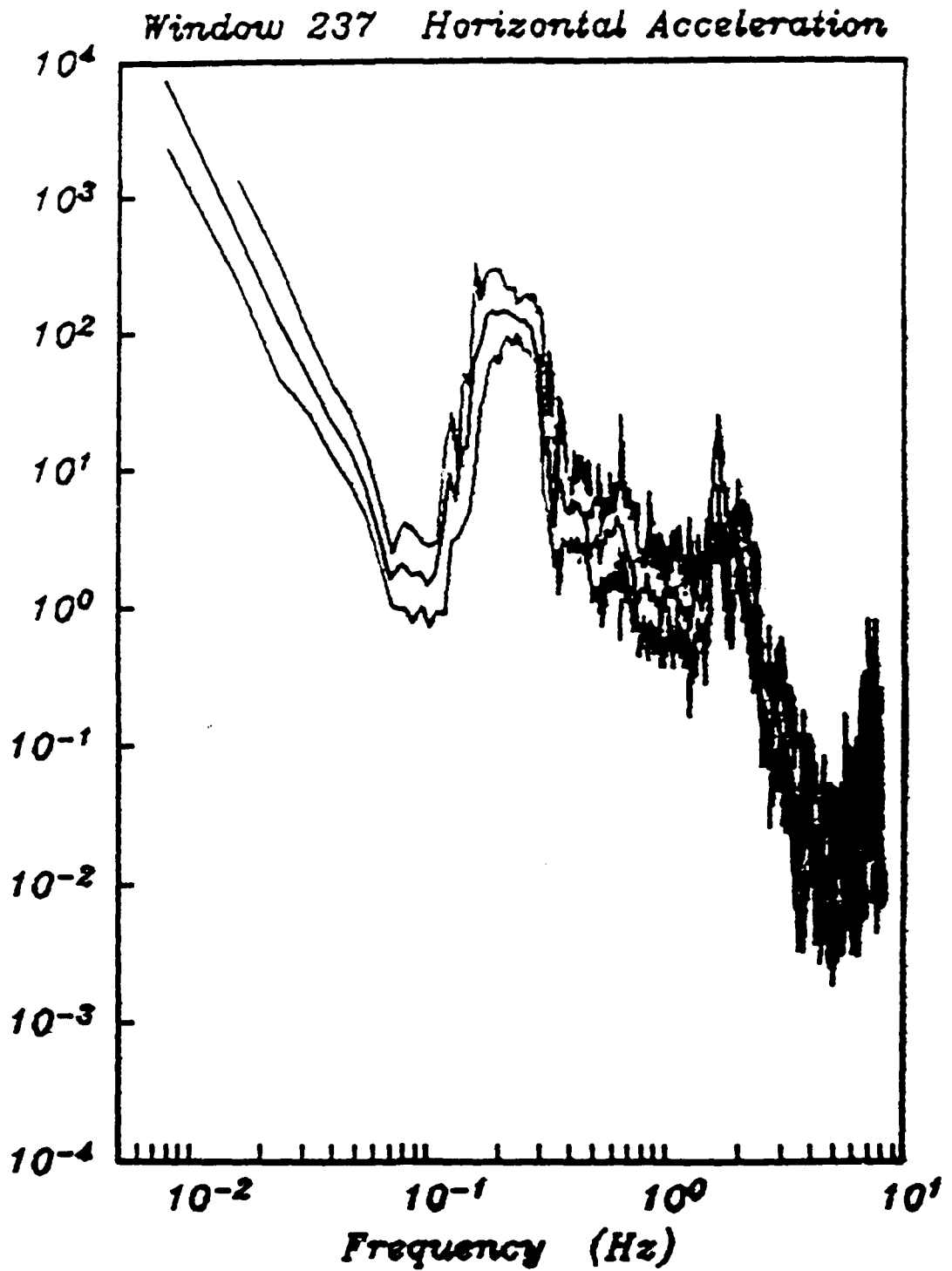


Figure 11. As Figure 10a except for a horizontal component. Note that the vertical and horizontal are similar below about 0.4 Hz but the horizontal becomes larger above that frequency. This may indicate that the mode structure of the noise is different above and below that frequency.

Window 237 Pressure

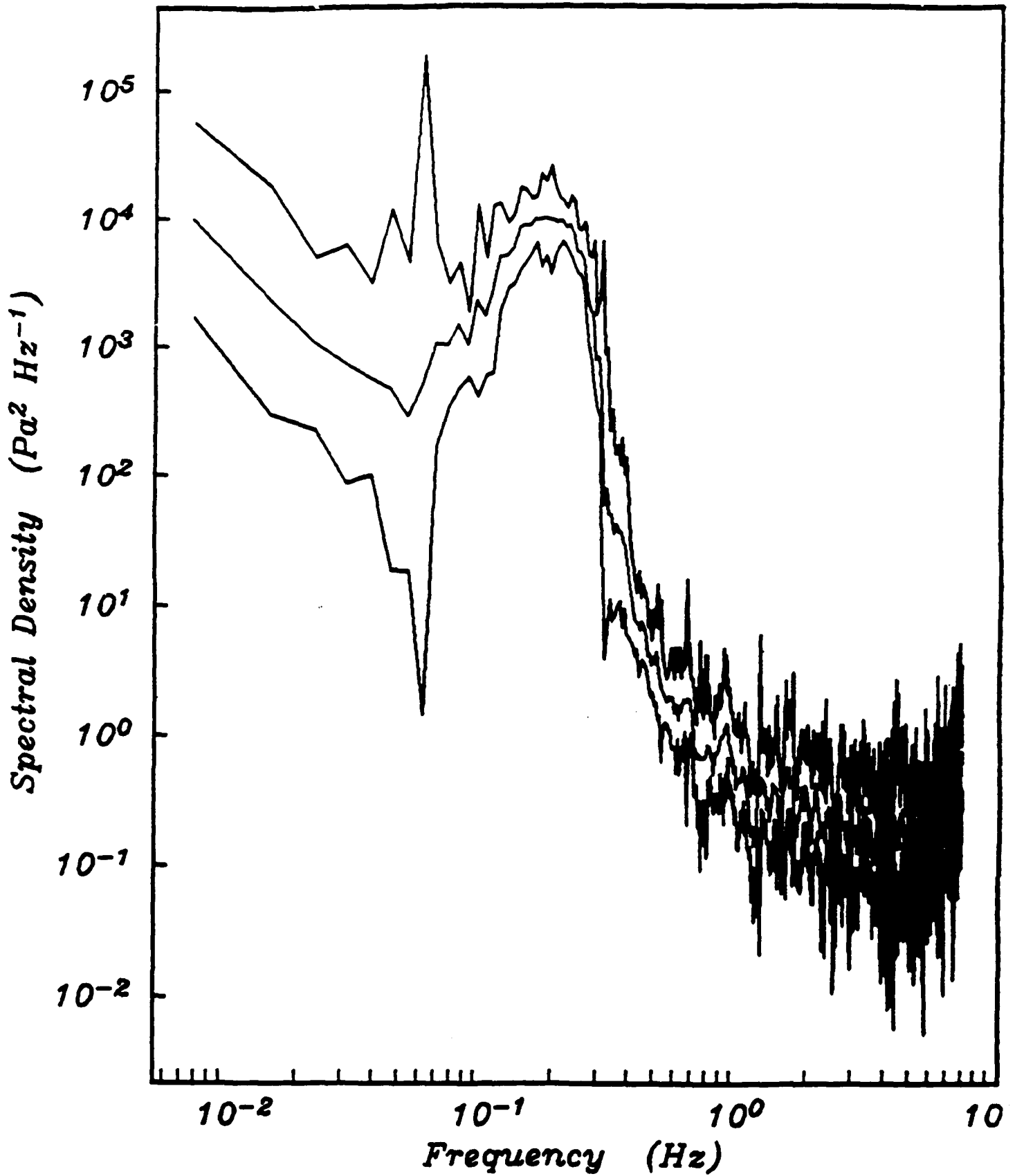


Figure 12. As figure 10a except for pressure.

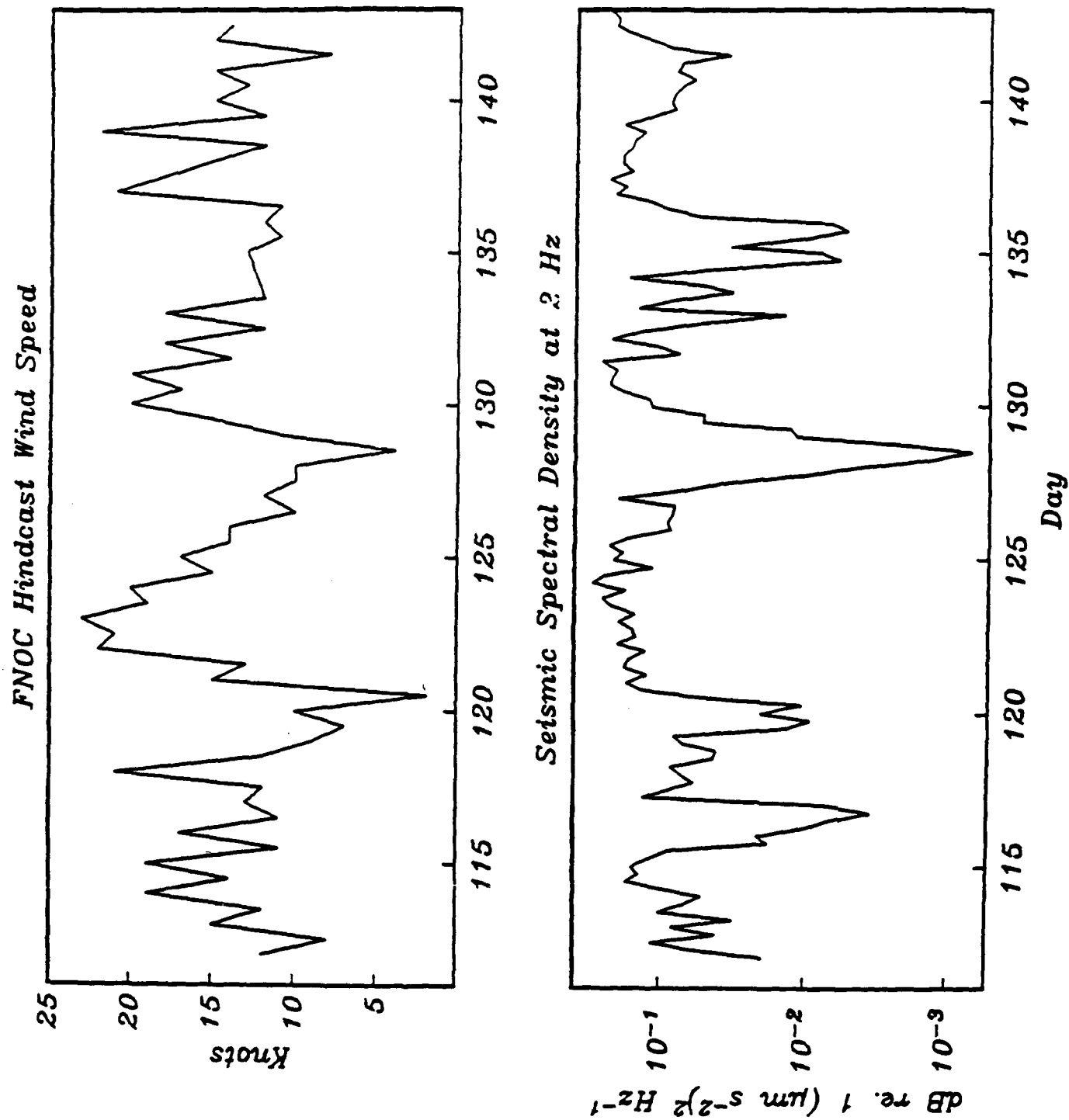


Figure 13. The upper panel shows windspeed from the FNOC meteorological hindcast while the lower panel shows spectral density of vertical component noise for the same time period. If you imagine the upper trace to be clipped at about 14 knots it strongly resembles the lower trace, in agreement with the observations of McCreery (1989).

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THE 1989 LOW FREQUENCY ACOUSTIC-SEISMIC EXPERIMENT

27 November 1988

John A. Orcutt

Peter M. Shearer

The Low Frequency Acoustic-Seismic Experiment (LFASE) is a scientific endeavor scheduled to take place in the spring and early summer of 1989. The major objective of this experiment is to develop a better understanding of the physics of the excitation and propagation of low frequency noise (0.01 - 50 Hz) immediately above, at and below the seafloor. In addition to these noise experiments, we shall conduct signal experiments using a variety of impulsive and oscillatory sources at the ocean surface. Data from these signal experiments will delimit the elastic properties of the bottom for use in the noise studies as well as provide unique data for understanding the attenuation of sound in the coupled ocean-seafloor system.

The investigators will develop and exploit a new ocean technology to locate and probe DSDP holes with a maneuverable, tethered deep submergence vehicle. Using this technology they will emplace a multi-node seismic sensor within the cased portion of DSDP borehole 418. The overall system will consist of the borehole, three-component inertial sensors and borehole hydrophones as well as ocean bottom seismographs and a vertical hydrophone array. The experiment will be preceded by a visit of the re-entry system to the borehole to determine the condition of the re-entry cone using sonar and photographic means as well as a re-entry of the hole with a caliper log.

The actual LFASE experiment will consist of two complementary stages. In the first stage, the R/V Melville will emplace the instrumentation on the seafloor and within the borehole while remaining coupled to the borehole seismic and acoustic sensors through the re-entry vehicle and its tether. Other ships will shoot a series of radial and circular lines using airguns, explosives and tuned sources to provide data required to characterize the seafloor including the sediments, crust and uppermost mantle. The subsequent data analyses will employ a full suite of techniques for determining the vertical elastic properties of the seafloor as well as the anisotropic behavior of the ocean crust and uppermost mantle.

The second stage of the experiment is designed to provide recordings of long time series of unadulterated seafloor noise in the absence of ships. The R/V Melville will divorce itself from the borehole sensors and return to port with the shooting ship. The ocean bottom seismographs and the borehole sensor recording systems are presently being modified to provide several gigabytes of recording capacity in order to allow nearly continuous seafloor recording. Data from all the sensors will be jointly examined to develop a full understanding of the noise at the bottom. The R/V Melville will return to the recording site after several weeks to recover the seafloor apparatus and extract the borehole array from DSDP Hole 418.

Overall coordination for the program is provided by the Johns Hopkins University Applied Physics Laboratory with assistance by a group of scientists from government and private organizations including the Science Applications International Corporation (SAIC), the Naval Oceanographic Research and Development Activity (NORDA), Woods Hole Oceanographic Institution (WHOI) and the Scripps Institution of Oceanography (SIO).

Fiscal Year 1988 tasks include the purchase (from CGG of France) of the multinode and multicomponent broad band seismograph for emplacement in the seafloor (WHOI/MIT), design and construction of the bottom control unit for the array (WHOI), the updating of the electronics, timing and recording capacity of available ocean bottom seismographs (SIO and NORDA), developing a modern vertical hydrophone array

(NORDA), and the preparation of a Remotely Operated Vehicle (ROV) for borehole re-entry (SIO).

The Ocean Drilling Program (ODP) and the Joint Oceanographic Institutions, Inc. (JOI) have supported related research objectives and planning for future experiments. The JOI U.S. Science Advisory Committee (USSAC) sponsored a workshop in 1987 entitled *Science Opportunities created by wireline re-entry of deepsea boreholes* and the USSAC Program Plan calls for the development of a wireline re-entry system for general seafloor use during the next three years and a Request for Proposals for the development of such systems was published in summer, 1988. Borehole seismometry and sub-seafloor instrumentation were the subjects of another JOI-USSAC workshop, held at Woods Hole in April 1988, *Permanent Ocean Bottom Geophysical Observatories*.

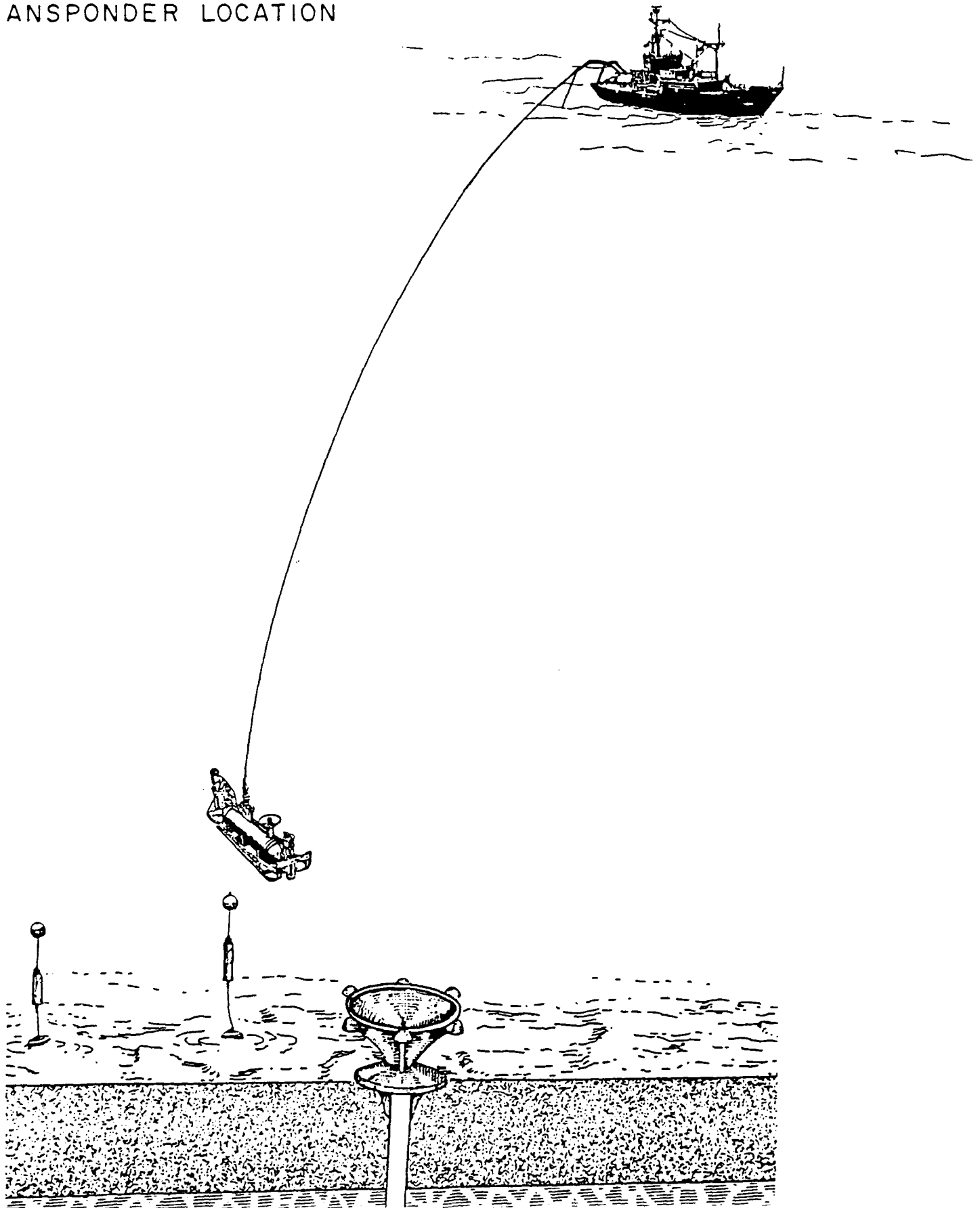
This project is made possible by the successes of the Deep Sea Drilling Project (DSDP) and the Ocean Drilling Project (ODP) which have been sponsored by the National Science Foundation and several non U.S. partners. This research follows directly from the earlier DSDP studies in the Atlantic in a re-entry and recovery from Hole 395A in 1981 (Leg 78B) and in the Pacific at Hole 581 (leg 88, 1982) and the later Ngendei Experiment (Hole 595B during Leg 91 in 1983). These earlier experiments were funded by the Defense Advanced Research Projects Agency (DARPA) and this agency is providing a share of the funds for this experiment. The U.S. Navy through OP-21 and the Office of Naval Technology are the other sponsors. The development of reliable and affordable deep sea maneuvering systems that can operate from conventional research ships will extend the scientific yield from the seafloor boreholes. The ODP regularly exploits the holes drilled in the seafloor from the D/V JOIDES Resolution through petrological, geochemical and paleomagnetic studies of the samples and logging, electrical and seismic studies of the holes. These decades of studies recognize that the existing boreholes are a scientific legacy that are available for further exploitation. Studies such as LFASE are required as ocean scientists seek to exploit seafloor measurements in the global study of the Earth through the deployment of long term observatories.

The first actual tests of a re-entry system were carried out in France using the submersible Nautile in 1986. The French approach used a special frame (NADIA - Navette de Diagraphie) fitted with a logging winch and 1,000 m of cable which was docked in the re-entry cone by the submersible. The next step in the French program was to re-enter DSDP Hole 396B in the Atlantic in late 1988. Scientists at the Pacific Geoscience Centre in Canada intend to use an advanced ROV for re-entry with a NADIA-like system. At a later stage, the Canada group would use the ROV to guide instruments, suspended from a surface ship, into a re-entry cone. This is very similar to the approach being taken in LFASE.

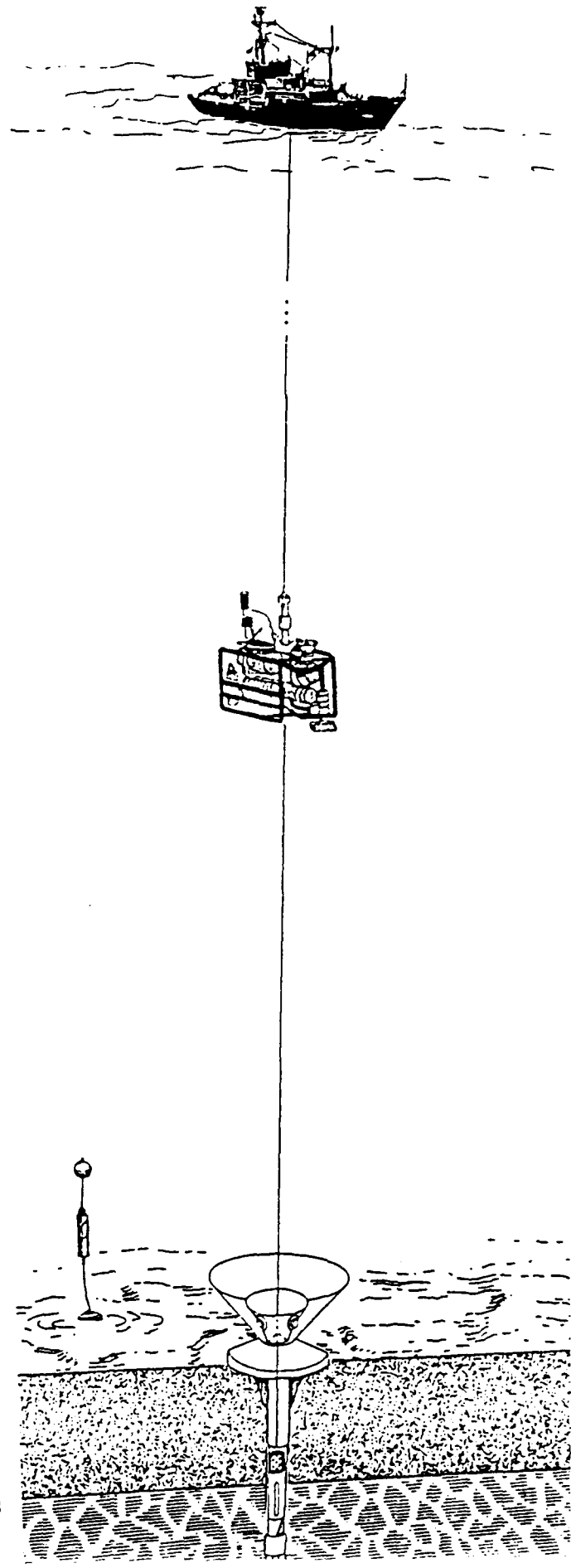
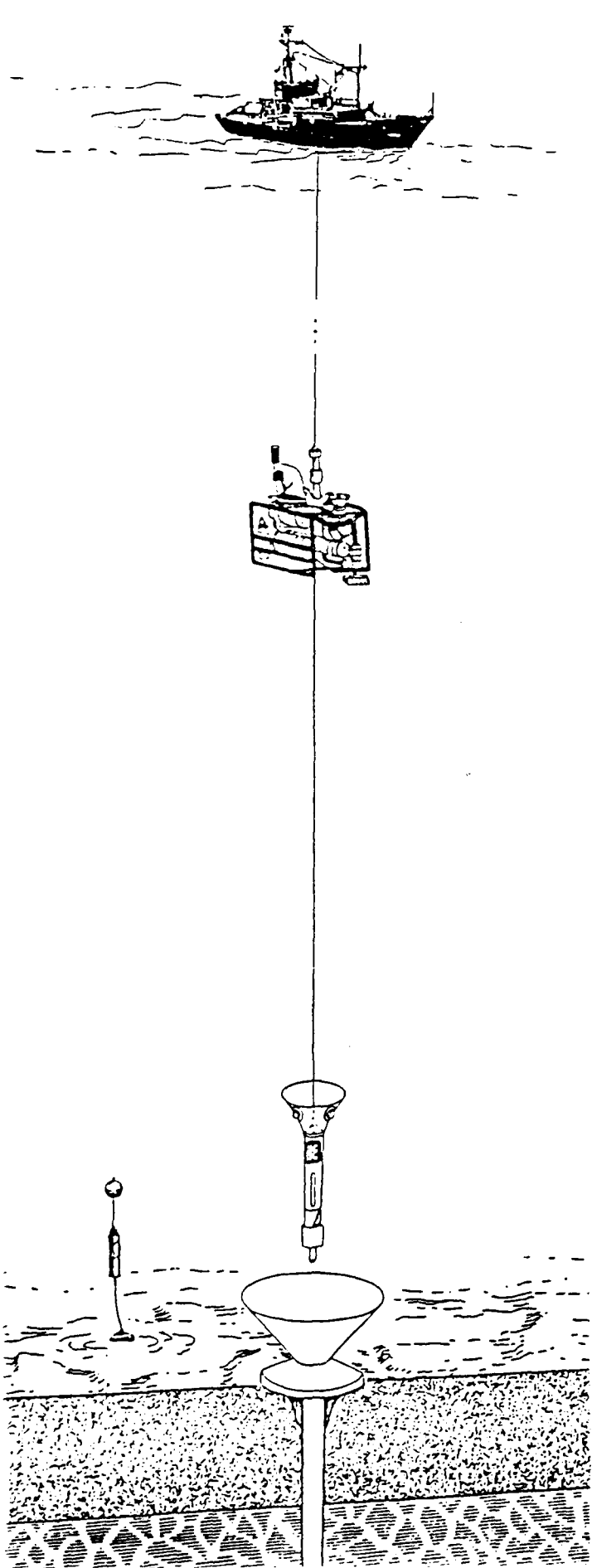
The series of figures provided by Dr. Fred Spiess of the Scripps Marine Physical Laboratory illustrate the insertion of the borehole array in DSDP Hole 418A. Figure 1 illustrates a Deep Tow survey of the borehole which will be conducted in April, 1989. The Deep Tow *Fish* will be located within a seafloor transponder array. The array will remain on the bottom through the final stages of the main experiment. Figure 2 is a schematic of the hole reconnaissance which will be conducted in April to ensure that the hole is open and not bridged by debris or sediments. The seismic nodes will be inserted only in the cased portion of the hole. The reentry probe contains a pair of caliper logs, a low-light television camera and lights and an acoustic transponder for location. This system was fully tested, with a set of refurbished ocean bottom seismographs in October, 1989. Figure 3 depicts the thruster package with the bottom recording package and attached seismic nodes. The ocean bottom seismographs and vertical hydrophone array will be placed on the seafloor in the vicinity of the hole using the Scripps' thruster package. The wire to the ship will be

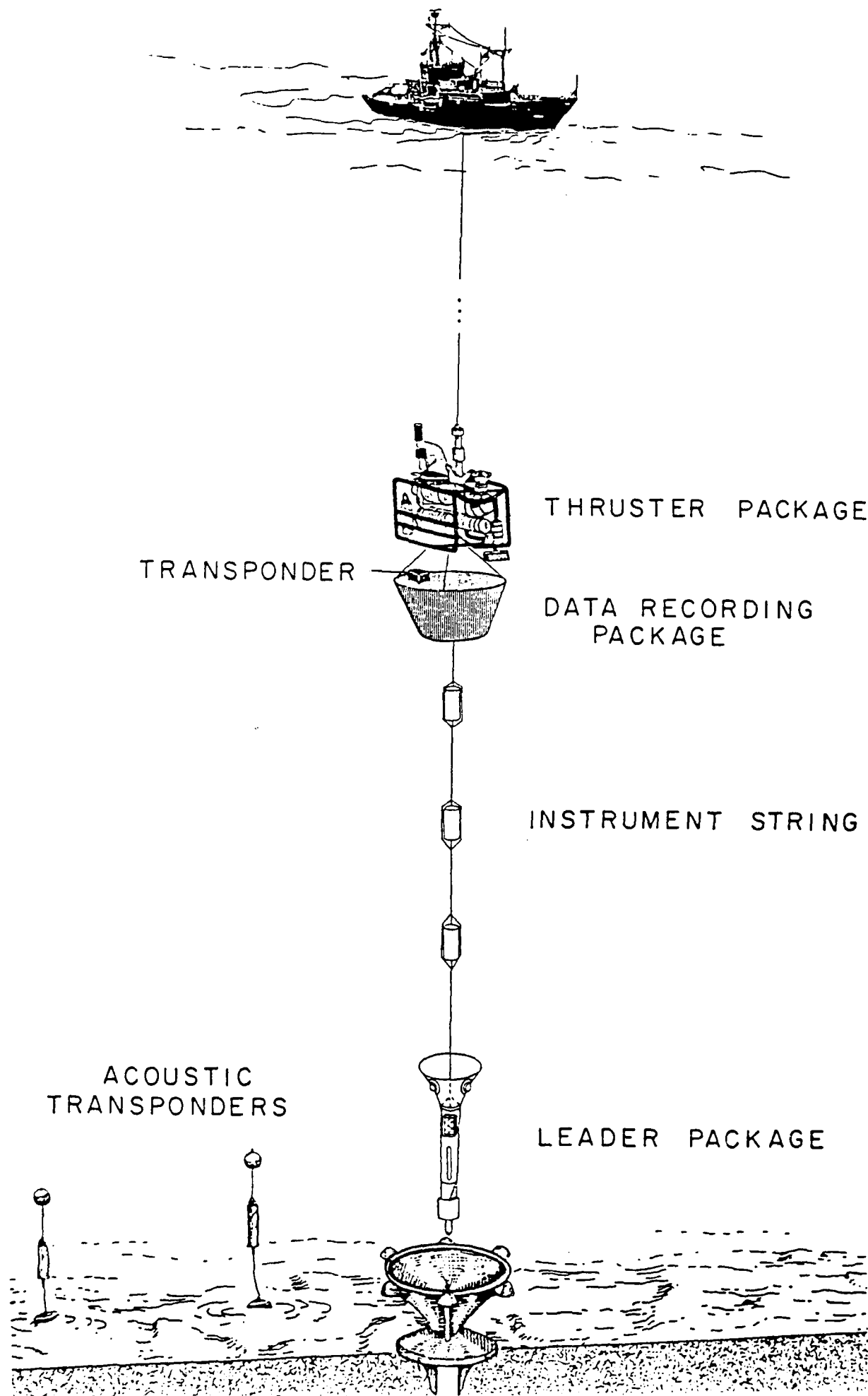
used to telemeter data to the R/V Melville for the active phase of LFASE as shown in Figure 4. The thruster will be detached from the array for the noise portion of the experiment and the ships will leave the area for an extensive term of seafloor recording. The borehole array and data recording package will be recovered as shown in Figure 5 at the end of the experiment. The data recording portion of the package can be released remotely by acoustic means in the event that the array cannot be successfully withdrawn.

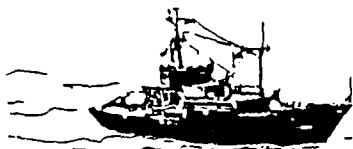
LOWERING 1
AREA SURVEY AND
TRANSPONDER LOCATION



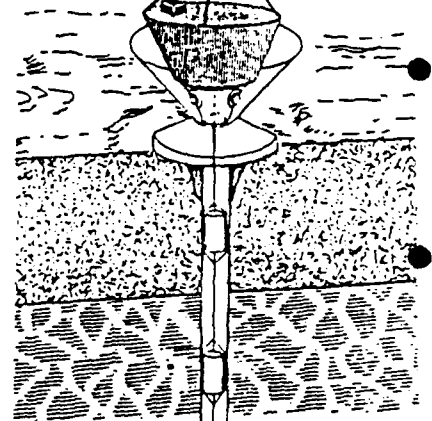
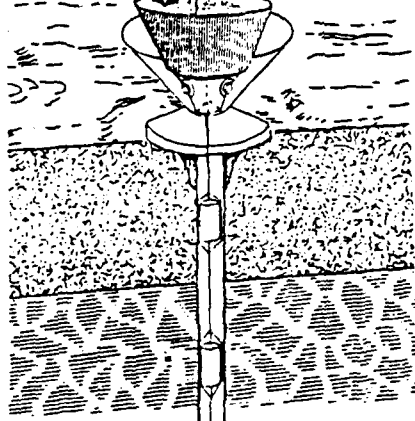
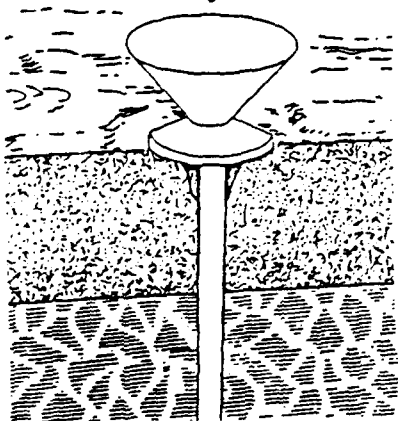
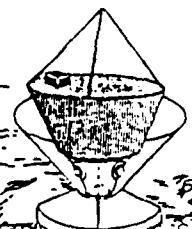
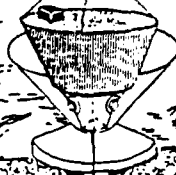
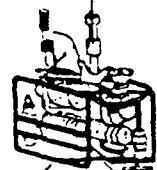
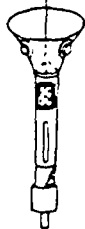
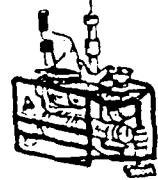
HOLE RECONNAISSANCE



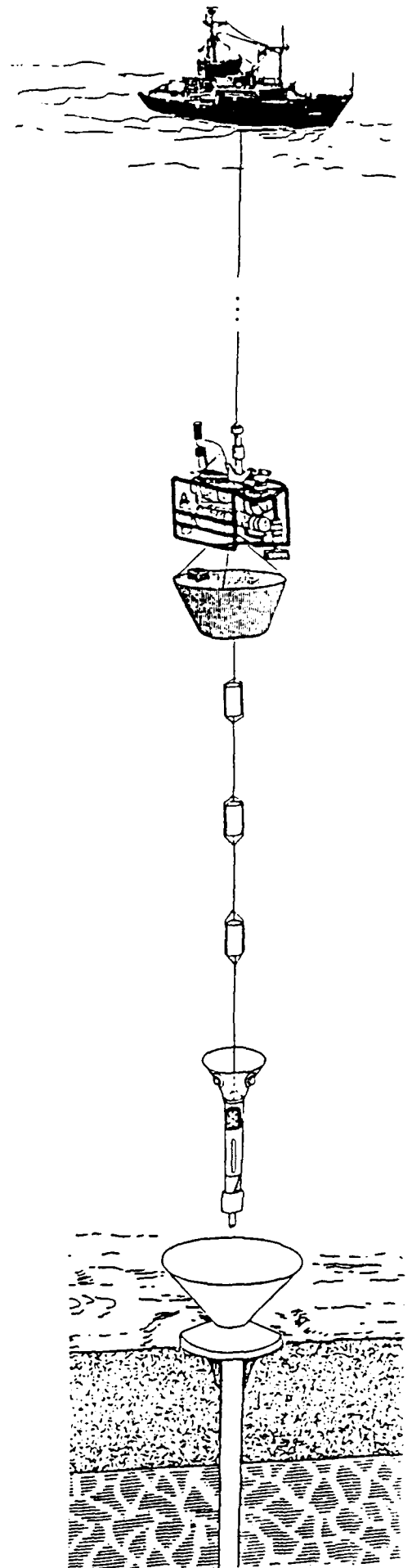
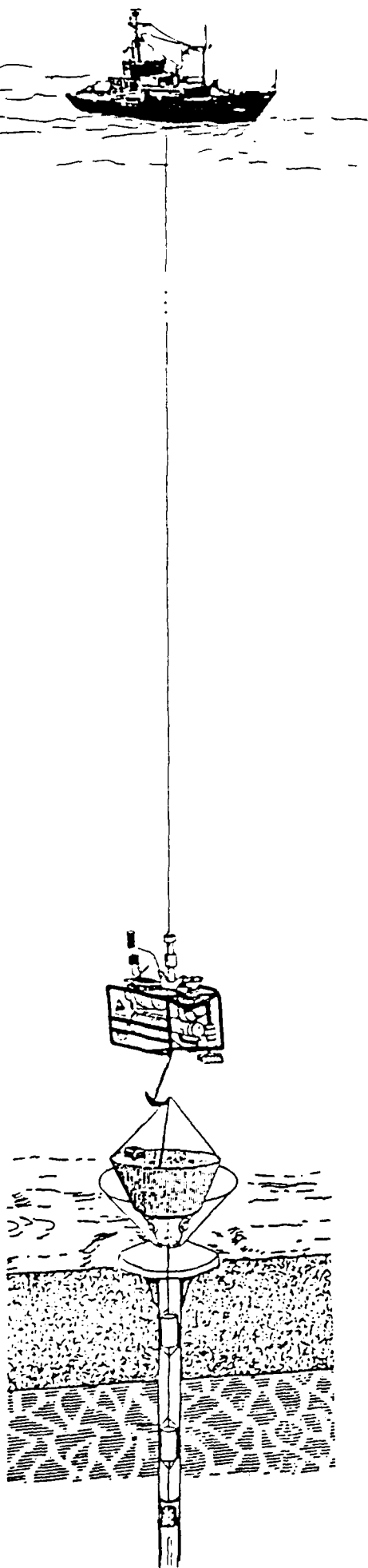




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Coupling

By F. K. Duennebier

Ocean bottom seismometer studies done during the last 20 years, and a few specific tests aimed at the problems of coupling of seismic instruments on the ocean floor have told us a great deal about how not to build ocean bottom seismometers:

don't: have large, massive instrument packages. They tend to be inertial at low frequencies, with resulting poor high frequency response, and they tend to yield resonant data.

don't: have a large cross section in the water, the instrument could be strongly effected by local currents and the shear discontinuity at the ocean floor.

don't: have a very small area in contact with the ocean floor. The effective "spring constant" of the sediment-instrument system increases with area in contact with the bottom; the higher the spring constant, the better the high-frequency response.

don't: have a very large area in contact with the ocean floor. The effective (virtual) mass of the instrument increases as the cube of the radius of the instrument cross section, and the spring constant increases as the square of the radius. A large area also reduces sensitivity to low velocity shear waves, that can have very short wavelengths in sediment.

don't: have the instrument package density much different from the density of the ocean floor. Coupling problems are minimized if the densities are matched.

don't: have large density contrasts in the instrument. The low density parts tend to move faster than the high density parts, creating tilts and torques.

don't have large asymetries. Sensors should be located near the center of the package to minimize effects of rocking and mechanical cross coupling. Asymmetric packages respond to seismic waves differently in different directions.

don't: have antennas and other mechanical moving parts connected to your sensor package. They tend to generate noise.

A solution to these problems is to separate seismic sensors mechanically from OBS recording and recovery systems, placing the sensors in a buried or low-profile package. Although this can cause problems with ease of emplacement and recovery, the potential benefits in signal fidelity are worth the trouble.

Design considerations for experiments interested only in ULF data (1 Hz and below), have not yet been delt with by the community. The increased sensitivity to tilt as lower frequencies are detected requires a more stable platform than is necessary at higher frequencies. It may be desirable to have a relatively massive, wide base area sensor package (with associated loss of fidelity in the VLF band) to obtain more stability against tilt. Burial is an excellent plan for ULF sensors, although emplacement and retrieval complications may not warrent the increase in fidelity. Depth of burial necessary to reduce current-generated noise is probably less than a meter. Burial to greater depths will reduce noise levels, as the sensors are placed in higher velocity material, and

away from the ocean bottom shear wave guide, but S/N from signal sources above the sensors does not appear to change as burial depth increases. S/N from arrivals from below should increase as burial depth increases.

In summary, OBS's should be able to detect the motion of the ocean floor with reasonable fidelity if proper care is taken in the design, and with a bit of luck (an OBS with one corner resting on a stone will probably not yield high fidelity data). Pressure sensors offer the easiest way to observe seismic signals over the total seismic band, as they are not affected by coupling problems. However, particle motion and shear studies require the directional sensitivity of seismometers.

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AN OCEAN BOTTOM SEISMOMETER SYSTEM
FOR THE OFFICE OF NAVAL RESEARCH

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and S.C. SOLOMON⁴

Submitted to the Workshop on ULF/VLF Noise
in the Deep Ocean,
Austin, Texas, Nov. 29-30, 1988.

1. Woods Hole Oceanographic Institution
2. Scripps Institution of Oceanography
3. University of Washington
4. Massachusetts Institute of Technology

Introduction

The Office of Naval Research is currently supporting the development and design of a new ocean bottom seismometer instrument with uniquely powerful and flexible capabilities of data acquisition and recording. This effort is being undertaken by a group of four institutions and universities: Woods Hole Oceanographic Institution (WHOI), Scripps Institution of Oceanography (SIO), University of Washington (UW), and Massachusetts Institute of Technology (MIT). Our goal is to construct an instrument that will satisfy the majority of the data acquisition needs of the seismology and low frequency acoustics communities in the deep ocean for at least the next decade. It follows then that the design is highly modularised. We recognize our inability to confidently predict all the varied future needs and thus by building the instrument of independent units, the modifications and design changes that will be inevitably required in future years will be achievable with minimum impact on the established reliability of the instrument. The design and construction responsibilities are divided among the four co-operating institutions. MIT will design and construct the triaxial seismometer packages and SIO has responsibility for all the data acquisition hardware including the analog amplifiers, timing and primary computer, with help from UW who will build to A to D subsection. WHOI has overall responsibility for the complete project with the specific tasks of building the recording system and configuring the total package.

System Overview

The heart of the new instrument is two 3-foot-long, 7-inch I.D. cylindrical pressure cases: one contains all the analog circuitry and the acquisition computer; the other contains the data recording system and the alkaline batteries that power all the electronics. There are many reasons for such clear separation of these fundamental units. It permits the major portion of the electronics to be operated in 'sealed-case' mode. Unless a component failure occurs there will be no requirement to open the acquisition package pressure case, thus increasing reliability by minimizing human interaction with the hardware. Great flexibility in configuration of the total package is allowed because two cylinders of such modest dimensions can be configured in many ways. They are sufficiently small to be easily handled at sea and ashore. Their complete independence simplifies the process of evolution to, almost inevitably, higher capacity recording systems or perhaps in the longer term, higher dynamic range (24 bit?), lower power consumption acquisition systems. Most importantly it allows reconfiguration of the system into the type of dual package design that we envision would be necessary for permanently installed monitoring stations. In this scenario the acquisition system would sit permanently on the ocean floor beside some sensor array and be linked to the recording and power system by several kilometers of conducting wire. This latter package could then be recovered and repowered at periodic intervals without the need to disturb the permanently installed (buried and downhole?) sensor system.

Fig. 1 shows a block diagram of the complete system as currently planned, and Fig. 2 illustrates one possible configuration as a single unit OBS with an external deployable sensor package. The intention is that the instrument be operable with a wide range of sensors: six channels of hydrophone and long-period pressure sensor, two 3-component seismometer packages, and for the vast majority of experiments are three component seismometers with a Cox-Webb long-period pressure sensor and perhaps a conventional pressure compensated hydrophone. As shown in Figure 2, buoyancy will be supplied by conventional glass balls because of their low cost, proven reliability and flexibility in configuration. Recovery will be achieved using two completely independent commercial acoustic releases mounted in tandem.

Acquisition Package

A block diagram of the acquisition system that shows more detail than that in Fig. 1 is shown in Fig. 3. It can handle up to six channels of primary data at sampling rates between one sample every 8 seconds up to 256 samples per second per channel. A 16 bit A to D converter and both software-controlled and fast auto gain ranging result in a dynamic range capability in recording in excess of 120dB. The acquisition tasks are controlled by an Onset 80C88 computer on a C44 bus. The clock is designed and manufactured by the Webb Research Corp. and provides a timing accuracy of a few milliseconds over one year. Data is transferred to the recording package over a serial data link at a rate of 38.4 kilobaud. System status (particularly with regard to sensor disposition) is transmitted for a fixed period of time following deployment via a simple code using one of the acoustic release transponders. This transponder is activated by a small transducer mounted internally in the acquisition pressure case which in turn is controlled by the acquisition computer.

Two additional auxiliary data channels are provided for low data rate information (e.g., current, sensor azimuth, etc.). Provision is also made to record any one of the primary data channels (at any one time) through a 500 Hz baud pass and envelope detail circuit. This is invaluable for the precise recognition of water borne events that is critical to the accurate determination of instrument location on the seafloor.

Recording Package

In the recording package (Fig. 4) a second Onset 80C88 computer receives the data from the acquisition system down the serial data link and controls its temporary storage in 4 Megabytes of RAM and its periodic removal to optical disc. By standardizing on a SCSI interface for this recorder we have reduced the impact that the choice of a particular model will have on the system design. Given the volatile state of this market, this is an important advantage that will hopefully enable us to upgrade as the state-of-the-art advances in the coming years. Our tentative choice of drive for the prototypes is the Maxtor 400 MB unit, but this choice will be reviewed before production commences.

In addition to the recording system, this three-foot-long pressure case contains sufficient alkali batteries to permit two months of continuous operation of the electronics. Power for the optical disc drive is provided by an external lead-acid battery.

Sensor

For the two prototype units under construction at the time of writing sensors have been developed based on the successful MIT deployable package using three component 2Hz seismometers. However, plans now exist to use a set of 1Hz seismometers mounted in a 12" O.D. aluminium sphere. Although the cost of such a system is relatively high, it combines the critically important advantages of the good coupling characteristics in the VLF band resulting from its small size with useful sensitivity down to frequencies perhaps as low as 0.05Hz. A prototype deployment scheme for this package is illustrated in Fig. 1. This scheme retains the relative orientation of the sensor package and the main instrument frame, thus negating the need to add the bulk of an azimuth sensor to the external package itself. Studies of simple and adequately accurate methods for azimuth measurement continue.

Too little is known of possible interactions between the main instrument package and the deployed sensor to allow optimization of the main instrument anchor design or sensor - main package separation. Because of the ease with which our planned package can be

reconfigured, it will be relatively straightforward to respond to this information as our learning progresses in the coming years.

System Operation

Any instrument with the capabilities described here - six channel recording at 256Hz per channel, 120dB dynamic range, the option for one year deployments and hundreds of megabytes of data storage - is necessarily a complex piece of hardware. In order to achieve reliability at sea, however, the operation and check out of the instrument must be rigorous and simple. For the acquisition and electronics packages this can be achieved with the development of good software controlled monitoring, predeployment checkout and task definition procedures. Before deployment all instruments will be connected as a serial data loop that a 386 'mother' computer will use to routinely check correct functioning and status of, e.g., recording medium, clock, A to D converters, memory, power supply voltages, etc. The software in the 386 will detect failures or deviations from the norm and alert the operator. If desired these check-out procedures can proceed with the instruments on the fantail and only be stopped immediately before deployment. The same data loop will permit changes to the data recording tasks to be made right up to deployment time. If last minute failure of a recording or acquisition system occurs this would be overcome by replacement of the complete package. Well before commencement of deployments, and before acoustic releases were attached to the instrument frames their correct operation would be established in the most thorough manner possible by simply clamping them to a hydro wire and operating them at some appropriate depth. The same 386 computer used for instrument check-out will be used for data quality assessment upon instrument recovery and thus will be outfitted with an optical disc drive identical to that in the recording package. The instrument programming is carried out using Aztec 'C' and the 386 will run the MS-DOS system.

Status Report

Construction of two prototype units is underway. SIO and UW are building two acquisition packages, WHOI is building two recording packages and instrument frames, and MIT is constructing two 2Hz deployable seismometer packages. We plan to bring the first complete system together in early calendar 1989 and carry out shallow water tests in spring 1989. Manufacturing will then commence immediately. Within present budgetary constraints we plan to build 30-35 complete instruments by the summer of 1990. Each of the four cooperating institutions is responsible for the manufacture of the systems that they have designed. Current plans call for these initial 30-35 instruments to be fitted with a three-component 1Hz external seismometer package, and a Cox-Webb long-period pressure sensor. Then ownership will remain with ONR, who will also determine by whom within the U.S. academic community they will be operated.

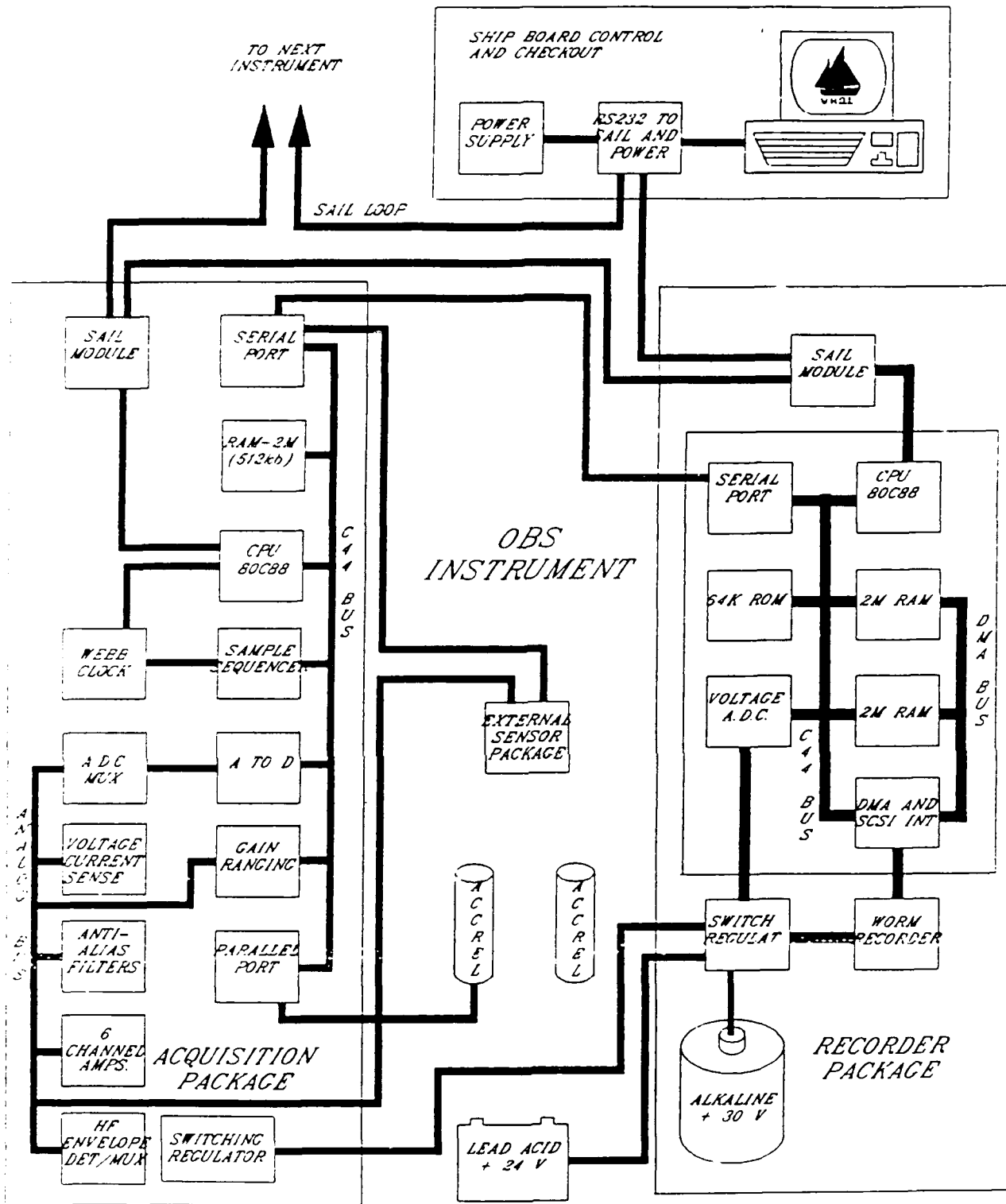
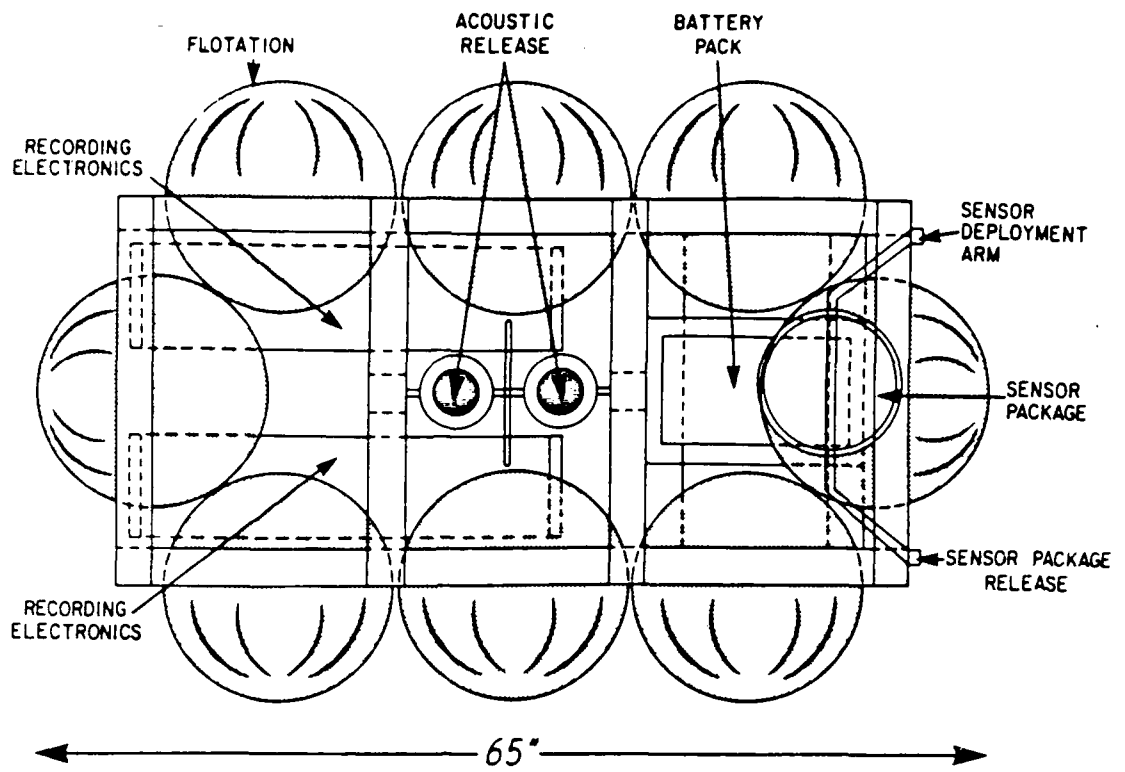


Figure 1



PLAN VIEW

THE NEW Q.N.R. O.B.S.

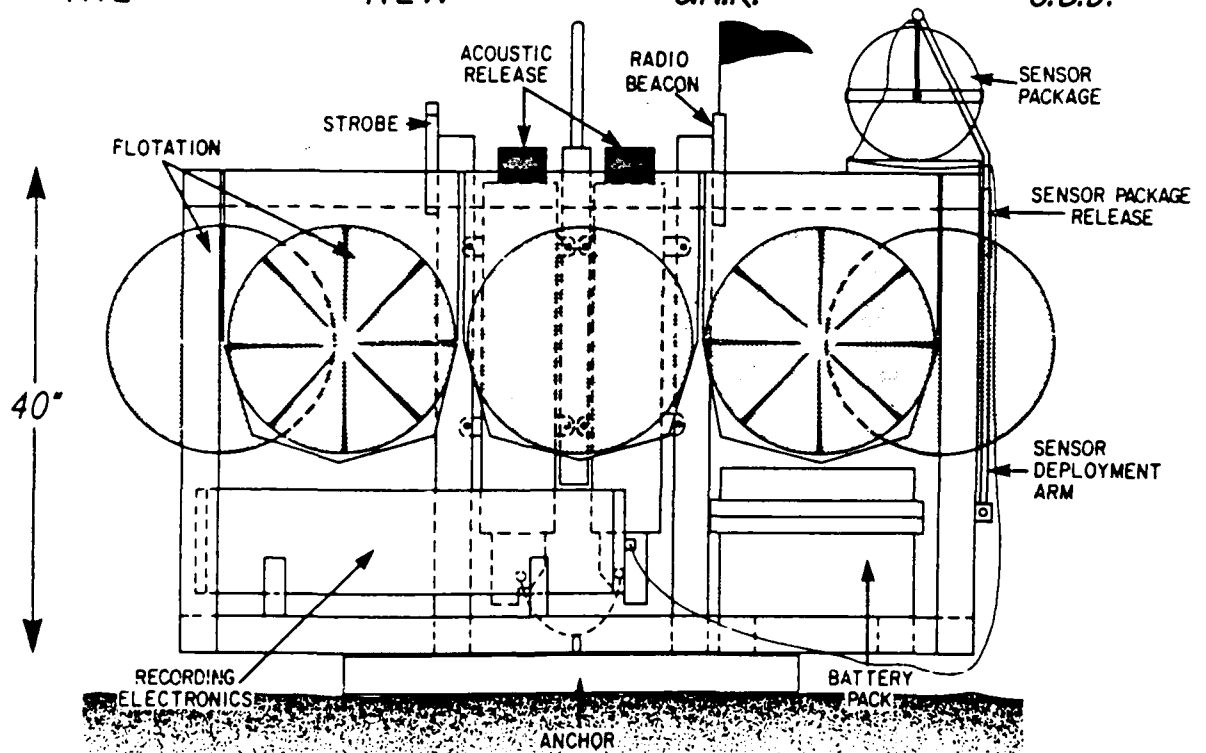
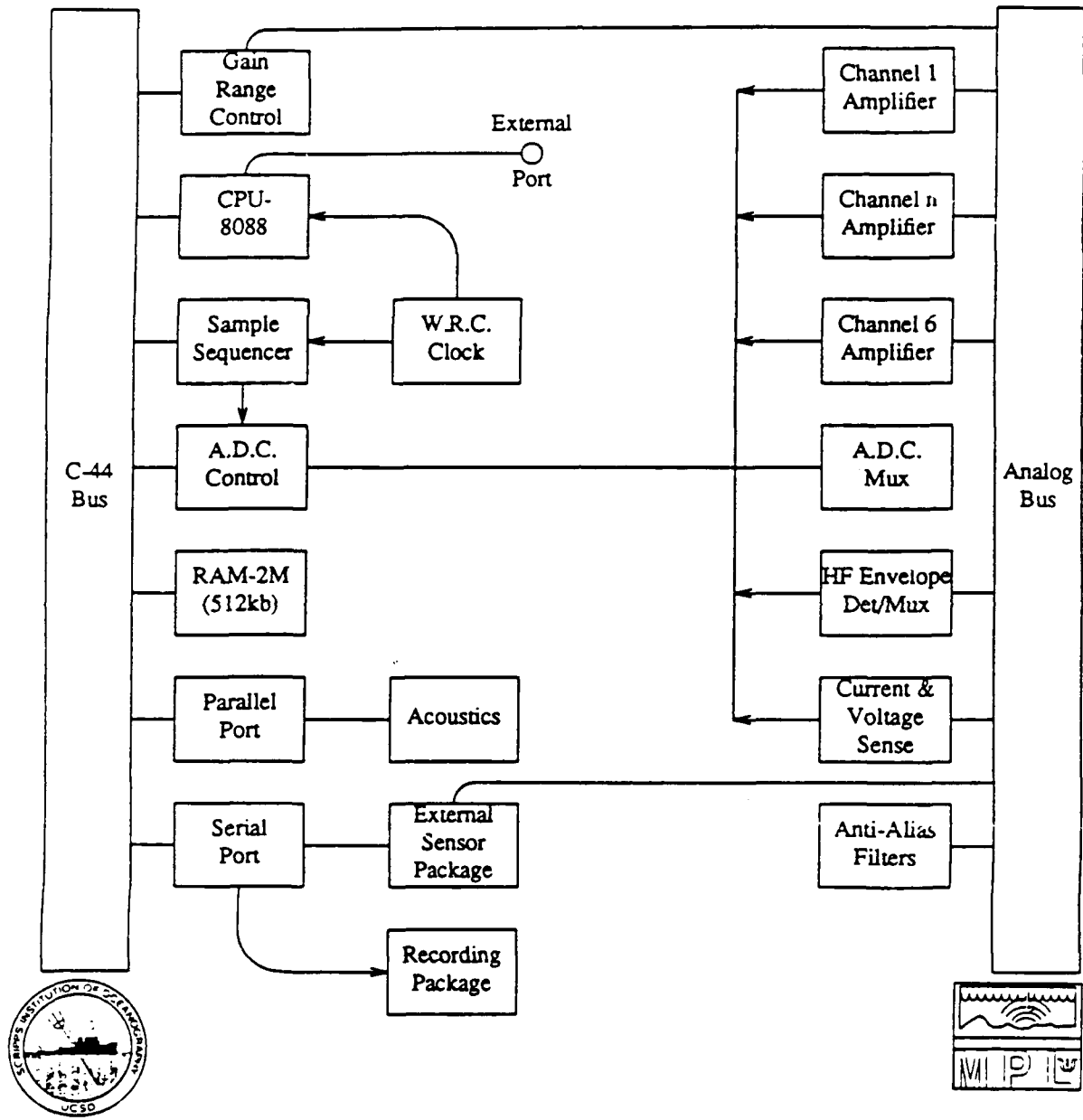


Figure 2



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Figure 3

OBS RECORDER

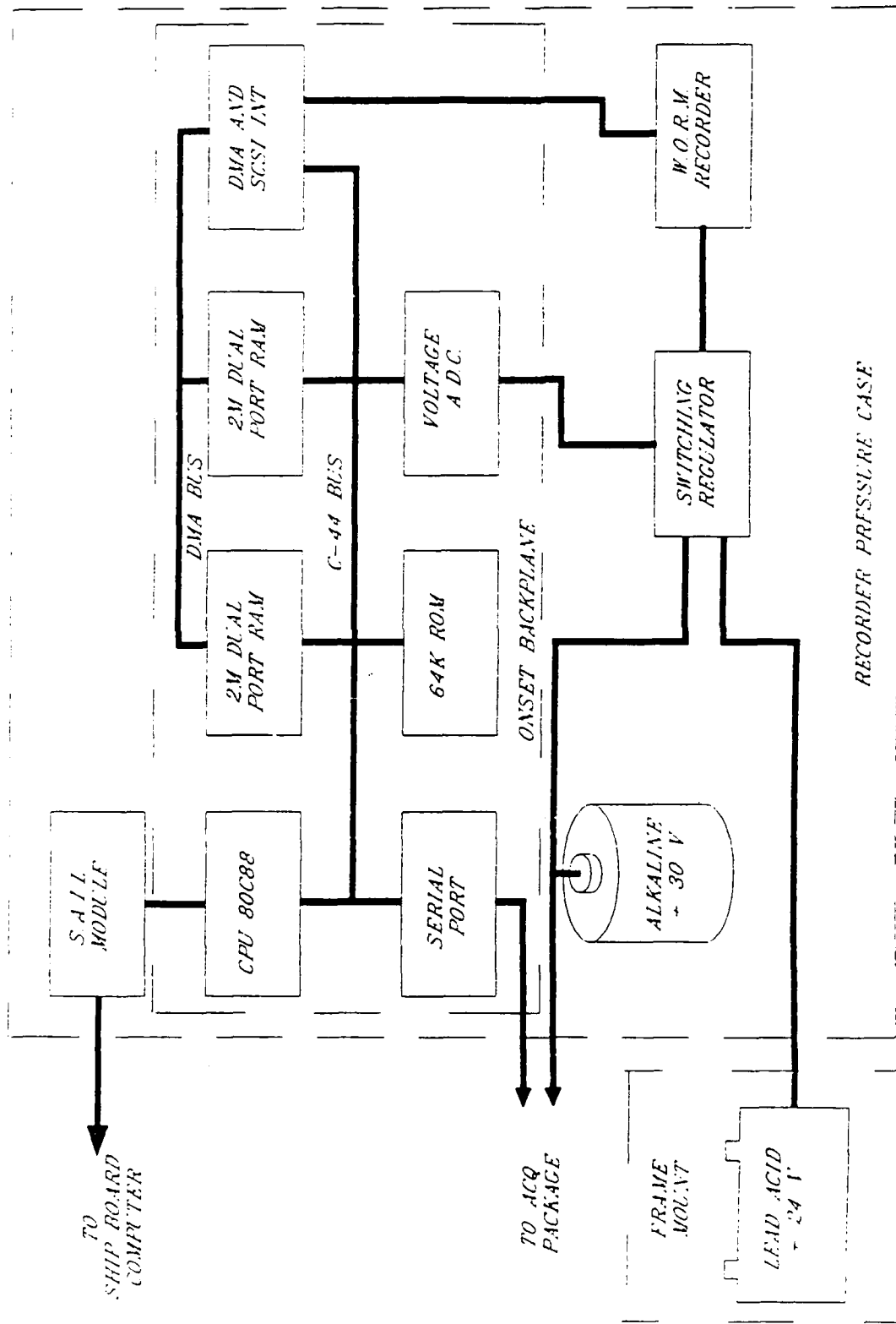


Figure 4

THE SENSOR PACKAGE FOR THE ONR VLF OBS

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Introduction

As part of the ONR Research Initiative on ULF/VLF ambient noise, MIT has been building prototypes of the external sensor package for the VLF ocean bottom seismometer (OBS). The design philosophy has been to utilize the heritage of the present MIT external geophone package in terms of small size, low profile, and sensor configuration. This arrangement has generally yielded good coupling to the seafloor and an absence of spurious resonances in the working frequency band [*Duschenes et al.*, 1981; *Trehu and Solomon*, 1981].

The sensor-package development effort has been divided into two phases. During fiscal year (FY) 1988 we designed and constructed two prototype packages with three-component 2-Hz seismometers. These packages are in the final stages of bench and vault tests and will be utilized in the first wet tests of the VLF OBS early next year. During FY 1989 we are designing and constructing a three-component 1-Hz seismometer package. A preliminary design concept for this 1-Hz package has been completed and is described below. We anticipate that the 1-Hz package, following initial tests, will be the principal seismometer system for the production phase of the VLF OBS.

The 2-Hz Sensor Package

The 2-Hz sensor package includes three matched seismometers (2 horizontal and 1 vertical) configured orthogonally. The geophones, support hardware, and electronics are housed in a cylindrical pressure case that is 30.5 cm in height and 18 cm in diameter. The pressure case sits vertically atop a circular base plate 40.5 cm in diameter. The package weighs 16.3 kg in air and 7.7 kg in water. The cable to the main instrument package is

detachable. The cable connector is located at the top of the package, on the central vertical axis, to preserve symmetry of mechanical response. Prior to deployment the package will be held to the deployment arm of the main OBS package by means of a handle attached to the top of the package.

Sensors. The seismometers are Mark Products model L-22E units. These devices are internally damped to 70% of critical, thus requiring no external shunt resistance. The coil resistance was chosen to be 8540 Ω , which provides a transduction constant of 10.97 V/cm/s.

Sensor Mounting and Gimbaling. The three sensors are mounted in a vertical array, gimballed such that it is self leveling up to 15° off vertical in any direction. The gimbal assembly is immersed in a highly viscous oil to damp relative motion between sensors and package in the instrument passband.

Electronics. The package electronics are housed in an oil-tight container mounted directly beneath the top end cap. Each of the three seismometer signals is routed through a variable-gain preamplifier stage. The gain is digitally controlled over a range of 54 db in nine 6-db steps. Gain control instructions are received from the main instrument package via a serial data link. The amplified signals are then sent to the main package.

An experiment is now being conducted at the seismic vault of the MIT Wallace Geophysical Observatory in Westford, Mass., to determine whether rigid clamping of the sensors is required for proper instrument response at the lowest frequencies of the VLF band. The two completed 2-Hz sensor packages are placed side by side on an instrument pier, one with its sensor array rigidly coupled to the case, the other with its gimbal assembly free to move within the highly viscous damping medium. The outputs of the two sets of sensors are routed through the actual VLF OBS acquisition electronics and digitally recorded simultaneously. Data will also be recorded with no sensors connected (equivalent source impedances installed) in order to control for variations in the electronics and recorder. Noise and signal spectra will be compared to assess differences in the responses of the two sensor configurations.

The 1-Hz Sensor Package

We have arrived at a preliminary design concept for the 1-Hz sensor package. The sensors, mounting hardware, and electronics are to be housed within a spherical pressure case. The pressure case will consist of two hemispheres, each with an outer diameter of 30.5 cm and a wall thickness of 1.6 cm, separated by an equatorial ring (33 cm O.D., 2.9 cm wall thickness, 5 cm height). The hemispheres and ring will be fabricated out of 7075-T6 aluminum. The total weight of this package will be approximately 30.4 kg in air and about 10.7 kg in water. The low weight of the package compared to designs incorporating a cylindrical pressure case has been a major factor in selecting a spherical geometry.

Three 1-Hz sensors (2 horizontals and 1 vertical) will be mounted in a gimbal that will be free to move 30° off vertical in any direction. The sensors will be Mark Product's L4-C seismometers that will be installed in a orthogonal mounting manufactured by Mark Products. As with the 2-Hz unit, the gimbal assembly will be suspended in a highly viscous oil.

The electronics for this package will be the same as for the 2-Hz package, with the addition of an *in situ* calibration capability controlled from the main package via the serial data link.

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Electric Field and Pressure Gradient Measurements

S.C. Webb and C.S. Cox, Scripps Inst. Oceanography

During the last several years, we have pursued the development of both electric field and pressure gradient measurements as alternatives to horizontal component seismometers for deep ocean seismic measurements. This work is partly motivated by the problems associated with the coupling of horizontal components of OBS's to soft sediments, but also by the interesting and unique properties of these new sensors. The coupling problem is complicated and depends on the detailed geometry of the OBS and the elastic moduli of the surface sediments. The two types of sensors described here are stretched horizontally across many meters of seafloor and generate measurements averaged over the entire length of the sensor. Presumably because the piping associated with the pressure gradient gauges and the cabling of the electric field antennas lie directly on the seafloor these sensors more accurately follow seafloor motions than a seismometer in a pressure case above the seafloor.

The electric field sensors are fully described in other publications. A voltage is induced along an antenna that moves with the seafloor through the geomagnetic field. At frequencies below one hertz a significant electric field is also induced within the seawater by the motion of the seawater through the geomagnetic field. The net result is that below one hertz the electric field detected by the antennas is $F \times (u_S - u_W)$ where F is the vector geomagnetic field and $(u_S - u_W)$ is the difference in the velocity of the antenna and the velocity within the water just above the seafloor.

The pressure gradient gauge might be more properly described as a pressure difference gauge. A long copper pipe is used to connect two points on the seafloor. A differential pressure gauge at one end or in the middle of the pipe is used to block the pipe. The pressure difference from one end of the pipe to the other is detected by the sensor allowing for the propagation delay down the pipe. A sensor at the center of the pipe sees the same propagation delay from both ends of the pipe and correctly measures the pressure gradient along the pipe. A pressure sensor at one end of a pipe can detect an

acoustic wave propagating perpendicular to the pipe, but will be insensitive to waves propagating along the pipe toward the sensor because of the propagation delay. The pressure gradient gauges exhibit antenna-like behavior because of this dependence on propagation delays. An apparent pressure gradient is also generated by horizontal accelerations of the pipe. This problem has not yet been fully investigated.

The pressure gradient associated with a plane wave is related to the pressure in the wave simply as $\mathbf{k}p$, where \mathbf{k} is the vector wavenumber and p is the pressure signal. The phase velocities associated with low frequency seismic noise can be derived from a combination of the pressure gradient and pressure measurements if one assumes the directional spectrum is isotropic and that only one mode is present. This calculation performed on some recent data from the NACHOS expedition demonstrates that reasonable values for phase velocity in both the surface gravity wave band (200 m/s) and in the microseism peak (.1-.3 Hz) can be obtained using these arguments. The spectrum of ambient pressure noise is probably not as simple as this model, but the measurements do suggest useful measurements can be obtained from pressure gradient devices.

The pressure gradient and electric field sensors have been deployed during the last year as part of two large seafloor arrays called NACHOS (May 1988) and Pegasus (Nov 1988). Measurements from these sensors will be compared with predictions from wavenumber-frequency spectral estimates derived from the pressure and seismic sensors in the arrays.

Five sets of single component electric field and pressure gradient sensors were deployed during last month's Pegasus experiment. These measurements will be particularly useful because four of these sets were deployed as orthogonal pairs. The electric field sensors detect motions perpendicular to the direction of the antenna, so signals on the electric field antenna on one instrument should correlate with the pressure gradient measured by the instrument laid out orthogonally.

ESTIMATION OF SURFACE NOISE SOURCE LEVEL FROM LOW-FREQUENCY SEISMO-ACOUSTIC MEASUREMENTS

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1 Summary

The spatial structure of low frequency noise from distributed sources such as those causing surface generated noise in deep or shallow water is governed by the waveguide defined by the water column bounded above by a pressure release surface and below by a stratified viscoelastic medium. Therefore, in determining the actual source spectrum level of the noise generating sources from measured data, it is necessary to account for or "subtract out" the ocean waveguide environment. Only after this procedure is followed, can source levels be derived from noise measurements performed in arbitrary ocean environments and consequently, be compared between experiments performed in different environments and seasons.

In this paper we apply a wave theory of surface distributed noise in a stratified ocean to data collected in a low frequency shallow water experiment. The results explain the waterborne and seismic partitioning of low frequency distributed noise in shallow water. This same analysis applies to the wavelength-scaled deep water distribution of surface generated noise. We find by accounting for the ocean environment in a set of diverse experiments previously summarized by Kibblewhite et al [1], that the spread of the reported noise source levels as a function of frequency and parameterized by wind/sea state is considerably reduced over the levels originally reported which took the noise measurements as the noise source levels. Not

only does this demonstrate the importance of including propagation factors in processing noise data for estimating source levels, but the results indicate a consistency in the frequency dependence of the noise source levels which suggests that there are only one (or possibly two) dominant natural noise source mechanisms that contribute to ambient noise below 10 Hz.

At intermediate frequencies and above, the acoustic waveguide nature of the noise field has previously been reported [2,3,4]. Recently, Ingenito and Wolf [5] has studied the site dependence of shallow water noise data in terms of waveguide theory [2]. At lower frequencies, the stratified ocean environment supports not only body waves but also surface waves associated with the interfaces of the layers. Whether or not discrete modes exist in the environment, there exist interface waves in a viscoelastic environment which are never cutoff [6]. The amplitude distribution of a surface wave decays exponentially away from the guiding interface, which implies that interface waves can only be excited by sources close (in terms of wavelength) to the interface, a condition normally satisfied for ocean bottom interfaces at acoustic frequencies below cutoff. Hence, at frequencies below waterborne propagation cutoff, these interface waves provide a mechanism for sound, including ambient noise, to be propagated and sensed in the water column and on the sea bed in particular. This partitioning between body waves and interface waves provides an explanation for the spectral distribution of noise observed experimentally in shallow water [1,7] and also reported elsewhere for deep water environments [8,9,10] where it is believed that some sort of activity at the air/sea interface is the source of low frequency noise often referred to as microseisms [11].

The Kuperman-Ingenito theory describing the distribution of surface generated noise has been modified to include expressions for the vector quantities which a seismic sensor with three geophones would measure. It is shown that this theory predicts with decreasing frequency below cutoff a large increase in the vertical and radial components of the outputs of the geophone as compared to a much smaller increase from the pressure output of an adjacent hydrophone. Above cutoff, the noise spectrum level oscillates in amplitude as function of frequency corresponding to the cutoffs of the individual modes as seen experimentally, for example in deep water data [12].

We describe a shallow water experiment with an ocean bottom seismometer (OBS) and present its results which are consistent with the theoretical predictions. Since the theory predicts only relative levels because the spectral level of the noise sources is an unknown input, the experiment

combined with theory can be used to derive the spectral distribution of the noise sources. Then the same procedure is applied to other shallow water data as summarized in Ref. [1], and finally the water depth dependence of the spectral distribution is investigated.

It is demonstrated that the excitation of seismic waves below waterborne mode cutoff accounts for a significant frequency dependent "magnification" in observed noise levels as compared to the actual source levels. Accounting for the propagation in the spectral distribution of noise is therefore crucial at low frequencies, resulting in a significantly lower spread in the derived source levels as opposed to the spread of the directly measured noise levels. Further, this procedure will separate propagation from sea state effects.

It should be stressed that the present paper is not intended as a contribution to the ongoing discussion concerning the nature of the source mechanisms [13], but rather as a theoretically derived statement based on existing data of the importance of properly accounting for propagation effects when evaluating different theories by comparison to experimental data.

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Finite difference modeling of scattering

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We present finite difference forward models of elastic wave propagation through laterally heterogeneous upper oceanic crust. The finite difference formulation is a 2-D solution to the elastic wave equation for heterogeneous media and implicitly calculates P and SV propagation, compressional to shear conversion, interference effects and interface phenomena. Random velocity perturbations with Gaussian and self-similar autocorrelation functions and different correlation lengths are presented which show different characteristics of secondary scattering. Heterogeneities scatter primary energy into secondary body waves and secondary Stoneley waves along the water-solid interface. The presence of a water-solid interface in the models allows for the existence of secondary Stoneley waves which account for most of the seafloor 'signal generated noise' seen in the synthetic seismograms for the laterally heterogeneous models.

'Random' incoherent secondary scattering generally increases as ka (wavenumber, k , and correlation length, a) approaches one. Deterministic secondary scattering from larger heterogeneities is the dominant effect in the models as ka increases above one. The strength of scattering and efficiency of coupling into interface waves is affected by the size and strength of the velocity variations as well as the depth profile of Poisson's ratio. As the shear wave velocity near the seafloor decreases down to 0.8 km/sec, coupling due to scattering into shear body waves and Stoneley waves decreases dramatically.

The utility of the finite difference method for investigating seafloor noise propagation has been unquestionably established by the results seen in our pulse scattering models. We would like to apply these results to further investigations of noise propagation in the following areas;

1. Distributed source fields. While a pulse response can provide improved insight into scattering mechanisms, the actual real world source, or 'signal' which generates ULF/VLF noise along the bottom is obviously much more complex. Source wavefields proposed from experimental results can be tested for their response upon interaction with various heterogeneous bottoms.

2. Shear velocity structure and/or sediments. The code currently in use can be used for a wide range of velocity structure and/or Poisson's ratio. How does Poisson's ratio or very low shear velocity affect the generation and propagation of scattered 'noise'?

3. Data analysis. Can these effects be seen in existing controlled source data? If not, how can we design an experiment to look closer at 'signal' generated 'noise'?

THREE-DIMENSIONAL NOISE FIELDS IN COMPLEX OCEAN ENVIRONMENTS

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1 Summary

The structure of surface generated noise in the ocean depends on both the noise source mechanisms and the characteristics of sound propagation in large regions of complex ocean. The latter environment includes three-dimensional variability in the ocean and bottom structure. With respect to the ocean bottom, low frequency acoustic noise invariably involves seismic propagation and hence the coupling and interaction of waterborne acoustic waves with the compressional, shear and interface waves associated with visco-elastic media. The modeling requirements for this problem, therefore, involves the computation of acoustic and seismic fields from all positions on the ocean surface to arbitrary positions in space. Furthermore, the interesting features of the noise are described by its cross-spectral density, which is a non-homogeneous quadratic form of the field structure.

A previously developed noise theory [Kuperman and Ingenito, JASA 67, 1980] (KI) was modified and combined with a recent 3-D modal method (WRAP) [Kuperman, Porter, Perkins and Piacsek, 12-th IMACS, Paris, 1988]. The wave equation solution of the acoustic field (or its cross-spectral density) from any point to any point (or to any pair of points) reduces to a "spreadsheet" type manipulation of precomputed modal solutions. With the surface represented by a statistical distribution of sources, any order pole radiator can be constructed according to the specific noise mechanism. Furthermore, local disturbances and storms can be included in this 3-D environment by changing the source strength and character of the appropriate patches of the surface [Perkins, Pasewark, and Kuperman, JASA Suppl. 1, 85,1989]. The KI model was also recently extended to include elastic media effects [Schmidt and Kuperman, JASA 84,1988] for studying the partitioning of noise field into waterborne and seismic paths in a stratified medium.

The next step should be the combination of the noise theory with elastic media in a three-dimensional environment. The development of such a model is now possible because WRAP uses a complex eigenvalue solver applicable to elastic media, "KRAKEN" [Porter and Reiss, JASA, 77,1985]. This model would address the distribution of noise in the frequency regime of about .1- 50 Hz originating from an uneven spatial/surface distribution of sources in a 3-D range-dependent environment. Of particular interest would be behaviour of the noise fields in regions of complex bottom topography and structure. Problems such as the distribution of noise from storms far from or near to continental margins could then be addressed.

**Open Ocean Surface Wave Directional Spectra
from Doppler Acoustic Measurements
(Flip-Based Measurements)**

By Jerry Smith, MPL/SIO

Two components of the ULF sound can be forced by local surface wave interactions: (1) the "double frequency" spectrum (1 to 8 sec period), due to oppositely-directed surface wave components, and (2) bound waves generated by wave groups (of order 100 second periods and longer). Non local forcing is also possible - e.g., sound can propagate horizontally from strong sources such as storms or regions with a reflected swell component, and long period free waves (100 sec) can propagate from coasts. Thus, it is of interest to investigate what portions are locally forced, and under what conditions.

Two models of the typical wind-wave directional distribution are shown in Figure 1: (1) a $\cos^2 \phi$ distribution (dots), which would yield zero total opposing-component energy, and (2) a $\cos^8(\phi/2)$ distribution (solid), which has enough energy in opposing components to roughly match the observed levels of "double frequency" microseisms (note especially the non-zero energies of waves perpendicular to the wind). The similarity of these two curves suggests that resolving the opposing components is a serious challenge: the energy of one or the other "opposing component" is down by at least an order of magnitude from that of the directional peak. This general form for the directional distribution of wave energy suggests that the opposing component mechanism should be greatly enhanced immediately following a change in wind direction. Then new, growing components can oppose older, "left over" components. Indeed, source observations of the "double-frequency microseisms" support this (S. Webb, personal communication).

A system capable of resolving opposing components under "normal" conditions could measure the effects of wind veering as well. Also, it would be more than adequate to resolve wave groups, which generate a second order (in wave slope) variation in periods on the sea floor (see Chip's section).

The Doppler acoustic systems presently being developed by Rob Pinkel and myself for use from R/P FLIP may plausibly provide this level of resolution. These systems consist of several sonar beams each, which are broad in the vertical plane but narrow azimuthally, and are aimed so as to "scan" along the underside of the sea surface in different directions (see Figure 2). One is a 75 kHz (center band) system with 1.5 km range and 12m range resolution; the other is a 200 kHz system with 500m range and 3m resolution. These range resolutions are balanced against the velocity measurement error by requiring sufficient velocity resolution to detect a wave 100 times weaker than the typical "equilibrium" level of wave components parallel to the wind, at the cutoff (or Nyquist) wavelength (see Smith, 1988; submitted to JAOT).

A preliminary deployment of both systems took place in May 1988, with two beams each (see Figure 3). Another deployment of a 3-beam, 200 kHz system took place in November 1988. Analysis techniques are currently being developed for these and future data. The results should clarify the capabilities of these systems.

It is useful to view the response of a single sonar beam to surface waves, on the k plane. At a given frequency, there is a corresponding surface wavenumber ($k = \sigma^2/g$), so a frequency band translates to a wavenumber annulus in 2 dimensions (see Figure 4). A wavenumber-component band in the along-beam direction (k_x , say) can intersect this annulus in 1 or 2 "patches" (as illustrated in Figure 4). Thus, the directional response is widest for waves traveling nearly parallel to the beam. On the other hand, the Doppler shift yields only the velocity component parallel to the beam. Waves traveling perpendicular to the beam have orbital velocities almost completely orthogonal to the beam (except for a small vertical inclination of the beam), and are virtually "invisible" (see Figures 5 and 6). Thus, the best directional resolution of each beam occurs for obliquely incident wave components. For this reason, three beams can be expected to perform much better than two. Of course, additional beams provide additional degrees of freedom, so that the ability to resolve low-energy components in the presence of other high energy ones can be augmented by adding beams.

In summary, the FLIP-based sonar measurements have the following characteristics:

- More beams yield more (statistical) accuracy, but with decreasing gain vs. effort.
- 2° directional resolution is feasible with 3 or more beams
- Surface waves with periods from 1 1/2 to 20 seconds or so can be resolved with sufficient precision to detect a wave 2 orders of magnitude less energetic than the corresponding directional maximum.
- Most costs (e.g., development, construction) are already covered under the surface wave ARI. Salary, deployment, and analysis costs are needed (!).
- Need remains to evaluate whether the small-energy component can indeed be measured in the presence of high-energy ones.
- Finally, in addition to the sonars, FLIP can carry anemometers, etc. to monitor the local meteorology.

Figure Captions

Figure 1. Two models of the "typical" wind-wave directional distribution: (dots) a $\cos^2 \phi$ spread, which yields no energy in opposing components, and (solid line) a $\cos^8(\phi/2)$ spread, which has enough opposing energy to roughly match observations of "double frequency" microseism.

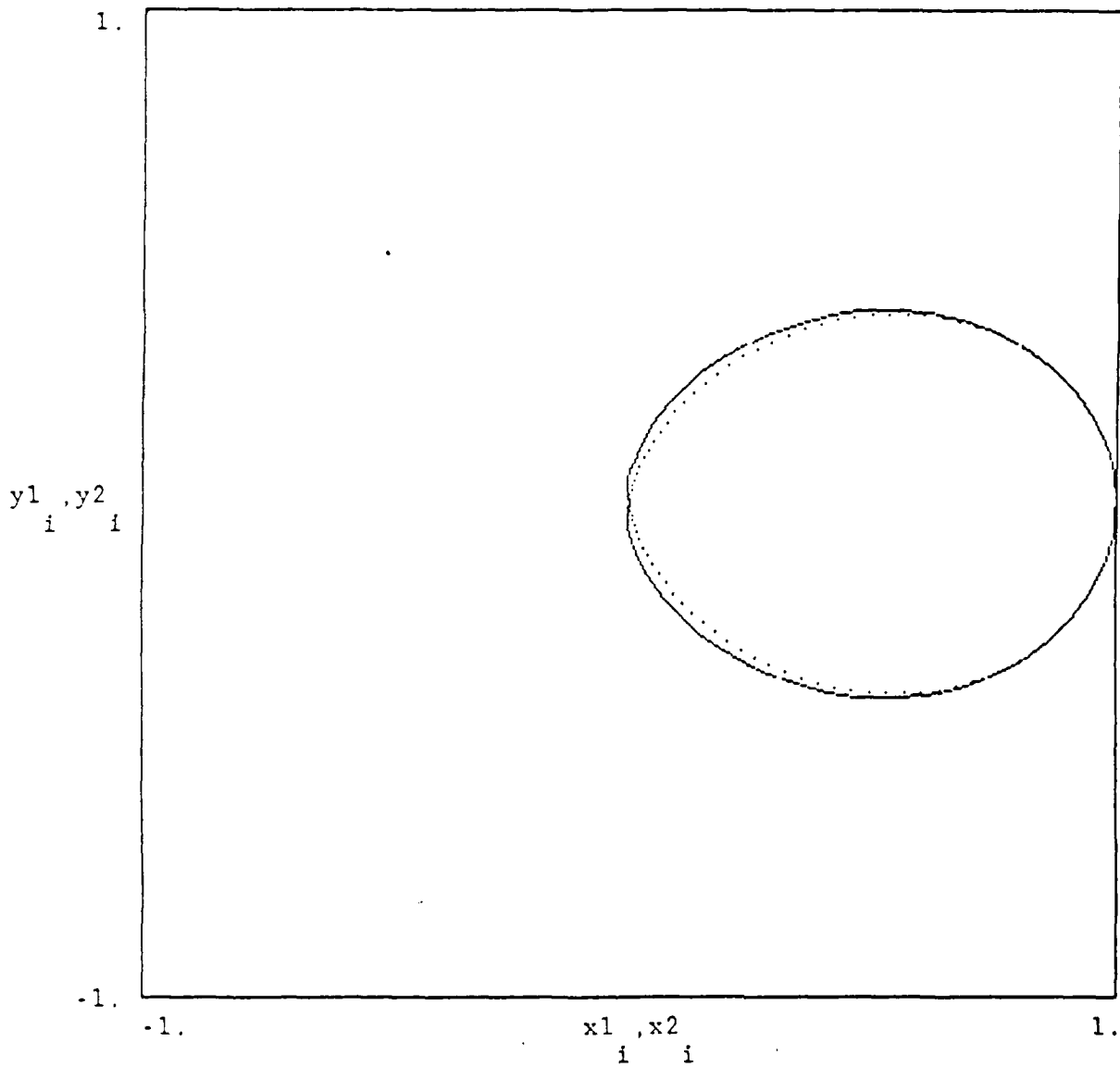
Figure 2. Schematic view of a surface-scanning sonar beam. The vertical measurement extent is limited by the surface-trapped nature of the scatterers (bubbles). The resulting measurements are stable with respect to tilting of FLIP by the waves.

Figure 3. Plan view of the two-beam sonars as deployed in May 1988. The 75 kHz and 200 kHz beams were parallel to permit comparisons.

Figure 4. The intersection of a frequency-band and a wavenumber-component band as arises from a single sonar beam. Waves traveling perpendicular to the beam (along the k_y axis) would be nearly "invisible", since they have no orbital motion parallel to the beam.

Figure 5. An example of a single-beam frequency-wavenumber spectrum for a beam nearly parallel to the wind. Solid lines show the linear dispersion relation. The peaks lie along this line as nearly as possible.

Figure 6. Another single-beam frequency-wavenumber spectrum, for a crosswind beam. The peaks lie inside the dispersion lines, as is appropriate for waves crossing the beam obliquely.



polar-map of directional energy distribution for
 $\cos^2 \varphi$ (dots) vs $\cos^2 (\varphi/2)$ (solid)

Figure 1.

Scanning Acoustic Linear Surface Array

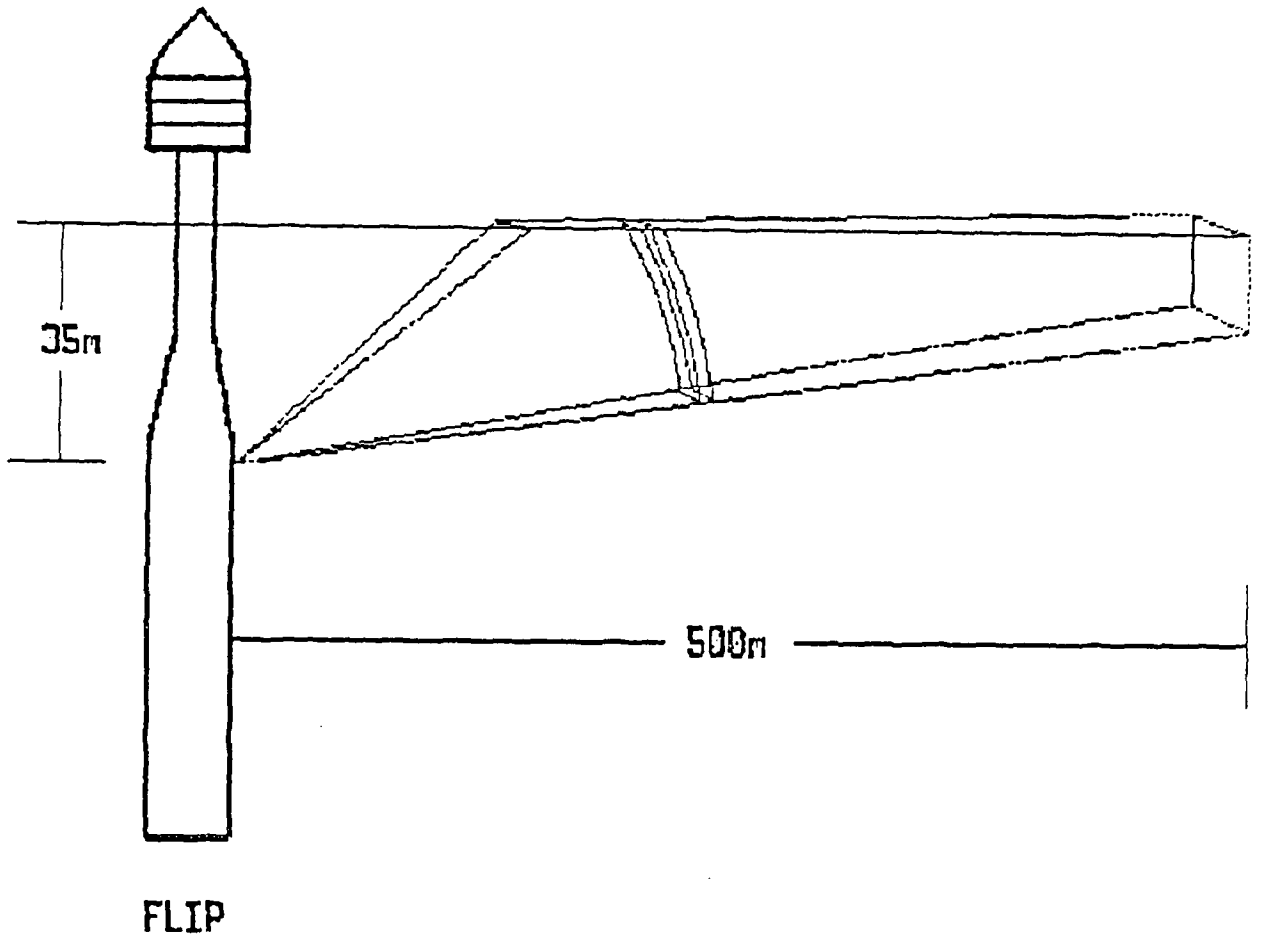


Figure 2



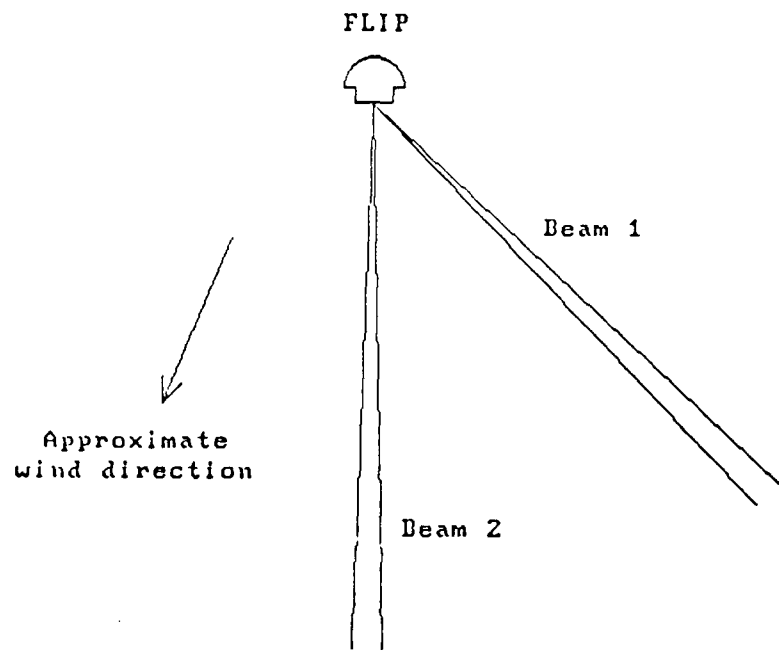


Figure 3

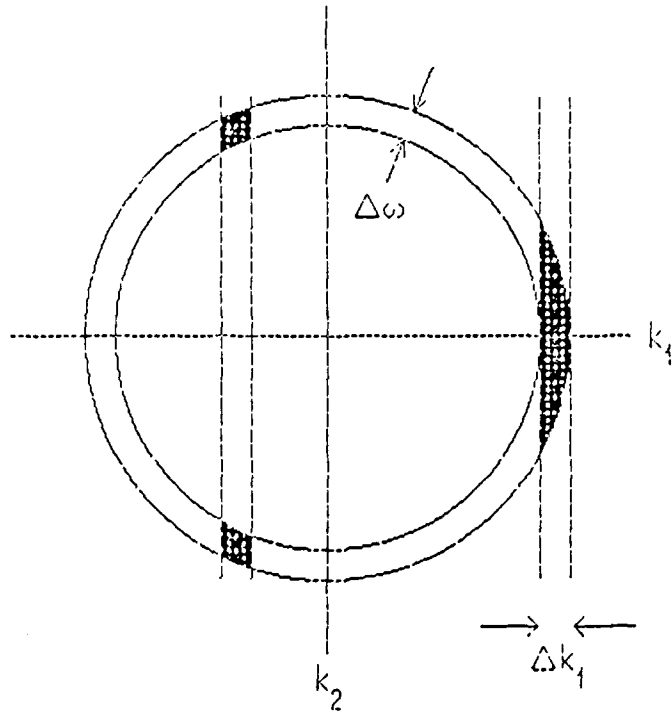


Figure 4

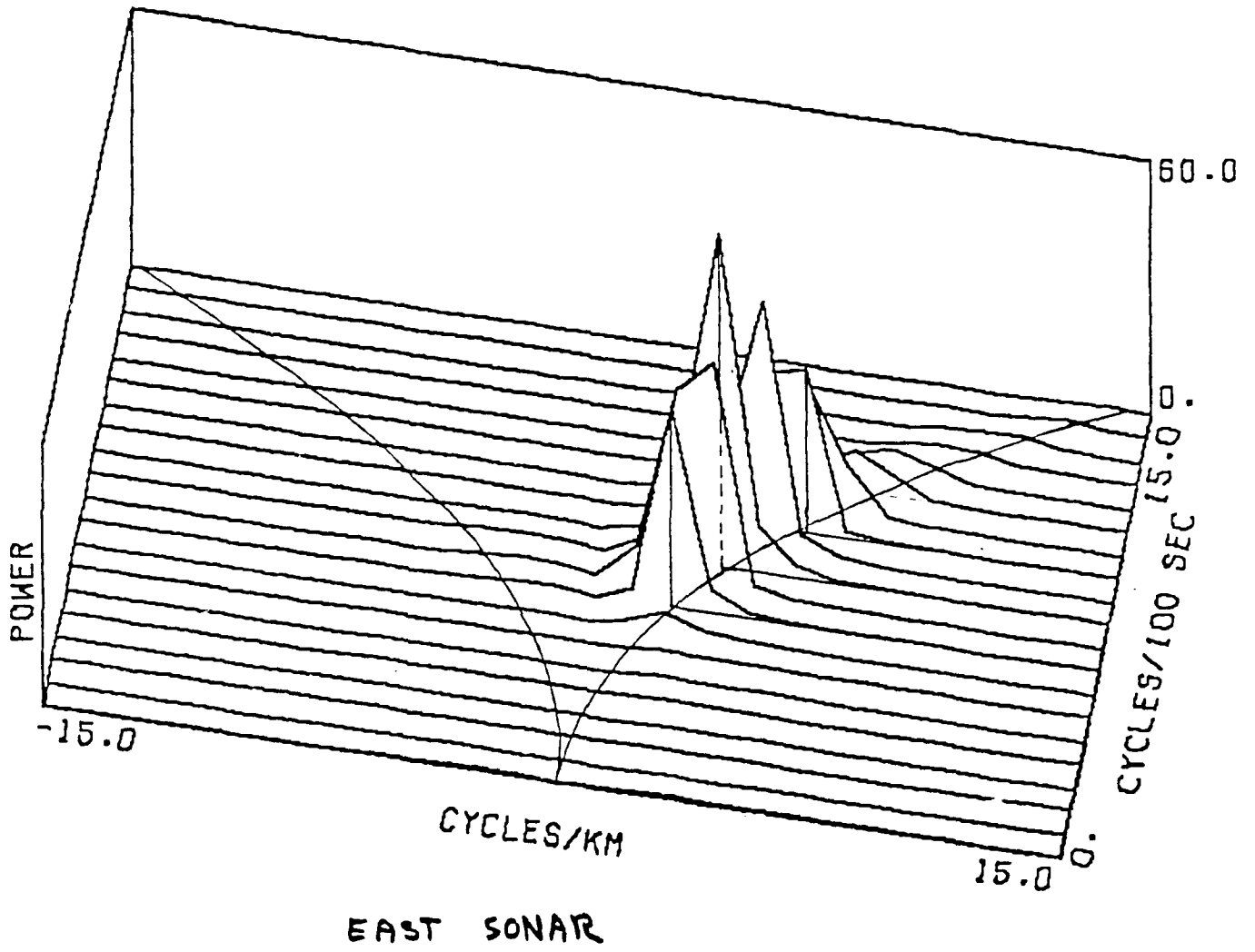


Figure 5

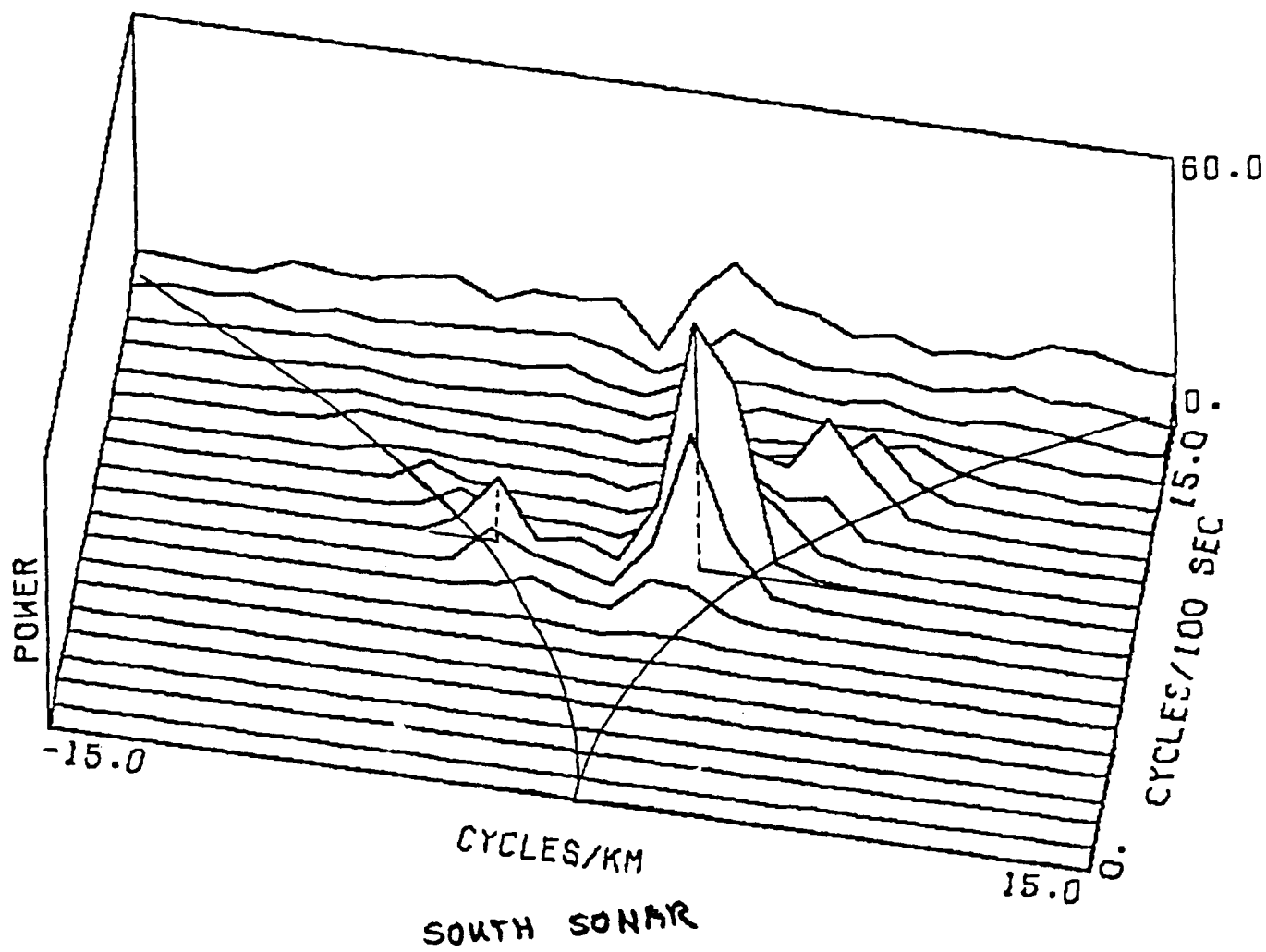


Figure 6

MEASURING OCEAN WAVES FROM SPACE: A SUMMARY OF THE LAST DECADE

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I. Some Highlights of the Last Decade

The Seasat L-band SAR in 1978 provided the first opportunity to monitor the spatially evolving directional wave spectrum over anything approaching basin scales. Several Seasat passes were analyzed over tracks of up to 1000 km. One such pass along the U.S. East Coast exhibited clear evidence in its evolving spectra of deep water dispersion, convergence to storm centers, refraction in the Gulf Stream, and shallow-water wave shortening [Beal et al., 1983, 1986]. Other passes have been analyzed by European investigators, and have shown a similar convergence to storm centers.

Techniques for optimal extraction of SAR spectra developed gradually during the early 1980's^[Talley, 1986]. Much of the European experience is described in the volume edited by Allen [1983]. Unfortunately, there were few auxiliary estimates of the simultaneous and coincident directional wave spectrum under the Seasat SAR. As the number of analyzed spectra began to accumulate,

however, it became clear that the Seasat SAR lacked the ability to respond to all but the longest waves travelling in the azimuth direction of the spacecraft and, perhaps worse, appeared to rotate other wave systems away from the azimuth direction. This Doppler-induced contamination has come to be called "the azimuth fall-off problem" [Beal et al., 1981]. The problem is fundamental for SAR imaging of the moving ocean. Because the SAR is a Doppler measuring device, it misregisters scene scatterers having radial velocity components according to the relation [Harger, 1970]

$$\Delta x_a = \frac{R}{V} v_{rs}$$

where Δx_a is the azimuth displacement
R is the range from the platform to the scene scatterer
V is the velocity of the platform, and
 v_{rs} is the radial component of the scatterer velocity.

A typical R/V ratio for Seasat is about 130 seconds, and a typical rms scatterer velocity might vary from 1 m/s to 3 m/s, depending upon sea state. This rms velocity ensemble of scatterers produces a "smearing" in the azimuth direction, i.e., an azimuth fall-off or low-pass filter in the wavenumber spectrum. There appears to be no practical way to reduce this limit, except to lower the altitude of the orbit.

The Seasat SAR mission had only a short 100 day lifetime, but it was long enough to make a convincing case for the future potential of spaceborne SAR. In spite of its severe azimuth fall-off problem, the SAR revealed

wave, current, wind, and bathymetric patterns of a richness and variety that had not been seen before [Fu and Holt, 1982]. Indeed, it can be expected that the large scale oceanographic features in SAR imagery will be of intrinsic oceanographic value, for example in the study of air-sea interaction, and in mesoscale circulation dynamics.

In 1984, the Shuttle Imaging Radar-B (SIR-B) provided a second opportunity to acquire spaceborne SAR ocean wave spectra. The Shuttle SAR was of particularly high interest because its low R/V ratio (~35 seconds) offered the potential for relief in the azimuth fall-off problem. With SIR-B, several data sets were acquired in the presence of independent directional wave measurements. One such experiment occurred off the southwest coast of Chile [Beal, 1987; Monaldo and Lyzenga, 1988]. In this experiment, underflights by a NASA P-3 containing a Surface Contour Radar [SCR; Walsh et al., 1985] and a Radar Ocean Wave Spectrometer [ROWS; Jackson et al., 1985] produced nearly simultaneous and coincident estimates of the spectrum. In addition, the U.S. Navy Global Spectral Ocean Wave Model (GSOWM) produced forecast spectra at specific grid points every twelve hours.

On most days, the three sensors (SAR, SCR, ROWS) agreed quite well, confirming the advantage of the low altitude SAR orbit. But on the lowest sea state day (1.6 m significant wave height), when most of the spectral energy was concentrated at wavelengths shorter than 100 m, substantial SAR azimuth fall-off occurred even at the low shuttle altitude. Fortunately, the waves on this day were of sufficiently low energy to be of little operational value.

On the other three days, when the wave height ranged between 2 and 5 m, the SAR spectra agreed with the two aircraft estimates as well as the aircraft estimates agreed with each other, but the model forecast was usually substantially different from any of the three sensor estimates. These results suggest that spaceborne SAR could be used to advantage to validate and improve global ocean wave forecasts.

II. Some Recent Results

Anticipating a reflight of SIR-B in early 1987, substantial European and North American resources were gathered to further examine the ability of spaceborne SAR to correctly image ocean waves in extreme sea states. The southern Labrador Sea was chosen as the central site, based upon both its proximity to aircraft bases and its extreme wave climate. According to the US Navy Climatological Atlas, some regions in the southern Labrador Sea experience up to 20 "events" of greater than 10 m significant wave height in an average winter. The SIR-B reflight (designated SIR-B') was planned for a near-polar orbit, and would have passed over the southern Labrador Sea twice daily for several days.

The Challenger accident in early 1986 caused a major reconfiguration of the experiment. SIR-B' was replaced with an aircraft C-band SAR, newly-developed by the Canada Centre for Remote Sensing. In three ways, the aircraft SAR provided even more capability than SIR-B' would have: 1) it was C-band, rather than L-band, better emulating the planned European ERS-1 and Canadian Radarsat, 2) its overpasses could be scheduled to more nearly

coincide, both temporally and spatially, with ship and wave model estimates, and 3) the aircraft altitude could be chosen to simulate the range-to-velocity ratios of both low and high altitude satellite orbits. The major disadvantages of the aircraft SAR were its relative lack of absolute calibration and its higher temporal (both minute-to-minute and day-to-day) transfer function variability.

The SIR-B' experiment evolved into the Labrador Sea Extreme Waves Experiment (LEWEX), and occurred when it was originally planned, in March 1987. Analysis and interpretation of the LEWEX results are still proceeding, with participation from eight countries in North America and Europe. Over a seven day period, thousands of estimates of the directional wave spectrum were obtained from several ship-deployed directional buoys, two ship radars, three aircraft radars (the Canadian C-band SAR, the NASA SCR, and the NASA ROWS), and nine wind-wave models, including first, second, and third generation models.

The LEWEX results, although still preliminary, appear to confirm the potential value of SAR for improving global wave forecasting. On 13 and 14 March, a major event passed directly through the LEWEX area, creating dynamic and rapidly changing wind and waves.

The LEWEX preliminary results appear to indicate that:

1. the low R/V ratio SAR estimates describe the directional spectrum of waves longer than 100 m as well as the other two aircraft estimates do, reinforcing the SIR-B Chile results,

2. the wave model forecasts, when driven with separately derived winds, can be in gross disagreement, particularly in describing the more dynamic phases of a storm, and
3. the wave model hindcasts, even when driven by common winds, continue to exhibit substantial differences which often can be resolved by the remote sensing estimates, in spite of their inherent 180° ambiguity in wave direction.

The more general implications, which still require more precise quantification, are 1) present methods for wind field determination are lacking (no surprise to advocates of satellite scatterometers), 2) at least most (and at most, all) existing wave models are lacking in some of their critical assumptions, and 3) spaceborne estimates of the global directional wave spectrum, if properly assimilated into the models, could substantially improve the present situation.

Research continues today to better understand the SAR in terms of fundamental scattering processes. For example, the dominant scattering mechanisms for range (cross-track) and azimuth (along-track) waves are fundamentally different. See Hasselmann et al. [1985] for a review of the major issues. Recent collections of papers in the Journal of Geophysical Research [Shemdin, 1988 and following 8 articles; Holt, 1988 and following 4 articles] provide insight into some of the major unresolved issues. Most of these issues involve a search for a more fundamental understanding of SAR in terms of both the long wave/short wave coupling mechanisms and the

associated radar backscatter mechanisms.

III. Future Opportunities: The ERS-1/SIR-C Conjunction

In the next five years, two major opportunities will present themselves for extending our understanding of how best to utilize global estimates of ocean wave spectra. The first opportunity will be the European ERS-1, and the second will be the NASA SIR-C. The higher altitude ERS-1 orbit will be deficient in azimuth response, however, so the ERS-1 SAR will probably offer only minor assistance to wave modelers, and even then only to the extent that its transfer function is well understood and accurately modeled.

This situation presents an interesting dilemma for the international wind-wave research community. Clearly, neither of the two upcoming opportunities is ideal, but the combination (for at least a several day period) could provide the best of both worlds: global scatterometer-enhanced wind fields from ERS-1 and global low R/V SAR-enhanced wave fields from SIR-C. One could even envision SAR-enhanced wind fields from SIR-C, since there is now good evidence that surface stress is a function of wave age, or state of development [Glazman et al., 1988]. This unprecedented combination of global wind and wave data, if properly utilized and assimilated into the best global models, could not only advance our understanding of the fundamental physics, but could demonstrate the value of spaceborne wind and wave measurements in a compelling way.

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RADAR RESPONSE TO EXTREME SEA SLOPES AND BREAKING WAVES

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A facet scattering model of microwave scattering from sea surfaces is being developed at the Applied Physics Laboratory of The Johns Hopkins University to supplement conventional two-scale models. The conventional Bragg resonant model of phase modulated scattering by a stationary and homogeneously rough sea is appropriate when applied with an instrument that matches its resolution to the electromagnetic and hydrodynamic (EMH) limit of the two-scale model. However, for an L-band synthetic aperture radar (SAR) or real aperture radar (RAR) with spatial resolution 10 to 1000 times their wavelength, specular facet scattering amplitude modulated by surface slope becomes the dominant imaging mechanism for ocean scenes that are not homogeneous over the long scale nor strictly flat within a resolution cell. Specular conditions have been observed to prevail when ocean waves break over a continental shelf, refract at the boundary of a coastal current or diverge when stressed by a strong wind. Such discontinuities in the sea surface elevation are expected to contribute significantly to acoustic noise along coast lines and in the ice margins bordering the northern oceans.

The imaging properties of a SAR in the downrange coordinate are much like those of a RAR. Steep surface slopes that directly face the radar will contribute large amplitudes to the backscatter cross section instantaneously. Temporally transient point specular reflections

[Winebrenner and Hasselmann, 1988] associated with range travelling waves are spatially periodic with regard to the measurement of the phase aperture but are probabilistic in time over the synthetic intervals defining the cell size along track. A Rayleigh-Poisson model of SAR measurement statistics [Tilley, 1986] has been shown to agree with space shuttle imaging radar (SIR-B) data for stationary scenes of both land and water. The Gaussian concept of a homogeneous surface characterized by an equilibrium wave spectrum may not be adequate for sea states that contain steep wavelets or breaking crests. Such surface conditions are manifested as spikes in the microwave radar return that are evidence of a sporadic and transient scattering mechanism. The number density of sea spikes per unit time and per unit area is in general a random process subject to deterministic correlations imposed by regular wave fields, current profiles and bottom topographics. The effect of these isolated amplitude modulations on the wavenumber response of the SIR-B instrument has been mathematically modelled and estimated [Irvine and Tilley, 1988] to approximate wave slope spectra in the Agulhas Current where wave power is intensified along the southern coast of Africa. Figure 1 shows the result of applying a stationary SAR wavenumber response function to estimate the unbiased Fourier image spectrum as well as the result of applying the dynamic response model to approximate a wave slope spectrum with both a purely random (i.e., $M=0$) and a partly deterministic (i.e., $M=1$) sea spike contribution.

The principal parameter in the facet scattering model (i.e., the mean number of sea facets detected per resolution cell during the signal integration period) contributes a discretely improbable term to the Rayleigh measurement of radar cross section. Since radar sea spikes describing the

wave breaking process are essentially hard limited (i.e., thresholded to some limiting maximum value) and sparsely distributed over the ocean surface. Such rare events stretch the Fourier concept of an infinite ensemble of spectral components. Hence, a proper consideration of the statistical nature of the specular cross section may be better formulated in terms of the surface elevation in space and time, rather than the spectral energy density in wavenumber and wave frequency. SAR or RAR images of breaking waves that are amplitude modulated by extreme sea slopes will allow visual identification of the precise spatial locations of those isolated surface events that are the source of acoustic noise in the underwater environment. Figure 2 depicts an isolated group of developing wind waves breaking in the Sea of Japan and an organized series of coastal breakers witnessed there by the SIR-B in 1984.

Microwave radars are capable of both general geographic estimates and site specific documentations of wave breaking induced by coastal currents, bottom topography and atmospheric stress. A specular cross section model can be employed to characterize the sea surface in terms of the mean number of specular facets per unit area per unit time. The facet model may be interpreted in either the Fourier spectral domain or the SAR image domain to specify slope distributions and wave height discontinuities that contribute to acoustic noise below the ocean surface.

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Figure Captions

Figure 1

The Fourier power spectrum for the SAR image of the Agulhas Current scene has been calibrated in units of meters² to portray wave slope variance. Note that the solid white graph depicts the dynamic response correction estimated with a random approximation ($M=0$) and applied assuming a deterministic correlation ($M=1$) between sea slope amplitude and swell wave phase.

Figure 2

The SIR-B synthetic aperture radar recorded specular enhancements of breaking wave groups for 100 meter wavelengths in the Sea of Japan.

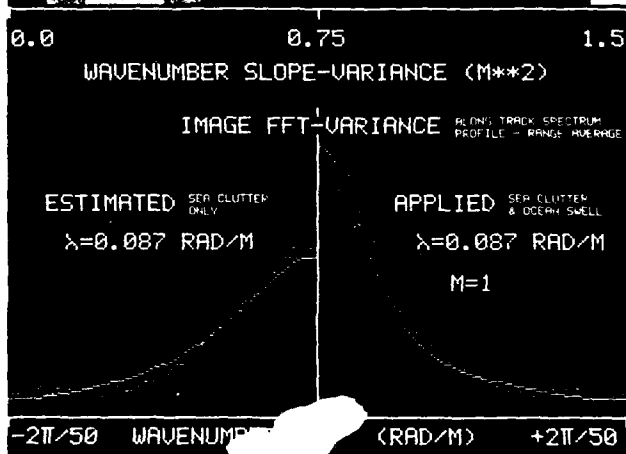
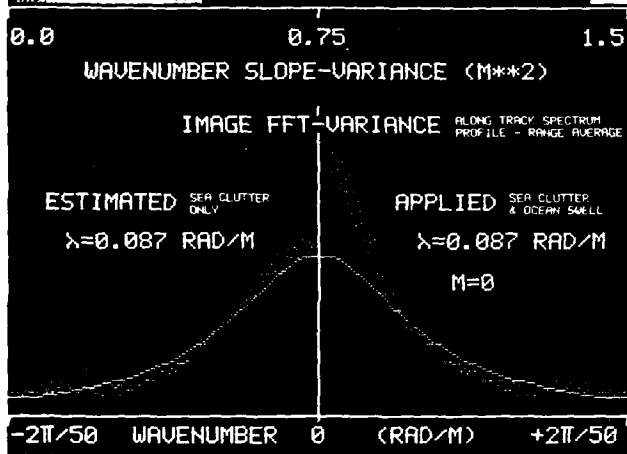
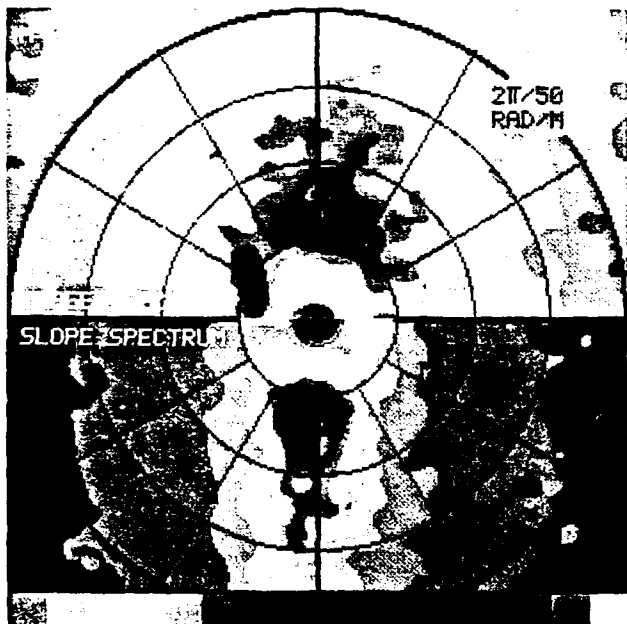
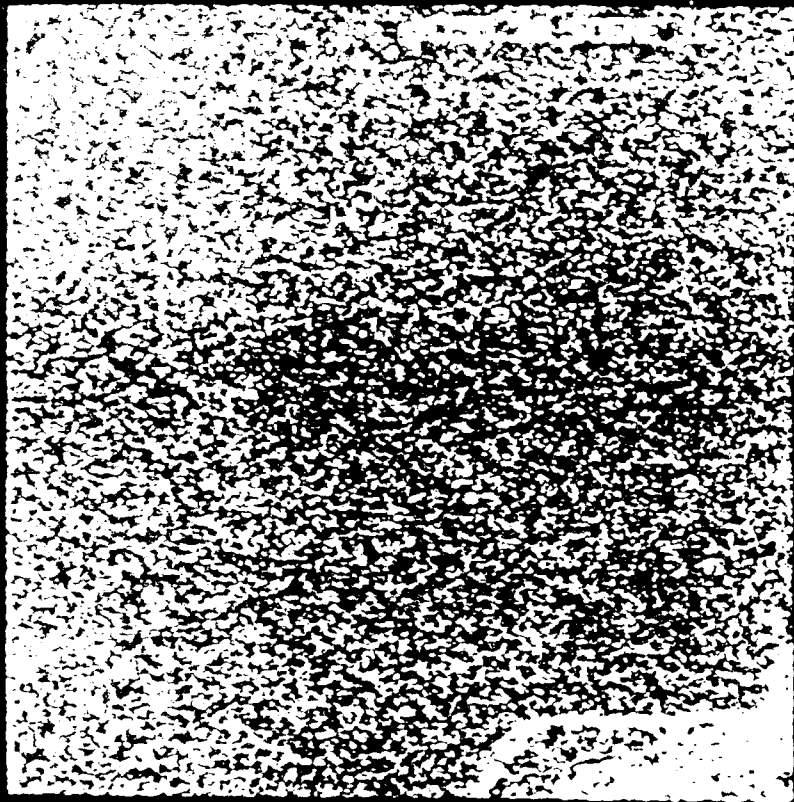


Figure 1 - Tilley



SOUND GENERATED BY BREAKING WAVES

Notes for the ULF/VLF Workshop, November 1988, Austin Texas.
David Farmer

Almost all previous studies of surface generated sound in the ocean have focussed on long term averages of the measured spectrum. In fact the designation "ambient noise", indicates the presumption that the signal is unwanted and not normally considered for what it can tell us about ocean surface processes. A deeper understanding of the physical mechanisms of generation and propagation will surely be useful, both in overcoming interference of this "noise" and also in using it as a probe of ocean conditions. Measurements of the sound spectrum and of its directionality need to be extended to include temporal variability at the time scales of the generation mechanism, and to be incorporated in experiments where the surface processes are adequately resolved.

Recent work on the temporal variability of ambient sound, and of fine structure in its spectrum, have exploited the higher frequencies (>100 Hz), but the results suggest the merit of combining high frequency with low frequency measurements as a useful probe of surface processes. Some of these results are briefly outlined below, with suggestions as to their potential significance in future experiments.

Bubble clouds and the ambient sound spectrum

The Knudsen curves are heavily averaged and reveal none of the fine structure in the spectrum. Our recent observations show that although the spectrum of individual breaking events generally satisfies the -17 to -20 dB/decade slope, there are peaks in the spectrum that can be

explained by the presence of the bubble layer. The near surface bubble layer is approximately exponential with e-folding depth 1-4m (see Fig.1); it produces a distinct sound speed reduction near the surface, thus leading to an upward refractive condition. Sound is trapped in the resulting wave-guide, but near the cut-off frequency of each mode there is leakage of the signal into deeper water causing peaks in the observed spectrum. First mode leakage has been observed at frequencies of 1-3 kHz.

Bubbles influence ambient sound in a different way. At higher frequencies sound is so strongly attenuated by the bubble clouds that the ocean actually becomes quieter with increasing wind speed² (at frequencies greater than about 10 kHz).

Temporal Variability

In a fetch limited study² we found that wave-breaking was modulated by the group structure, consistent with previous visual observations³ (see Fig.2). Thus the sound generated by breaking waves included a time varying component at twice the nominal wave period. (This may be compared with the hypothesized nonlinear interaction source that leads to a pressure fluctuation of one half the wave period.)

On the other hand more recent observations in the open ocean suggest that although there is a longer term modulation of the pattern of breaking events, the dominant period of sound fluctuation is the same as the dominant wave period. The way in which wave breaking is influenced by wave conditions, especially presence of swell, and wind conditions, is not well understood. Acoustic observations provide a sensitive probe of these processes and can provide valuable information

in any experiment where the wave field, and wave breaking in particular, is important.

Use of temporal variability to identify lower frequency contributions

The fact that we can easily identify the occurrence of breaking events, both with video cameras on our instrument and also from their high frequency emissions, allows us to search for lower frequency contributions that occur simultaneously and can therefore be attributed to the breaking event. Our observations with a hydrophone 14m below the surface in 200m water depth, show unambiguously that the breaking wave, radiate sound at least down to 50 Hz (see Fig.3a,b).

Variance and pattern of sound from breaking waves

The depth of measurement is clearly important in determining the signal variability. Separation of contributions from individual waves requires near surface observation, and in fact the variance of the signal as a function of depth has been directly related to the distribution of breaking events on the ocean surface² (see Fig.4a,b).

For a hydrophone close to the surface, it is obvious that the relative position of the breaking wave is important, especially since this position changes during the breaking process. The actual shape of the wave, as well as the depth and nature of the source, will modify the radiation pattern. Preliminary calculations suggest a contribution to anisotropy (Fig.5). It would be desirable to see whether observed horizontal anisotropy in the sound field can be related to the wave shape and distribution.

Summary

Various features of the surface wave field and bubble layer contribute to low frequency sound in the ocean. Future experiments should include a strong oceanographic component so as to allow the connection between ocean surface processes and the ambient sound field to be clearly established.

¹Farmer, D.M. and D. Lemon, 1984. The influence of bubbles on ambient noise in the ocean at high wind speeds, J. Phys. Oceanogr., 14, 1762-1778.

²Farmer, D.M. and S. Vagle, 1988, On the determination of breaking surface wave distribution using ambient sound, J. Geophys. Res., 93, C4, 3591-3600.

³Donelan, M., M.S. Longuet-Higgins and J.S. Turner, 1972. Periodicity in whitecaps, Nature, 239, 449-450.

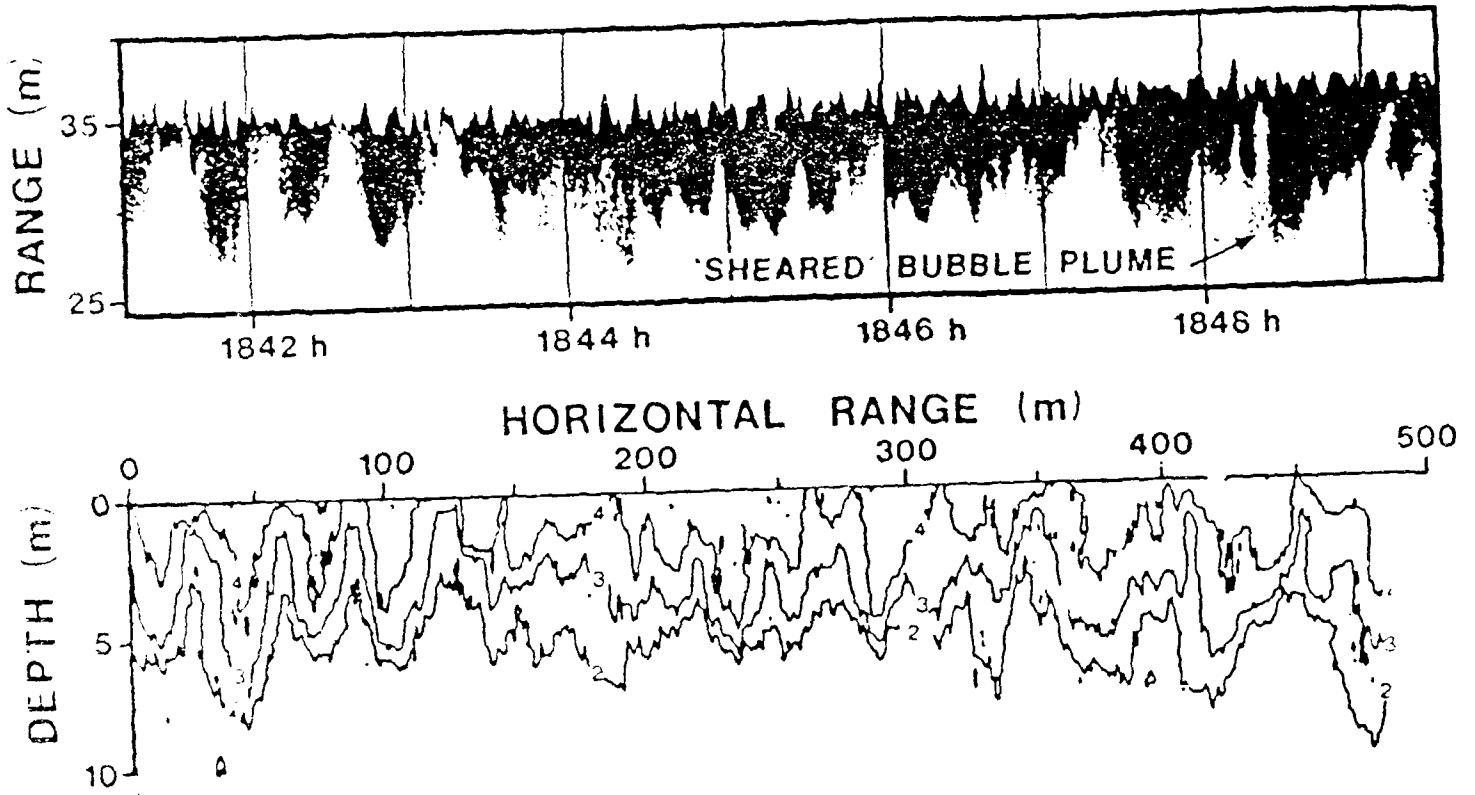


Fig. 1. Bubble clouds (above) and calculated target strength contours obtained with an upward pointing sonar mounted on a submarine (after Farrer & Crawford, J. Geoph. Res., 92(08), 8231-8243, 1987).

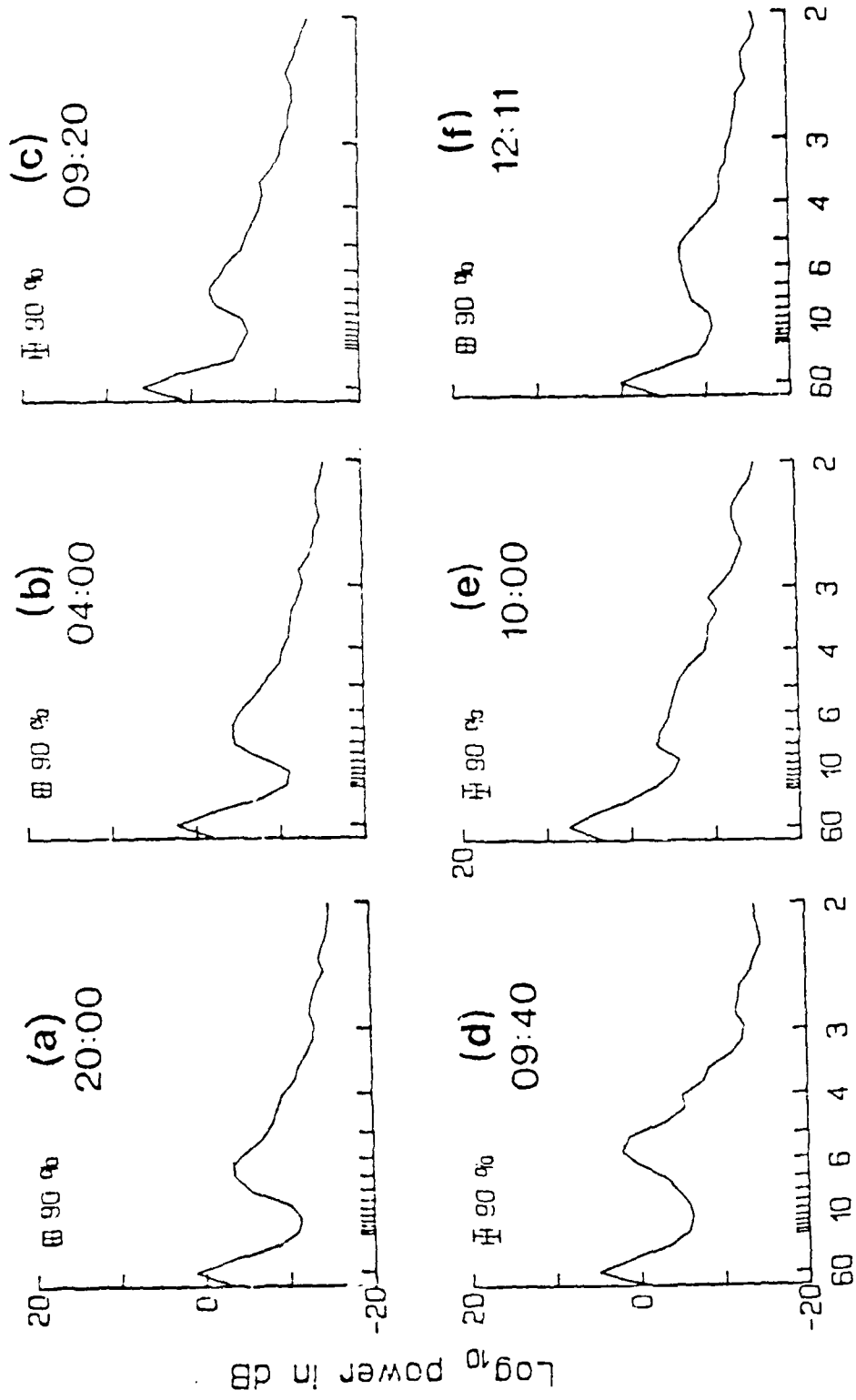


Fig. 2. Sequential spectra calculated from the time series of ambient sound spectral level at 4.3 kHz, recorded beneath a fetch limited sea. Note the two peaks, one at about 7.5 periodicity, associated with the repetitive nature of wave breaking at twice the dominant wave period, and a 60s peak associated with passage of wave groups over the instrument (adapted from [1]).

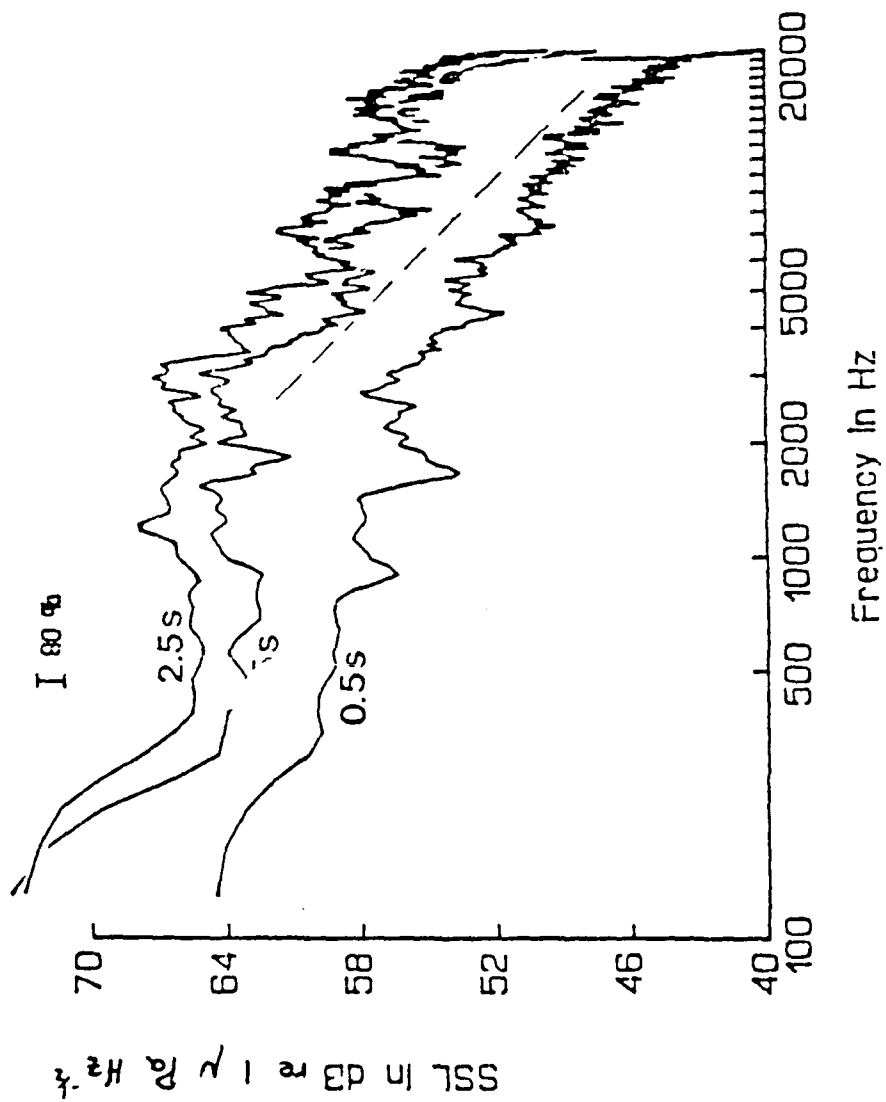
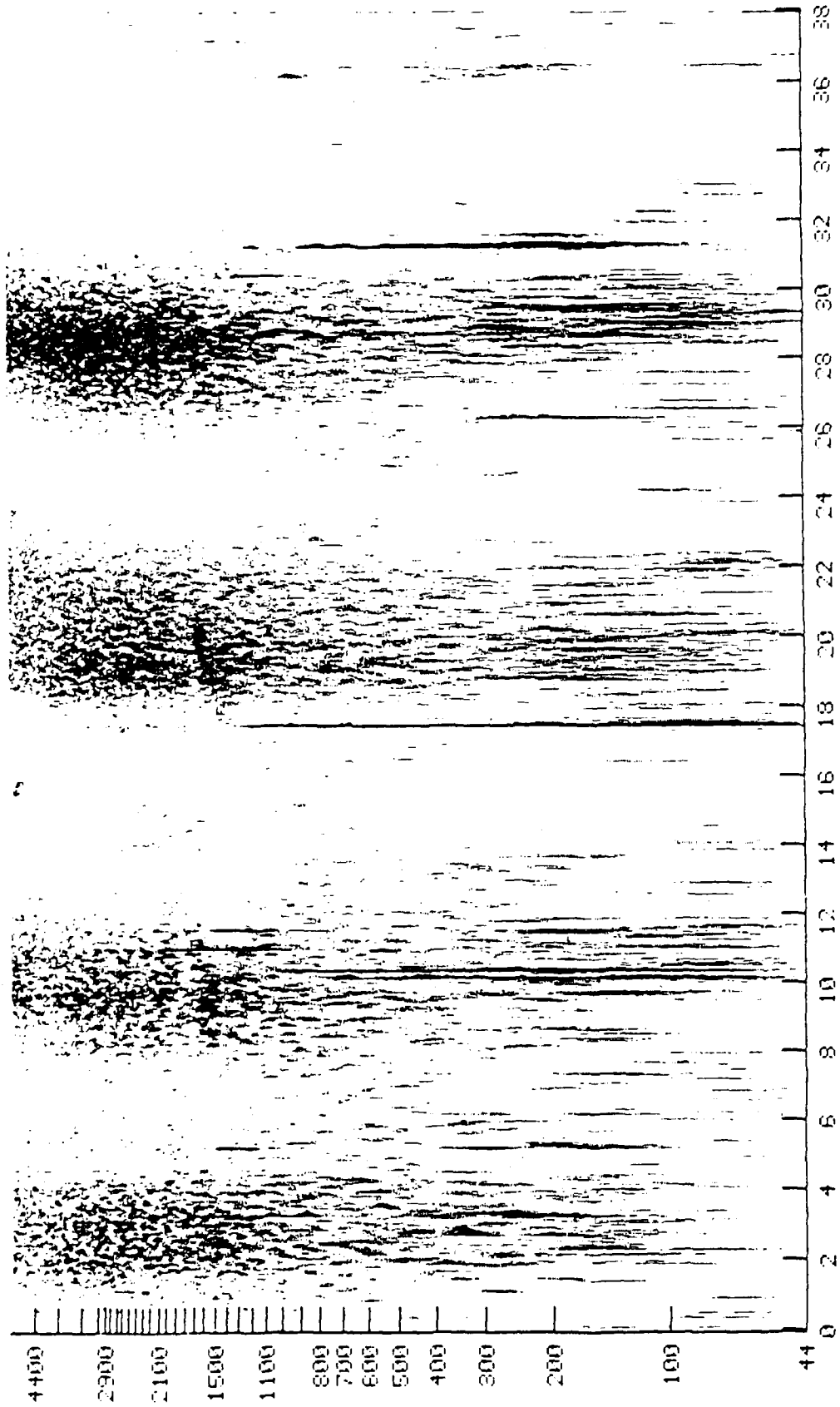


Fig. 3a. A sequence of 3 ambient noise spectra obtained at 1s intervals, at a depth of 14m, during passage of a breaking wave. Note that the spectrum is modulated in amplitude over the full range, 100 Hz - 20,000 Hz, shown here.

Acoustic Spectra: La-Perouse Bank 8, 9, 11, 13

SSL (dB) 0.0 4 .8 1.2 1.6 2.0 2.4 2.8 3.2 3.6 4.0 4.4 4.8 5.2 5.6 6.0

Time(PST): 17:54:32 -- 18:04:31 1024 points fft 5 averag



Time (Seconds) S. Nagle Nov/88

3b. Spectrograms of a breaking waves, showing that sound is radiated during the breaking event at least down to 50 Hz.

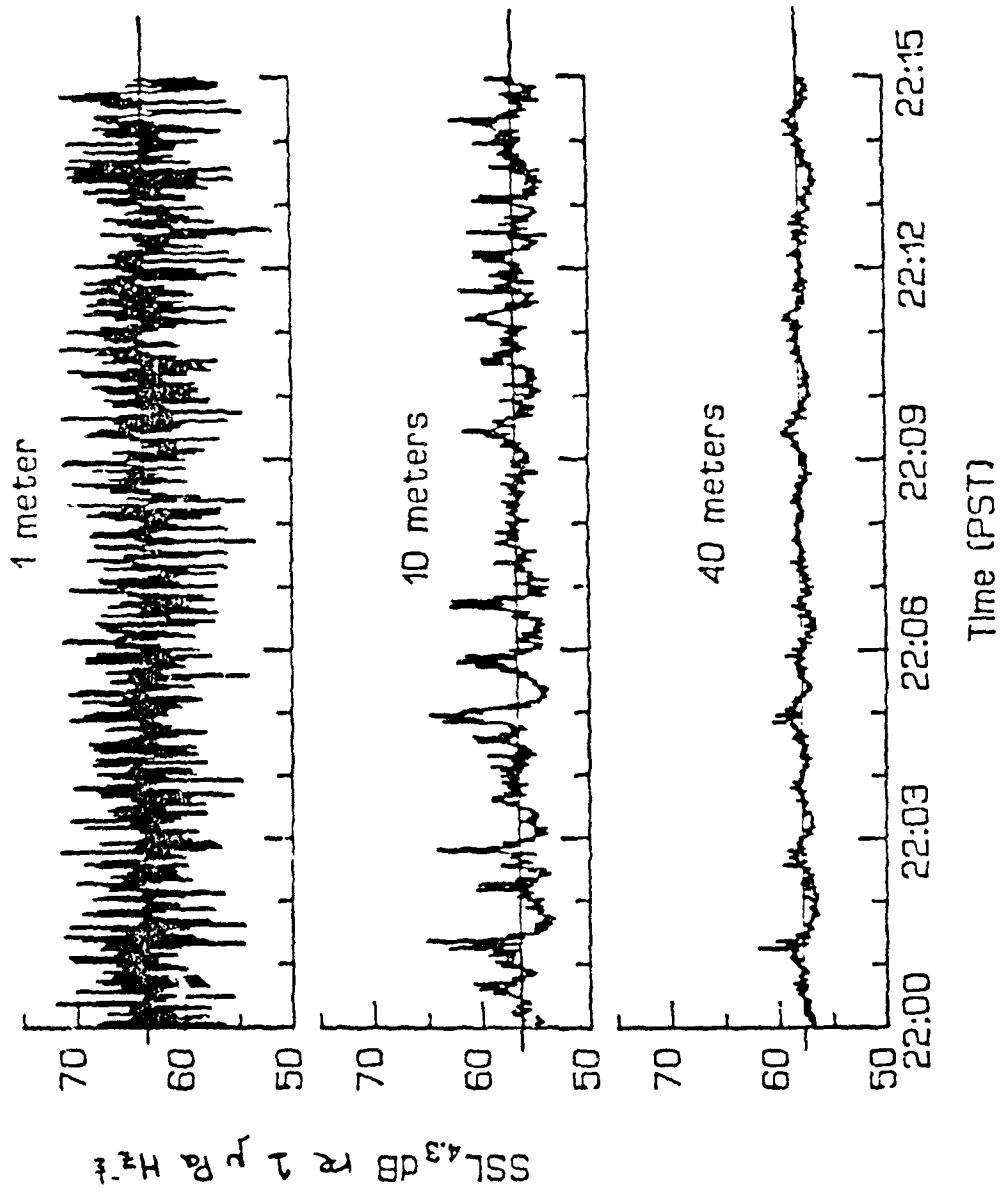
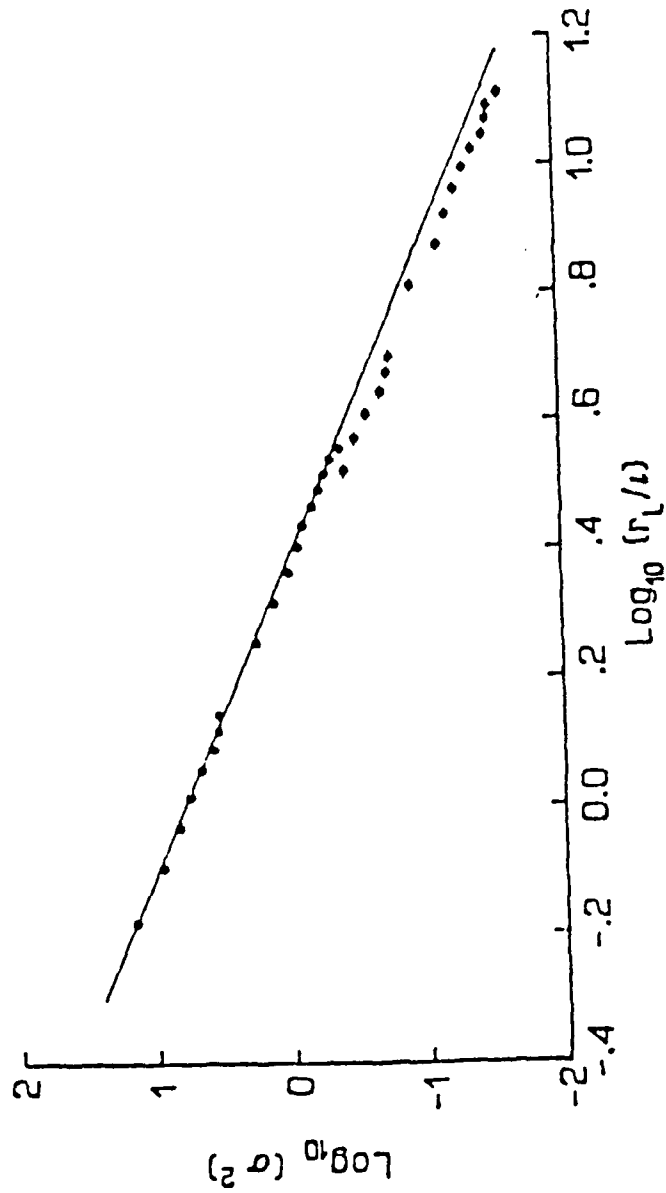


Fig. 4a. Time series measurements of sound spectrum level at a frequency of 4.3 kHz, at depths of 1m, 10m and 40m, showing the decrease in variance with increasing depth (adapted from ref. 2).



4b. Relationship between variance, σ^2 , and effective listening radius r_2 divided by mean spacing of breaking waves λ , derived from a simple computer model of random source on an ocean surface. The listening radius increases with hydrophone depth and is dependent on effects of spherical spreading and attenuation.

Sound Pressure Contours (Infinitely Wedges)

Source position: $r=0.5m$, $\theta_{max}=115.359$

Frequency: 4.5 kHz

$r_{max}=0.000$ $r_{min}=1.000$

Interval= 1.000

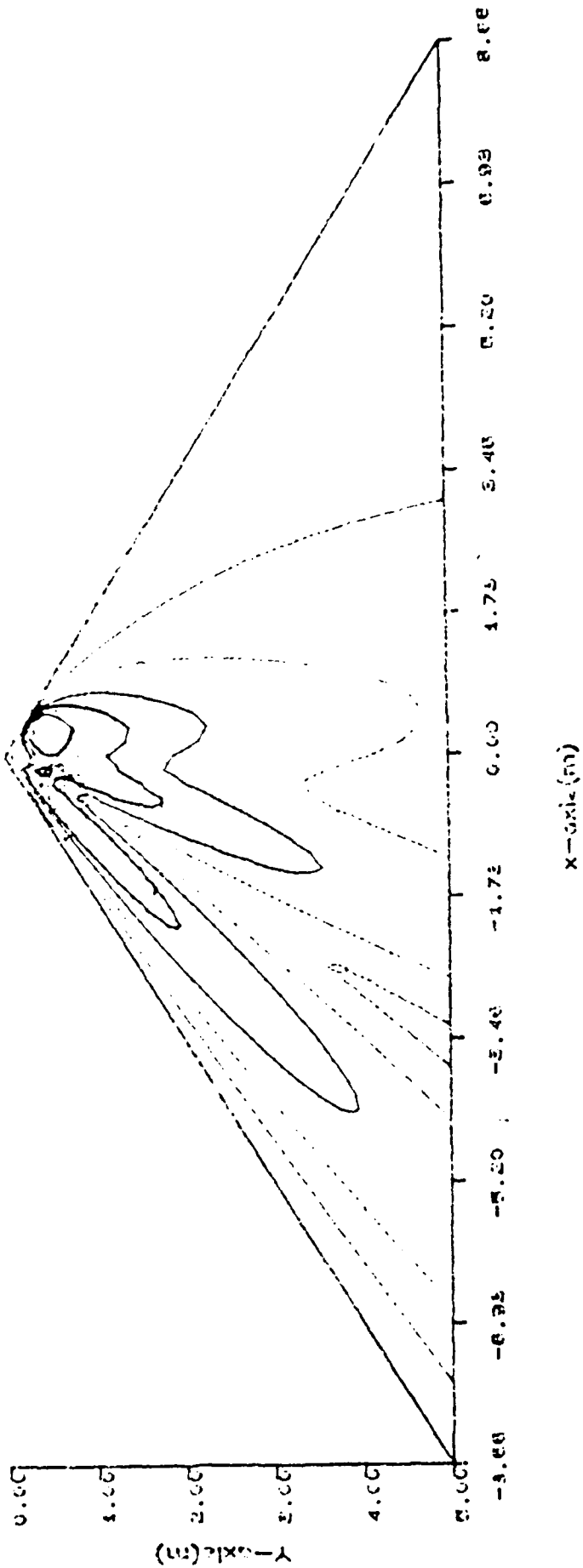


Fig. 5. Calculated sound pressure field in a wedge, which might provide a very simple model of a surface wave, showing anisotropy introduced by placing the sound source slightly off-axis as would be expected in a real breaking event.

Low Frequency Energy in the Nearshore

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Energy input to the nearshore from wind-driven waves and swell is nonlinearly transferred into motions with time scales from fractions to thousands of seconds. The low frequency (from a seismic view) component of the spectrum is classed roughly into incident (gravity) band (1 - 0.05 Hz) and infragravity band (0.05 - 0.005 Hz).

Incident Band

The bottom pressure signal of a surface gravity wave is depth-attenuated by a factor of $[\cosh(kh)]^{-1}$, where h is the local depth and k the wavenumber. Thus, typical ocean waves have negligible pressure signal in water deeper than the shelf break. The shoaling of the incident band of the spectrum on narrow continental shelves is reasonably well understood. On wide shelves, shoaling can not currently be modelled due to anomalous dissipation at intermediate depths. In shallow water, surface gravity waves become depth-limited and dissipate energy through a surf zone. However, a component can be reflected at the shoreline, producing a locally standing pattern (with associated second order pressure signal at twice the frequency that is detectable seismically), with shoreline amplitude, a_s , given by

$$\epsilon_s = a_s \sigma^2 / g \beta^2 \approx \kappa.$$

κ is approximately constant with value in the range 1 - 3, σ the incident frequency, g is the acceleration of gravity and β the beach slope. The standing wave has a cross-shore energy decay given by

$$a^2(x) \propto \left(\frac{\sigma^2 x}{g \beta} \right)^{\frac{1}{2}}$$

for cross-shore positions greater than about one wavelength offshore but still in shallow water ($\cosh(kh) \sim 1$).

Infragravity Band

Infragravity band motions are driven by non-linearities in a modulating incident wave field. In deep water these result in a bounded long wave, directly forced and trapped to the incident wave group. Surface magnitudes are typically 10% of the incident wave height. Bottom pressure signals again vary as $\cosh^{-1}(k_L h)$ where k_L is the wave number of the long wave. In shallow water the velocity of the forcing group and the phase velocity of the forced long wave are close and the interaction becomes near resonant.

The nearshore can act as a natural wave guide for these resonant interactions, with the trapping resulting from reflection at the shoreline and refraction offshore. At any frequency, the trapped modes on a plane beach of slope β consist of a series of longshore progressive modes called edge waves, with discrete wavenumbers, k_y , given by

$$k_y = \sigma^2 / \{g \sin [(2n+1)\beta]\} \quad n = 0, 1, 2, \dots$$

where n is the mode number. In the cross-shore, the wave has n zero crossings before exhibiting an exponential decay (figure 2). A convenient cross-shore scaling, the distance at which the phase velocity just equals shallow water phase velocity $(gh)^{1/2}$, is given by

$$\chi = \sigma^2 x / g\beta = (2n+1)^2$$

For smaller wavenumber, $k_y < \sigma^2/g$, a continuum of leaky modes exist that will lose energy from the nearshore to the deep sea. Cross-shore scaling is roughly the same as for standing incident band waves (figure 2).

For a typical infragravity period of one minute on a beach of slope 0.02, longshore wavelengths of edge waves range from 110 (mode 0) to 5.6 km (highest, or cutoff, mode). Shoreline wave heights (calculated from total variance in the infragravity band) are of order 0.7 H_s , where H_s is the offshore significant wave height. However, both height and typical period vary with the surf similarity parameter, $\sigma^2 H_s / 2g\beta^2$, an empirical measure of beach reflectivity for incident band waves.

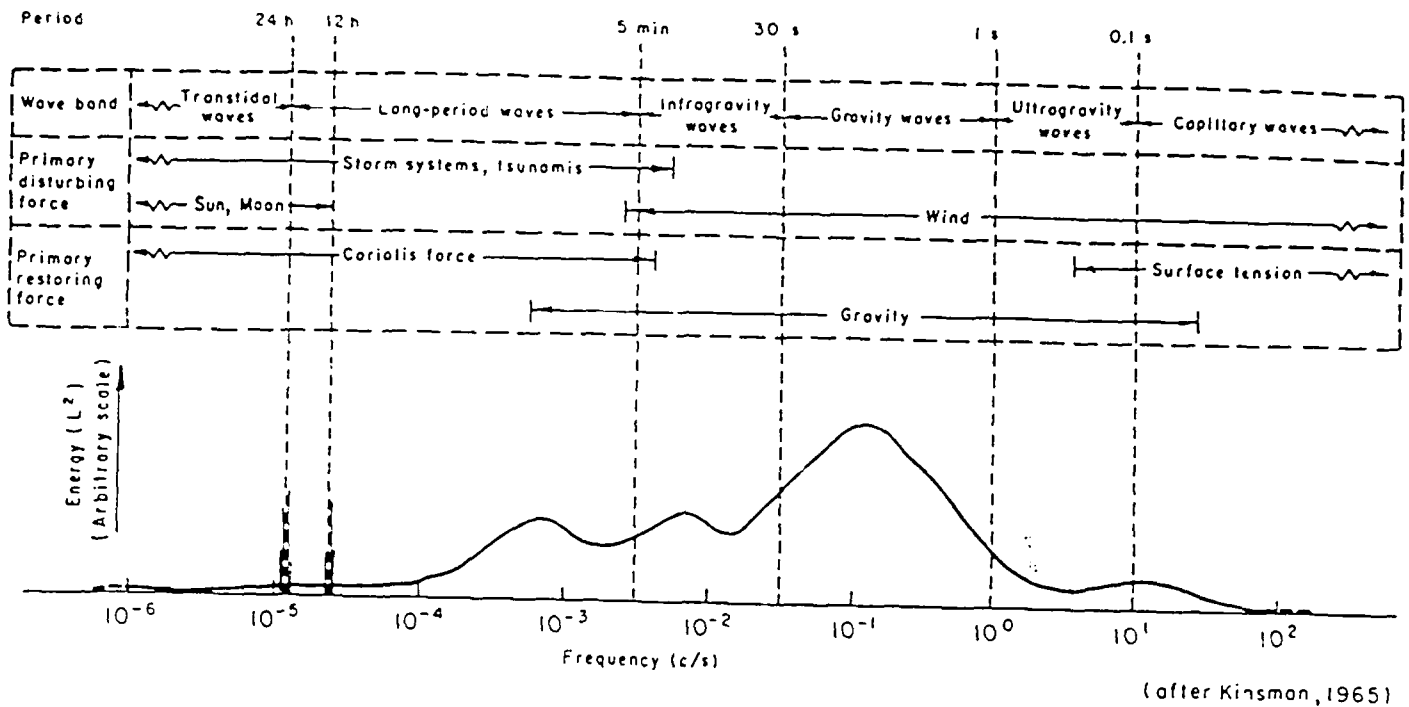


Figure 1. Approximate distribution of ocean surface wave energy illustrating the classification of surface waves by wave band, primary disturbing force, and primary restoring force.

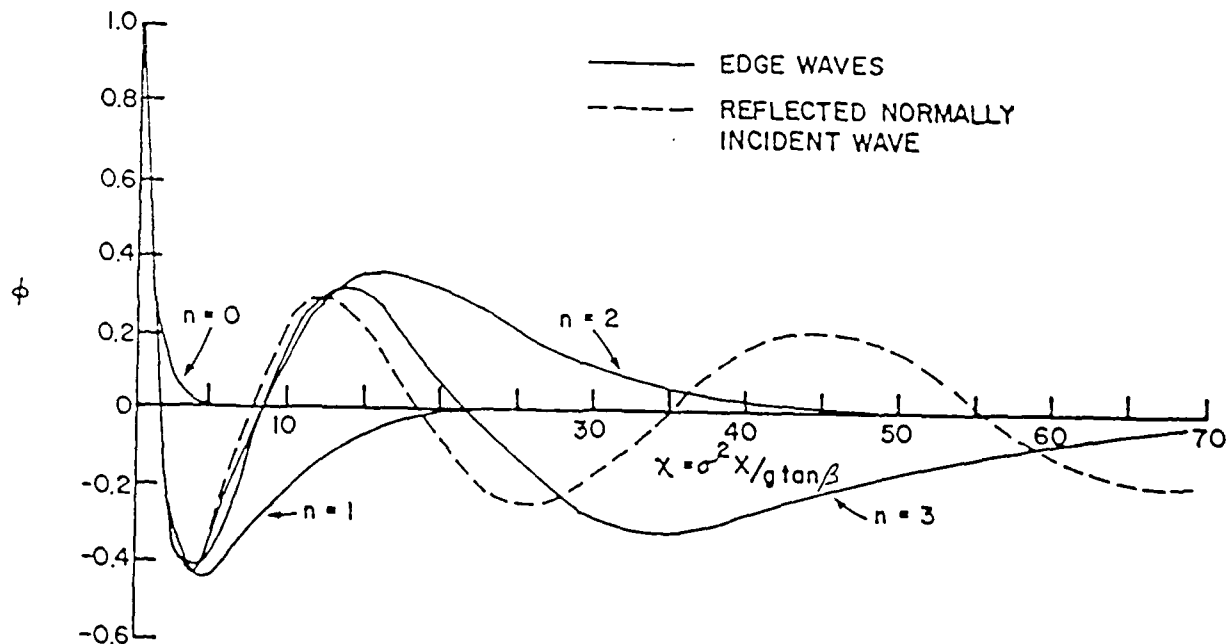


Figure 2. The offshore structure of edge wave modes 0 to 3 plotted in terms of nondimensional offshore distance χ . Note the similarity between modes, particularly near the shore. Note further that the standing incident wave, an example of the leaky waves, also looks similar.

VLF NOISE IN THE ARCTIC OCEAN

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The Arctic ocean is unique on this planet because it is mostly covered by ice. In most of the world's oceans, noise on the seafloor in the frequency range 0.01 to 50 hz is probably related to pressure changes produced at the sea-surface by wind stress or shipping, or to elastic stresses produced at the sea floor by geologic processes. In the Arctic ocean the ice covering the sea-surface will greatly dampen the coupling of wind stress and it will also considerably modify the acoustic reflection coefficient from its usual value of -1, thereby modifying sea-surface acoustic boundary conditions. This unique situation could greatly alter the noise characteristics. For instance, the microseism peak at about 0.12 hz might be shifted in frequency and greatly reduced in amplitude. Surface shipping is non-existent for much of the year in the Arctic and this will also reduce noise levels. The largest natural noise sources are expected to be impulsive and related to the breaking of ice.

Because it is a low energy environment in terms of physical oceanographic parameters and man made noise sources, the Arctic Ocean provides a unique laboratory to study the physics of VLF noise generation and propagation. To our knowledge there are no VLF noise data in the Arctic.

APPENDIX B

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ULF/VLF Workshop
29 November - 1 December, 1988

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FIELD	GROUP	SUB. GR.				
19. ABSTRACT (Continue on reverse if necessary and identify by block number) The objective of the ONR Accelerated Research Initiative (ARI) in ULF/VLF seismo-acoustic noise in the ocean is to improve our quantitative understanding of the physical processes that generate the acoustic/seismic ambient noise field in the frequency band from 0.001 to 50 Hz as a function of location and time (pp. A2-12): - to experimentally identify and quantitatively define the important noise sources including the discrimination between noise from local and distant sources; - where possible, to theoretically establish and describe the physical processes which are the noise source in the ULF/VLF frequency band; and - to determine the effects of ocean, bottom and sub-bottom acoustic/seismic parameters on the observed noise field, i.e., the effects both of propagation and of local site conditions. It is understood that only "natural" noise sources will be emphasized and that, e.g., sources from shipping will be considered only as required to evaluate their contribution to the noise field.						
20. DISTRIBUTION/AVAILABILITY OF ABSTRACT UNCLASSIFIED/UNLIMITED <input checked="" type="checkbox"/> SAME AS RPT. <input type="checkbox"/> DTIC USERS <input type="checkbox"/>			21. ABSTRACT SECURITY CLASSIFICATION Unclassified			
22a. NAME OF RESPONSIBLE INDIVIDUAL Dr. George H. Sutton			22b. TELEPHONE NUMBER (Include Area Code) 914-687-9150		22c. OFFICE SYMBOL	

General characteristics of important noise sources and of the noise power spectrum are already fairly well established over much of the frequency band of interest; most are related to ocean waves and their meteorological sources (see, e.g., reports of Frequency Oriented Working Groups, pp. 2-9 and talk abstracts, pp. A97-136). Ice motions, earthquakes, volcanic eruptions, turbulent tidal (or other) water currents, and atmospheric acoustic disturbances (also discussed in the Proceedings), are other possible sources that might be considered where appropriate.

Propagation through the ocean/crustal waveguide between source region(s) and receiving instrument arrays produces important effects on the observed ambient pressure and the seismic motion components. Relationships among these components depend strongly on the wave type (e.g., free traveling waves, such as body, Rayleigh, Love, Stoneley/Scholte, or forced deformations of the ocean-bottom produced by short-wavelength pressure disturbances) and upon the detailed velocity/attenuation structures of the oceanic waveguide; in general, the relationships also are strongly frequency dependent.

The elastic properties of the ocean floor differ by orders of magnitude between soft sediments and igneous rock - sometimes over very short distances; such local site variations can degrade array performance. Improper coupling of seismic sensors to soft sediments results in serious signal distortion and excess noise, especially for the horizontal components of motion.

The program outlined in this paper of

- 1) theoretical investigations,
- 2) analysis of existing data,
- 3) development and testing of improved measurement techniques, and
- 4) integrated field experiments,

if followed during the ARI, should produce a significant improvement in our knowledge of the ULF/VLF ambient noise field and our ability to predict its geographical and temporal variations. Each of the four elements of the program are considered to be of equal scientific priority.