'The global coverage will remain uneven without deployment of instruments on the ocean floor. While the necessary technology exists for continuous operation of land-based stations, many developments are still necessary for the ocean bottom stations. ...the availability of these data is of critical importance to the truly global sampling of the earth... All efforts that would make it practical in the near future should be strongly encouraged.'

[Science Plan for a New Global Seismographic Network, IRIS, 1984]

'We also propose that several crustal holes be used as seismic observatories. Recent major advances have been made in improving upon radially symmetric models of the Earth's interior, both in the core and mantle. These results provide fundamental information about the driving mechanisms of plate tectonics. The resolution of tomographic imaging of the Earth's interior, however, is severely limited by lack of seismological data from the ocean basin.'

[Recommendations of the COSOD II Report, JOIDES/ESF, 1987]

Cover: Surface of the Earth is divided into 128 "squares" of a dimension, roughly, 2000 x 2000 km. Light shading means that a station in a given square either already exists, is planned or there is land, including oceanic islands, where a standard seismographic station could be deployed. Dark shading shows the contribution to the global coverage of 20 ocean bottom observatories; only 5 isolated squares remain empty.

Acknowledgements

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We were particularly grateful for the substantial involvement of colleagues from Japan, Canada, United Kingdom, France, Germany, Switzerland and Italy whose participation was critical to the success of the workshop and without whom our breadth of knowledge concerning new results and planned programs would have been severely limited.

Discussion leaders Don Forsyth and John Orcutt played an important role in the workshop and were primarily responsible for the preparation of the text in Sections A1 and A2 of this report.

CONTENTS

SUMMARY

The Earth is the most accessible example of condensed matter in the universe. Understanding its structure helps us to understand its origin, the dynamic forces at work within it in the past and present, and its evolution as a planetary body. Within the Earth are two great machines. One involves convection in the Earth's solid mantle and drives the ponderous motion of the plates, giving rise to earthquakes and volcanic eruptions. The other involves convection in the liquid core and produces the Earth's internal magnetic field. An ultimate goal of geoscientific research is to understand the dynamics of this interior machinery.

The availability of data, which represent the remote sensing of the physical properties of the Earth's deep interior, is the necessary condition for the progress towards this goal. Some very encouraging results on whole Earth tomography have been obtained during the last several years. Yet, the current resolution of these images is very low. To improve it on a global scale, it is necessary to deploy geophysical observatories on the ocean bottom. This represents a number of technical problems; hence the need for a workshop.

The workshop was held at Woods Hole Oceanographic Institution (WHOI) on 26-28th April, 1988. The convenors were G.M. Purdy (WHOI) and Adam Dziewonski (Harvard University). The attendance for the workshop spanned the global and marine seismology communities and totaled 70 persons, of whom 12 represented institutions outside the U.S.

Four goals for the workshop were defined:

- describe, and establish the importance of the scientific problems that could be solved using data recorded by ocean floor or downhole broad band seismometers.
- determine how such an initiative would overlap and interact with existing and new initiatives in global and marine seismology.
- examine the technical challenges in, for example, sensor design and emplacement, data recording and telemetry, for which thorough solutions would have to be devised before the initiative could be considered feasible.
- generate a menu of ideas for further work and pilot experiments that would constitute tangible progress towards the eventual goal of a global observatory of ocean floor seismic stations.

The structure of the workshop itself is described in the Introduction to this report. The important conclusions of the workshop can be considered in three categories: the definition of the primary scientific objectives, the description of the major challenges related to the recording of high quality data, and a list of specific recommendations. Summaries of these three important sets of conclusions are presented here.

SCIENTIFIC OBJECTIVES

Scientific objectives that can be addressed with seismological data from long-term ocean floor observatories can be considered in six broad subject areas, the first three of which we consider to be primary.

- Global earth structure - some examples of key questions are: is the inner core heterogeneous or anisotropic? What is the geometry of the core-mantle

boundary? Are hotspots correlated with slow regions at the base of the mantle? What is the nature of lower mantle anisotropy? What is the geometry of the 400 km and 670 km discontinuities?

- Oceanic upper mantle dynamics and lithosphere evolution - can seismic anisotropy be used to map flow in the upper mantle? What is the degree of lithospheric thinning beneath hotspot swells? What are the spatial variations in the depth extent of anomalous structures beneath ridges? Do oceanic plateaus have roots like continents? What is the form of small-scale convection beneath plates?

- Earthquake source studies: Ocean floor stations are needed to improve source location (particularly depth), focal mechanism and rupture process determinations. These measurements are critical to studies of the depth of the seismic decoupling zone, the depth extent of outer rise events and the rheology of the oceanic lithosphere. Near field data, in particular ocean floor recordings, are needed to improve the resolution of the source mechanisms of events not caused by faulting but by slumping or magmatic injection. Such studies have important implications for estimation of long-term seismic hazard.

- Oceanic crustal structure.

- Tsunami warning and monitoring.

- Sources of noise and the propagation of noise.

Broad-band, long term ocean floor observatories are needed for all these studies. Generally only seafloor stations can provide uniform global coverage in areas without islands and they are needed for regional studies of individual tectonic features and for sampling wave propagation in 'normal' oceanic lithosphere. Oceanic islands are, by definition, located on anomalous structures with thick crust and, in many cases, unusual upper mantle velocities.

EXPERIMENTAL AND TECHNICAL ISSUES

Four problem areas need investigation and experimentation before plans for permanent observatories can be made:

- Seafloor and subseafloor noise: Measurements of inertial noise at intermediate frequencies (10 mHz) are very limited and at low frequencies (3-10 mHz) do not exist. Although knowledge of deep ocean noise sources and propagation mechanisms has increased substantially in recent years, insufficient understanding exists to guide emplacement of permanent observatories. A key parameter that remains unknown is the depth of sensor burial required (in various tectonic settings) to optimize the signal-to-noise ratio while minimizing required drilling penetration. Pilot experiments are required to study these issues.

- Islands and Seafloor Stations: Island seismic stations play an important role in the global seismograph network and are at present the only locations where permanent observatories may exist in the oceans. But how adequate are these stations? Would ocean bottom observatories provide substantial improvements in broad-band signal-to-noise? How does local structure influence seismic signals received on islands compared with an ocean floor site? Pilot experiments are needed to make these comparisons.

- Short-Term Technical Issues: 1 Hz geophones are the lowest frequency sensors routinely used on the ocean floor and they have little sensitivity to earth noise below 50 mHz. An urgent priority is to adapt a presently available broad-band sensor for operation on the ocean floor. One year recordings will be necessary during pilot experiments and although systems with the data storage capacity and timing accuracy necessary for this are currently under development, they have never been deployed.

- Long-Term Technical Issues: Telemetry, Power, Sensors: The major problems here are related to how a permanent global ocean floor network would be operated. With a data rate of approximately 50 MBytes per day, the problems of both internal recording (with periodic data retrieval) or real-time telemetry are extremely challenging. Costs associated with use of fiber optic or existing telecommunication cables can be huge but completely remote packages produce the problem of power source. Completely new (micropower) sensors may need to be developed.

These issues require three classes of pilot experiments.

- 'Island' experiments: to make comparisons of seafloor and downhole sensors (>100 km from shore) with high quality island sites.

- 'Borehole' experiments: to investigate influence of depth of burial upon signal and noise.

- 'Telemetry' experiments: for development and proving of new sensor designs, telemetry schemes, power sources and long-term deployment capabilities.

All these experiments must be carried out at more than one site so that environmental effects can be taken into account. Wireline reentry capability is essential if these experiments are to be performed efficiently.

RECOMMENDATIONS

We plan to establish a global network of approximately 15-20 permanent broad-band ocean floor seismic observatories. A solid base of scientific justification exists to support this ambitious goal. However, before it is realizable, a number of preliminary steps must be taken.

1. The single most important recommendation of the workshop is that during the next 2-3 years a number of pilot experiments should be carried out that will provide the data to:

    - understand sources, propagation mechanisms and environmental controls on ocean floor noise in the band 3 mHz to 50 Hz,

    - determine dependence of noise spectra upon depth of burial of sensor beneath the ocean floor in a range of tectonic regimes,

    - compare signal and noise data from seafloor and subseafloor broadband sensors with that from nearby island sites,

- prove the operational reliability of sensors, data recording and/or telemetry schemes, power sources and timing systems during long-term (>1 year) deployments.

2. These pilot experiments should be located so that they will provide useful structural and tectonic data as well as address environmental noise and technical issues. Specific suggestions for such sites are:

   - Nazca-Pacific-Cocos triple junction,
   - Outer rise seaward of a trench,
   - On the Hawaiian swell,
   - South America-Antarctic-Africa triple junction.

3. Because of logistic considerations, and the opportunity for comparisons with island site data, we specifically recommend that the first pilot experiments be carried out on the Hawaiian swell.

4. A steering group should be formed, co-sponsored by JOI and IRIS, with the following mandate:

   - to stimulate and coordinate efforts by groups of investigators to carry out the necessary pilot experiments to permit the informed design of an effective global network

   - to begin a process of long term planning for the emplacement of this network that includes substantial involvement of the international community

   - to ensure effective communication between IRIS, USGS and the global seismology community, and JOI-USSAC, JOIDES and the marine sciences community (with specific concern for the new global initiatives e.g., RIDGE)

   - to work with funding agencies to plan realistic goals and timetables for progress that can be supported by the available resources

   - to serve as a focal point for information concerning progress of non-U.S. initiatives in deep ocean seismic observatories

   - to provide periodic reports for the community and for the funding agencies that describe progress in these endeavors and better define the emerging needs for resources and technology.

   The membership of this steering group should be endorsed by JOI and by IRIS and should consist of approximately five members of the U.S. academic community.

5. Development of a routine wireline re-entry capability is critical to being able to carry out these pilot experiments efficiently.

6. Broadband seismic stations should be established on a greater cross-section of available island sites in order to begin the process of reducing severe spatial aliasing problems and to learn more about the ocean environment.

7. Although the primary function of the planned observatories is to record seismic data, consideration must be given to the measurement of other parameters (e.g., strain, magnetic field) that will also enhance our understanding of earth structure.

INTRODUCTION

The workshop was held at Woods Hole Oceanographic Institution, Massachusetts on April 26-28th, 1988. The agenda is included in Appendix 1 and the list of participants in Appendix 2. This report consists of two sections in addition to the appendices. The first presents the results of the workshop discussions in the form of reports from the two sub-groups charged with establishing the scientific requirements for a global network of broad-band downhole seismometers, and defining the technical requirements and pilot experiments necessary before such an effort is feasible from an engineering point of view. This section ends with a set of specific recommendations arising from the concluding plenary session of the workshop. The second section consists of six sets of papers that were presented at the workshop verbally or as posters, that in turn are concerned with Scientific Goals, Existing and Planned Global and Regional Networks, Noise, Marine Downhole Seismic Experiments, Instrumentation, and Other Measurements.

The specific short-term goal of this workshop was to identify the necessary experiments to unequivocally prove the scientific benefit and the technical practicality of a global network of long-term broad-band downhole seismometers. Towards this end the workshop proceeded through four steps: the first was to establish the fundamental importance of the scientific goals. Secondly we examined the overlaps and interactions that such an initiative would have with existing and new initiatives in global and marine seismology. Thirdly we studied the technical challenges in sensor design and emplacement, data recording and telemetry, and power and accurate timing requirements. And finally by consensus we generated a menu of ideas and recommendations that would result in progress towards our eventual goals of a permanent global network of broad-band ocean-floor seismometers.

It was not the intent of the workshop to create immediately a new large program and structure to tackle the objectives outlined here. In the near future progress should be made by a set of modest steps directed towards solving existing technical difficulties with sensor design and emplacement, making high quality measurements of ultra low frequency noise levels in the deep ocean and understanding their dependence upon depth of burial of the sensor beneath the seafloor. We hope that this report will stimulate the programs and proposals needed to solve these problems and acquire these measurements. With these solutions in hand we can then proceed to build the necessary plans for the emplacement of a permanent global ocean network that would, as is clear from the body of this report, have a pivotal impact on our understanding of the deep structure of the earth.

SECTION A

WORKSHOP PROCEEDINGS

1. Review of the Scientific Goals
2. Experiment Design and Technical Issues
3. Recommendations

A.1 REVIEW OF THE SCIENTIFIC GOALS

SCIENTIFIC OBJECTIVES

Scientific objectives that can be addressed with seismological data from long-term ocean floor observatories include six broad subject areas:

- **Global earth structure**
- **Oceanic upper mantle dynamics and lithosphere evolution**
- **Oceanic crustal structure**
- **Earthquake source studies**
- **Tsunami warning and monitoring**
- **Sources of noise**

In the following sections, some of the specific scientific problems in each one of these target topics are discussed. The two accompanying tables show the types of observations and the distribution of stations required. Three types of station distributions are desirable. First is a relatively uniform, global distribution of digital seismograph stations, i.e., completing the network of existing and planned continental and island stations by filling in large gaps in coverage with ocean floor instruments. 15 to 20 stations would provide practical completion of this coverage. For some applications either linear or two-dimensional array of stations is needed. Finally, some progress can be made with individual stations or pairs of stations strategically employed.

To achieve the scientific objectives, ocean floor observatories are essential. They are uniquely needed:

- **To provide uniform global coverage in areas without islands**
- **For regional studies of individual tectonic features**
- **To sample wave propagation in 'normal' seafloor**

Oceanic islands are by definition located on anomalous structures. The crust is known to be anomalously thick, and in many cases the mantle structure will also be anomalous. Waveforms of body waves have been shown to be anomalously complex at oceanic island stations. For these reasons, some stations in very simple structure of normal seafloor are important.

1. GLOBAL EARTH STRUCTURE: CORE AND LOWER MANTLE

In this section no distinction is made between the improvement in resolution of the data set that could be achieved by deployment of land based (which includes oceanic islands) stations and ocean bottom stations.

INNER CORE

Lateral heterogeneity and anisotropy of the inner core have been proposed in papers published in the last two years. These inferences were based on travel time anomalies of PKIKP and splitting of spectral peaks of PKIKP-equivalent modes.

The anomalous effect of the inner core structure is more prominent for rays that travel close to the center of the Earth (170°-180° of epicentral distance). Analysis of 20 years of data contained in the Bulletin of the International Seismological Centre (ISC) reveals major deficiencies in coverage caused by the lack of observatories in the ocean basins. The

TABLE 1

GLOBAL EARTH STRUCTURE: REQUIRED OBSERVATIONS

Aims are to map velocity and attenuation variations that reveal lateral heterogeneities related to convection and chemical inhomogeneity.

Scientific question	Observations	Stations
Core Inner core anisotropy or heterogeneity?	PKP, PcP	Uniform, Antipodal
Low velocity layer top of core?	SKS, SKKS	Uniform
Geomagnetic secular variations		Uniform
D" - CMB Boundary layer at base of mantle? Vertical dimensions and spatial variations.	ScS, ScP, Pdiff scattering	Uniform
Correlation of hotspots with slow	ScS, PcP	Uniform
Lower Mantle Slab penetration and geometry (fossil slabs?)	P,S waveforms travel times	Dense in limited area
Anisotropy in lower mantle?	P,S	Uniform
Vp/Vs Ratio Phase changes vs. compositional variations vs. partial melting	P,S, Free Oscillations	Uniform
Transition Zone Undulations in 400 and 670 km discontinuities	P,S waveforms SS, SSS	Uniform Arrays at specific distances
Continent/ocean difference?		

deficiencies at other epicentral distances are less simple to demonstrate, but the fact is that it is impossible at the present time to distinguish between the anisotropy and lateral heterogeneity beyond the C_2 term. Even at that level, some research groups made suggestions that the anisotropic effects may originate in the lowermost outer core.

OUTER CORE

It is generally expected that this region of the earth is chemically homogeneous, because of efficient mixing caused by the vigorous convection. A thin layer just below the core-mantle boundary (CMB) might be an exception. The presence of such a layer, characterized by significantly higher viscosity, could explain the difference between the large CMB topography inferred from seismological studies and its smoothness predicted by observations of changes in the length of day and Chandler wobble. Existence of such a layer has also been proposed to explain certain features in secular variation of the geomagnetic field downward continued to the CMB.

The data best suited to address this problem are SmKS phases: the shear waves, converted to P at the CMB and then multiply reflected on the underside. These waves can bottom just below the CMB and, therefore, be very sensitive to the properties of this region.

TOPOGRAPHY OF THE CORE-MANTLE BOUNDARY

Other than periods of nutations inferred from the VLBI measurements, the CMB topography has been studied using travel time residuals of seismic phases either transmitted through the CMB (PKP) or reflected from it (PcP). Rather large (±6 km) relief has been obtained even in the spherical harmonic expansion truncated at degree four.

The distribution of stations needed to determine the CMB topography with, roughly, uniform resolution is at present highly unsatisfactory. In particular, there are very few reflection points of PcP under the eastern Pacific. Ocean bottom seismographic stations are needed to improve the coverage. In addition to the PcP, other phases such as ScS, often difficult to identify clearly in analog records, could be used to study the CMB topography.

THE REGION D"

The anomalous properties of this region, spanning the deepest 200 km or so of the mantle have been identified nearly 50 years ago. It has been demonstrated in the mid-1970's that the rms level of lateral velocity variations is several times higher in D" than in the middle mantle. The global mapping with uniform resolution is now impossible, because of the paucity of rays bottoming in this depth range in the southern hemisphere. The high level of velocity perturbations has been attributed to both thermal and compositional anomalies. There is some evidence of thermal coupling between the lowermost mantle and the liquid core. Also, the hot spots - inferred from geological and geophysical evidence gathered at the earth's surface - are preferentially located over the anomalously slow (hot?) regions in the lowermost mantle. Some of the largest geochemical anomalies (so called Dupal anomaly) occur over a band centered on, roughly 20°S parallel, which also corresponds to the largest negative velocity anomalies in the lowermost mantle.

More uniform sampling of the D" with seismic data is needed to resolve the laterally varying properties of this region, which - many agree - may be most important to our understanding of the global dynamics. Some progress in mapping of D" could be achieved with temporary deployment of OBS arrays, where active source regions are available, but permanent ocean bottom stations may be needed to observe anomalies along paths originating from regions of low, diffuse seismicity.

LOWER MANTLE

The most striking feature of tomographic images of the lower mantle is the ring of high velocities circumscribing the Pacific, and the slow 'core' at its center. Some interpret the high velocity anomalies as the evidence of slab penetration in the lower mantle, others think that this pattern is an expression of very large scale convection - dominated by degree 2 - in the lower mantle, which may be actually determining the largest scale features in the pattern of plate motions. Indeed, there is good correlation for degrees 2 and 3 between the observed plate velocities (poloidal part) and that predicted by the dynamic response calculated from tomographic results.

There is good agreement among various models at the gravest orders of the expansion: the features are large and easier to resolve. In order to determine the nature of coupling between the upper and lower mantle, it is necessary to resolve higher order features. In particular, the interior of the Pacific needs to be better monitored with seismographic stations. Much of the volume of the lower mantle in this region can only be sampled using relatively rare intraplate events.

2. OCEANIC UPPER MANTLE DYNAMICS AND LITHOSPHERE EVOLUTION

The goals of research in this area are to map the thermal structure, the distribution of large-scale chemical inhomogeneities, and the pattern of flow in the upper mantle beneath the oceans. We know the overall pattern of increasing lithospheric thickness with age of the seafloor from seismic surface wave measurements. To address questions such as the thinning of the lithosphere beneath hot spot swells, the distribution of heterogeneities, and the pattern of convective flow, we need the increased horizontal and vertical resolution that can be provided by improving the uniformity of distribution of stations within the global seismic network and by deploying arrays of stations.

SEISMIC ANISOTROPY AND FLOW

One of the most exciting scientific prospects is the possibility of directly mapping the direction of flow within the mantle. Mechanisms of shear deformation at temperatures and pressures characteristic of the upper mantle all generate preferential alignment of crystals. Since crystals, particularly olivine, are anisotropic, measurement of the directional dependence of seismic velocity provides a direct indication of the preferred orientation of crystals, and, hence, the direction of shearing flow within the mantle. Anisotropy has been detected in the oceanic mantle with seismic refraction experiments, surface wave dispersion, and in the delay times of teleseismic S waves, but the depth extent of anisotropic material is poorly constrained. The horizontal resolution of azimuthal anisotropy is very poor at present, requiring averaging areas several thousand km wide.

THINNING OF THE LITHOSPHERE NEAR HOT SPOTS

Broad swells in the seafloor surrounding hot spots like Hawaii and Cape Verde are thought to be caused by the uplift associated with thermal expansion and thinning of the oceanic lithosphere. Heat flow, gravity and geoid observations provide some constraint on the depth of the thermal anomaly, but there is very little direct constraint on the depth of the anomalous structure. Surface wave measurements and body wave tomography using

TABLE 2

OCEANIC UPPER MANTLE DYNAMICS AND LITHOSPHERE EVOLUTION: REQUIRED OBSERVATIONS

Scientific Questions	Observations	Stations
Can we map variations in flow from seismic anisotropy, with adequate vertical and horizontal resolution? What are the directions of flow in mantle? To what depth is anisotropy frozen in lithosphere?	S splitting, Azimuthal anis. in vel. of body and surface waves. Polarization anisotropy. Particle motions.	Uniform 2-D array
Degree of lithospheric thinning beneath hot spot swells?	Amplitude variations & spectral ratios	2-D array
Mechanism of attenuation, particularly beneath ridges? Partial melt involved?	P,S delays	Ridge & off-ridge
Spatial variations in depth extent of anomalous structure beneath ridges.	Higher mode surface waves for depth resolution.	Uniform, ridge
Are long-wavelength velocity anomalies correlated with long-wavelength geochemical variations?	Lateral variations in dispersion.	Uniform
What is the form of small-scale convection beneath the plates?	Multiple S waveforms and travel times	Uniform, 2-D array
Do oceanic plateaus have roots like continents?		1-D array

teleseismic events can provide the needed data. Tomography experiments have already been successful in revealing deep structure beneath the Yellowstone hot spot.

MECHANISMS OF ATTENUATION BENEATH MID-OCEAN RIDGES

Mid-ocean ridge spreading centers attenuate S_n and surface waves. This is expected since the cold, high Q lithospheric waveguide is interrupted by the upwelling of hot material beneath the ridge. However, neither the geometry of the highly attenuating region nor the physical mechanism is known. Local measurements of spectral ratios of P and S waves and the frequency dependent attenuation of surface waves can indicate whether partial melting is involved.

VARIATIONS IN THE VELOCITY STRUCTURE OF RIDGES

Low resolution, global tomographic experiments indicate that in some areas, there seem to be low velocity roots extending to a depth of several hundred km beneath mid-ocean ridges, while other spreading centers apparently have roots extending to only about 100 km. This may be related to the pattern of whole mantle convection and the depth extent of upwelling beneath different mid-ocean ridges. Long-wavelength geochemical anomalies have been related to the depth of the axis of spreading centers on a global scale. Is there a similar relationship of shear wave velocity to depth or composition? Preliminary, regional studies are encouraging, with low velocities found beneath the shallow Reykjanes ridge and high velocities reported beneath the deep Australian-Antarctic discordant zone.

FORM OF SMALL-SCALE CONVECTION BENEATH THE PLATES

Geoid, gravity and topographic anomalies give some indication of the platform of small-scale convection beneath the plates. Particularly intriguing are the linear anomalies aligned in the direction of absolute plate motion in the Pacific and Indian oceans, the predicted direction of alignment of convective 'rolls'. These measurements, however, sample primarily the shallowest part of the convective system and yield very little control on the depth extent of convection. Rolls beneath the Central Indian Basin were first suggested on the basis of the delays of multiple S waves bouncing under the area, but again provide little vertical constraint. Seismic velocity studies with local stations may provide more control on the depth extent of convection.

DEEP STRUCTURE OF OCEANIC PLATEAUS

Oceanic plateaus are thought to be one of the basic terranes that are accreted to the edges of continents. Very little is known of either their crustal structure or the deeper mantle roots of these anomalously shallow regions. Do they have deep roots of depleted mantle material? Surface wave dispersion and other seismic techniques could be employed in an initial exploration of their structure.

3. EARTHQUAKE SOURCE STUDIES

Earthquake source studies require global coverage of the focal sphere. Seismic stations must be well distributed with respect to both epicentral distance and azimuth around the earthquake source. Since earthquakes can occur anywhere, it is important that seismic stations be uniformly distributed over the earth. For many regions, this can be accomplished with continental and island stations. However, there are many expanses of ocean that would need an ocean-floor seismic station. In addition to the fundamental goal of uniform global coverage, there are certain types of studies where ocean-floor stations would play a key role. One obvious case is a study of local or regional phenomena in oceanic area which cannot be accomplished by land stations at all. Another case concerns the events which cannot be explained by conventional models and require a detailed study using near field data. Recent studies have shown that some events exhibit very unusual seismic radiation suggesting that they are not ordinary earthquakes. In these problems, far-field data from land-stations do not have sufficient resolution to determine them.

The following list, though incomplete, further illustrates the key role that ocean-floor seismic stations would play in earthquake source studies:

LOCATION, FOCAL MECHANISM AND RUPTURE PROCESS DETERMINATIONS

Installation of ocean-floor stations would improve the accuracy of the epicentral parameters, particularly the depth. This is of critical importance for several problems in seismology and geophysics, e.g., depth of the seismic coupled zone, depth extent of outer-rise events and rheology of the oceanic lithosphere. Detailed studies of the focal mechanism based on wave-form inversion desperately require more stations in the ocean, the Pacific Ocean in particular. Due to the fact that the teleseismic distance range window falls completely within the Pacific Ocean over a wide azimuthal range for many subduction-zone earthquakes, detailed studies of fault geometry, slip direction, and the spatial variation of slip magnitude on the fault plane are not possible.

STUDY OF UNUSUAL EVENTS

Although most earthquakes are caused by faulting, recent studies suggest that some events which are generally thought to be ordinary tectonic earthquakes may be caused by some other processes like large-scale slumping and magmatic injection. Whether the earthquake is an ordinary earthquake or a large-scale submarine landslide has an important implication for estimation of long-term seismic hazard (the repeat times of a tectonic earthquake or a slump are obviously controlled by different processes). The difficulty in the determination of the mechanism of the unusual events is that far field data are not adequate to resolve the details of these events. Addition of near-field data, in particular, ocean-floor recordings, would greatly improve the resolution.

INTRAPLATE EARTHQUAKES

These relatively rare events, that occur away from plate boundaries are important for two reasons. First, they provide us with information on the state of stress within the plate; information important for our understanding of the plate dynamics, the driving forces as well as the intraplate deformation. Second, they excite seismic waves that allow us to sample regions that could otherwise be studied only through active seismic experiments (artificial sources).

The presence of the ocean bottom observatories would significantly lower the threshold of detection of intraplate earthquakes and improve the accuracy of their location, which is important in using the data generated by these sources in studies of lateral heterogeneity.

To conclude, there are several important research frontiers in earthquake studies, e.g., detailed locations in subduction zones, rupture process studies, detailed studies of faulty geometry, 'exotic' events, and tsunami research, that either require or would greatly benefit from permanent ocean-floor seismic stations.

4. TSUNAMI WARNING AND MONITORING

Tsunamis are generated throughout the rises of the Pacific Basin, South America, Alaska, and the northwest Pacific, and have historically been the source of destructive Pacific-wide tsunamis. When considering the directionality of propagation of tsunami energy, tsunamis generated in the Aleutians or along the Kamchatka coast pose a particular problem as the islands of the Hawaiian chain provide the first capability for land-based measurements.

An ocean bottom seismological observatory, combined with deep ocean pressure gauges, located seaward of the Aleutian Trench or the Kamchatka Trench would provide invaluable data for the determination of the source mechanism of a tsunamigenic earthquake as well as monitoring of the passage of a tsunami in its natural environment, the deep ocean. As the occurrence of tsunamigenic earthquake is an infrequent event, data must be collected continuously over a period of years. If seismological and tsunami data were available on a real-time or nearly real-time basis, they would be operationally useful in greatly enhancing the tsunami warning services provided by the Pacific Tsunami Warning Centre to national and international participants of the Tsunami Warning System in the Pacific. Even if not available in real-time, the data would provide significant research support in understanding the mechanisms of tsunamigenic earthquakes, particularly those of low magnitude, and the characteristics of tsunami propagation. The expected result would be an improvement in tsunami warning services both in timeliness and in development of a predictive coastal evaluation capability.

5. OCEANIC CRUSTAL STRUCTURE

Almost without exception permanent downhole seismic stations would not be emplaced with the primary objective of studying crustal structure. However, many important new and relevant observations would be possible as a by-product of the other primary studies described here. Useful auxiliary programs could be designed to tackle a range of fundamental problems, and in special cases temporary arrays may be the only feasible approach to a number of particularly intractable problems.

Studies of anisotropy in the shallowmost crust provide a means of estimating fracture orientations below the seafloor and thus can constrain models of the emplacement and tectonic evolution of the upper crust. Anisotropy in the lower crust (if it exists) could be related to crystal orientation and thus to magma chamber processes and the form of crustal accretion. Observations of shear wave splitting and particle motions that are key to robust determinations of anisotropy are only possible with well-coupled (downhole) sensors. Good measurements of Q in oceanic crust do not exist, but their possible relation with fluid content could permit mapping of the depth extent of hydrothermal circulation or the shape of magma melt zones. The primary knowledge of oceanic crustal structure comes from compressional wave studies: lack of knowledge of shear wave velocities, especially in the mid-ocean ridge environment, results in substantial ambiguities in interpretations of Layer 3 structure and the anomalously low P-wave velocities generally associated with the crestal regions. Earthquakes are the only effective high-energy shear wave source available to the experimental seismologist in the deep-ocean, and thus their monitoring with well-coupled broad-band sensors provides a rare method of obtaining high-quality shear-wave information. Temporary arrays of sensors could be applied to studies of particular classes of structures, e.g., oceanic plateaus or continental margins.

6. SOURCES OF NOISE

A more thorough treatment of noise generation mechanisms is given in section A2 of this report. Suffice it to say here that so few measurements of inertial noise have been made in the deep ocean at frequencies below a few hertz (and none below 10 mHz) that this is a fertile field of investigation. Active mid-ocean ridge processes (volcanic tremor, hydrothermal venting) may generate sufficiently distinctive signals that permanent seismic stations may provide a means by which the short time scale variability in the accretion process could be monitored. Monitoring the crustal response to long-period pressure variations provides a new approach to the determination of the physical properties of the crust and lithosphere. A thorough understanding of microseism propagation mechanisms awaits more deep ocean observations.

PILOT PROJECTS

Significant scientific returns can be expected from the deployment of single stations if the location is carefully selected. In making these selections, we are concerned only with geographic position, not with the logistic considerations of whether the station should be on the ocean-bottom or down a borehole, the mode of data transmission, and problems of servicing and emplacing the station. To some extent we have also ignored the problems of expected noise levels, although quiet sites are clearly preferred. Important advances can be expected from single, broad-band, long-term stations deployed at the following sites:

Nazca-Pacific-Cocos ridge-ridge-ridge triple junction. A station at this triple junction could sample upper mantle structure along three spreading centers ranging from intermediate to very fast spreading rates using surface waves and body waves. It would fill a significant gap in coverage of the global network. There is an excellent distribution of teleseismic events surrounding the station to act as sources at all azimuths. It would be expected to be a quiet site, with significant sediment thickness in quite young seafloor beneath the equatorial upwelling zone.

Outer Rise seaward of a trench. A station on the outer rise could be used to study the source characteristics of outer rise and trench earthquakes. It would be useful for determining structure of the downgoing slab with body waves. It is the ideal location for tsunami monitoring and warning. Depending on the location selected, it could fill a gap in global coverage. Candidate sites include seaward of the Japan trench to tie in with the Poseidon network, in the Aleutian-Kamchatka corner to monitor tsunamigenic events and fill in a gap in global coverage, and seaward of the Aleutian trench for tsunami studies and to monitor seismic gaps.

On the Hawaiian swell. A station on the crest of the swell sufficiently far from the islands could serve as an excellent site on 'normal' seafloor, isolated from rapidly varying structures that perturb the seismic waveforms. This site could be used to study lithospheric thinning beneath one of the best-developed, most-studied swells. If located within a few hundred km of the big island of Hawaii, it could be used to help constrain the location and mechanism of earthquakes in the vicinity of this seismically active island. There currently remains debate about the mechanisms of both the 1973 and 1975 Hawaiian events.

Near the South America-Antarctic-Africa triple junction. Although likely to be a noisy site, this would fill one of the largest gaps in global coverage of seismic stations. In addition, it could be used to study the upper mantle structure beneath mid-ocean ridges with very slow to intermediate spreading rates.

A.2 TECHNICAL CHALLENGES AND PILOT EXPERIMENTS

1. CRITICAL PROBLEMS

1.1 SEAFLOOR AND SUBSEAFLOOR NOISE

The most critical issues which must be addressed in the early stages of this program are the level and degree of variability of seafloor noise within a broad frequency band (3 mHz-50 Hz). Although our understanding of seafloor noise has increased enormously through the past several years, the number of measurements of inertial noise at intermediate frequencies (10 mHz-100 mHz) are very limited and at low frequencies (3-10 mHz) are nonexistent. The pilot experiments which are proposed for this program require that the highest priority be given to increasing the number and variety of these measurements.

1.1.1 Possible Noise sources

A discussion of the sources of seafloor noise is most conveniently broken into four frequency bands since the noise is dominated by different physics in each of these bands. The first, and most familiar band, is from 3 to 50 Hz. This band is known as the *Very Low Frequency (VLF)* or *Infrasonic* band by scientists with interests in acoustics but is termed *High Frequency Noise* for our purposes. The next lower band, from 80 mHz to 3 Hz, is commonly called the *Microseism Band* after the high level, relatively narrow bandwidth microseismic noise which is clearly recorded at all sites on the Earth's surface. The third band, the *Noise Notch* (20 mHz–80 mHz) has a variable bandwidth and is observed on both the continents and in the ocean. The final band, in which limited measurements of pressure have been made *[e.g. Filloux, J.H., Pressure fluctuations on the open ocean floor over a broad frequency range: New program and early results, J. Phys. Ocean., 10, 1959-1971, 1980]*, extends from DC to 20 mHz. With the advent of suitable sensors and accompanying measurements, this broad band will almost certainly be broken in additional bands (e.g. tides).

Figure 1 illustrates three seafloor pressure power spectra spanning the band from 0.7 μHz to 20 Hz. The solid and coarsely-dashed curves represent data collected in the Pacific while the data for the lower power, finely-dashed curve were collected in the Atlantic. Power spectral levels in the lowest band, from 0.7 μHz to 20 mHz, decrease monotonically at 25 dB/decade except in the band from 10 to 200 μHz where several tidal lines are recorded. The microseism band in both spectra is dominated by the double frequency microseism peaks above 200 mHz. Between the microseism peaks and 2 Hz, power spectral levels decrease at 40 dB/decade.

In Figure 2 the data in Figure 1 have been converted to acceleration units and superimposed on curves which illustrate the accelerations expected at an epicentral range of 30° from earthquakes spanning a range of magnitudes. The conversion from pressure to amplitude was made using a simple acoustic plane wave relationship assuming vertical incidence of seismic energy and continuity of vertical traction and displacement. Although this approximation is reasonably good at high frequencies, the conversion simply ensures that the units are correct at lower frequencies. Direct inertial measurements are simply not available at the lower frequencies. This figure is very encouraging and indicates that at this range all events with an M_w of 5.5 or greater lead to motions above the expected noise levels. Such earthquakes or explosions should be detectable with seafloor sensors. With regard to hydrophones deployed in the Atlantic, where ambient noise levels appear to be somewhat lower, earthquakes with M_w as low as 5.0 should be detectable.

1.1.1.1 High frequency Noise

Vertical component noise is comparable to continental sites in this frequency range. The best documented mechanism for the generation of ambient noise in the high frequency band (3-50 Hz) is shipping. Shipping produces noise that becomes trapped in the ocean waveguide either by reflections from continental margins or by interactions with the shoaling SOFAR channel at high latitudes. This noise is confined primarily to horizontally propagating modes although nearby shipping can produce very high noise levels which propagate directly to the seafloor. In the absence of shipping or for vertically directional arrays or sensors, high sea/wind states can produce measurable noise at the seafloor. The noise source, in this case, is felt to be breaking waves primarily because the noise levels increase at about 14 knots, a wind speed just sufficient to cause white caps. The noise levels increase with increasing wind speed beyond this level. Given that most of these data have been collected with hydrophones, comparatively little is known regarding the effects on inertial components. It is likely that this wind speed component will affect the vertical direction of motion the greatest since the directivity is largely vertical. In addition to these sources of noise, biological noise can be substantial, particularly that associated with whales, and vortex shedding from the instrument can produce noise in high current environments.

1.1.1.2 The Microseism Band

Seafloor vertical noise levels in this frequency band are generally somewhat larger than those observed on land presumably because the oceans provide the source for microseismic noise. The classical microseism band (80 mHz-3 Hz) is fairly well understood in the marine environment. Nonlinear interactions between surface gravity wavetrains in opposing swell are the primary sources of this noise. Opposing swell can be generated by spatially distributed storm systems, reflections from continental and island shores or by single storms. Winds which are responsible for exciting wind-driven swell are capable of generating waves which propagate in all directions, not just in the direction of the mean wind at the sea surface. At the lower frequencies, the so-called single frequency microseisms are generated by the interactions of swell with the continental shelf and shore break and are most prominent at ocean sites with well-developed shallow margins. Observations of the single frequency microseism are common on continents, but are rare for small volcanic islands.

Microseisms propagate largely as Rayleigh and body waves with high phase velocities (>1.5 km/s) in the ocean. At higher frequencies it is possible that energy scattered from these higher phase velocities can give rise to trapped interface or Stoneley waves. There is evidence that the noise field at the higher frequencies increases suddenly with wind direction changes, presumably because of the resultant richer wavenumber content of the surface gravity waves. The precise mechanism and frequency band of noise generation for these oblique winds and their effect on swell is unknown. Secondly, few direct observations of the seafloor wavefield have been made at sea in conjunction with surface environmental measurements. Additional simultaneous measurements of these phenomena would be very helpful in understanding this very important source of noise.

1.1.1.3 The *Noise Notch*

Very low noise levels have been measured in the upper intermediate band (20 mHz-100 mHz) and are generally consistent with similar levels and phenomena on continents. The bandwidth of the noise *notch* is controlled by the fall-off of the microseism peak at high frequencies and the increase of noise levels at lower frequencies which is associated with shallow water waves (see below). Noise levels within this notch appear to be controlled largely by currents and turbulence in the seafloor boundary layer and these noise levels have been observed to increase with current speeds at energetic sites in the Atlantic. The source of noise in this band is more problematic in the eastern Pacific where currents are less than 10 mm/s. Pressure fluctuations in a turbulent current field give rise to tilts in the seafloor and can be expected to be particularly problematic for horizontal inertial measurements in this frequency band. This noise field is analogous to the well-known meteorological phenomena which control horizontal instrument noise on the continents. In addition to this noise source, seafloor currents will also directly perturb the horizontal components of any seafloor seismometer package. While shallow burial of a sensor package will eliminate any direct action of the currents, substantial burial depths (100 m) will be required to eliminate the, probably larger, tilt field. Atmospheric turbulence, largely responsible for seismic noise on continents at these frequencies, can also contribute to seafloor noise. However, the influence will be necessarily small since pressure disturbances will decay exponentially away from the surface given that atmospheric acoustic velocities cannot directly couple into the oceanic water column.

1.1.1.4 Low Frequencies

The one oceanic site which provided data below 50 mHz (the Columbia OBS off Cape Mendocino) was much noisier than a noisy continental site (up to 30 db) and the horizontal components were as much as 30 db noisier than this. Pressure and velocity fluctuations associated with swell or gravity waves decay exponentially with depth with an e-folding distance which is inversely proportional to frequency squared. Although the depth of the ocean basins is sufficiently great that surface waves do not generate noise at relatively high frequencies, except for the single frequency microseisms, the ocean basin depths can be easily traversed by this exponential decay in the lower intermediate band above 10 mHz. Furthermore, as water depths decrease the shallow water waves can cause the noise notch to disappear entirely. Below 10 mHz, our knowledge of noise mechanisms and spectra is limited. The noise is hydrodynamic in origin and does not propagate as acoustic or elastic energy. Although the very long wavelength swell responsible for noise below 10 mHz has never been directly measured, the swell amplitude must be less than 10 mm. All measurements in this frequency band have been made with differential pressure sensors. Again, the pressure fields are likely to cause significant tilts in the seafloor and these tilts may control the level of horizontal noise. Edge waves, which are strongly trapped along coastlines, have large amplitudes at these frequencies caused by differential heights between packets of wind-driven swell at higher frequencies. Although these waves are strongly evanescent, weak leakage can occur into the deep ocean through scattering along the coastlines.

1.1.2 Directionality of Noise

Throughout the considerable bandwidth of interest, the largest numbers of measurements have been made using pressure sensors including hydrophones and differential pressure gauges. Where seismic or inertial measurements have been made, scientists have concentrated on the vertical component of motion. Horizontal inertial measurements are particularly rare since many instruments are capable of measuring only the vertical component and the horizontal components which have been available have proven to be particularly noisy. Since the rigidity of bottom sediments can be very low (shear velocities less than 50 m/s are common) much of the noise results from tilts and the horizontal components generally exhibit much greater current-induced noise than the vertical, often by 20 db or more in the intermediate and high frequency bands.

1.1.3 Effects of Burial Depth

Considerable efforts are made in continental seismology to couple the seismometer firmly to the earth by literally burying the instrument in the ground or attaching it to a pier in a vault or to the sides of a borehole. In addition to enhanced coupling, this practice removes the instrument from the direct effects of wind and biological encroachment. Clearly similar precautions should be observed on the seafloor although the remoteness of the environment makes this practice extremely difficult. The effects of burial of the seismic sensor were investigated during the Lopez Island Experiment and additional measurements have been made in shallow water environments. During the 1980's seafloor boreholes provided by the Deep Sea Drilling Project (DSDP) and the Ocean Drilling Program (ODP) have been used to investigate the effects of burial on seismic vertical and horizontal noise levels at high frequencies. Additional efforts must be undertaken to investigate the gains which can be achieved by both shallow and deep burial in an effort to increase the resultant signal-to-noise ratios.

Shallow burial of seismic sensors would eliminate the direct effects of current and may produce significant reductions in noise. Deformations of the bottom itself should fall off exponentially with depth at a rate proportional to the horizontal dimensions of the vortices; probably a meter at 1 Hz and 100 m at 10 mHz. However, the steep positive gradient in rigidity expected for most bottom sediments and the underlying basement rocks will increase the rate of attenuation with depth. It would appear prudent to explore thoroughly the benefits of shallow burial, which can be accomplished by an oceanographic vessel capable of piston coring before committing to a much more expensive program of borehole installation with concomitant high costs and fewer stations.

1.2 Island and Seafloor Stations

Island seismic stations play an important role in the global seismographic network. Without ocean bottom instrumentation, islands present the only locations where we can sample the Earth's ground motion in the vast expanse of the oceans. Island siting of seismic stations is motivated by economics and expediency—Economics in the sense that off-the-shelf seismic equipment routinely used on the continents may be used, expediency in the sense of availability of timing and power, relative ease of deployment and maintenance.

Figure 3 shows power spectra of noise data obtained from the IRIS/IDA seismographic stations on Easter Island (RPN), at Eskdalemuir (ESK), 40 km from the Irish Sea and at Piñon Flat (PFO), some 100 km from the California coast. The dashed lines show the level of minimum and maximum ambient noise at quiet continental locations. The high frequency portion of the spectrum for the RPN station was derived from data collected during the most noisy period experienced during a one-year period; it should be taken as representative of the maximum high frequency noise at an island site.

The spectra illustrate two principal features of the ambient seismic noise:

- For periods longer than the microseisms, there is only a small decrease in noise as distances to the coastline increases. Island stations should provide relatively low noise data in this frequency range.

- For frequencies higher than the microseisms, there is considerable reduction in noise as the distance from the coastline increases. Factors such as the coastal shoaling, barrier reef coverage, and surf conditions can be expected to cause large variations in the ambient noise in this frequency range.

Three areas of investigation are necessary to lay a firm scientific foundation for oceanic siting of seismic stations. The first consideration is the broad-band noise characteristics of island and adjacent seafloor sites. A second consideration is the quality of seismic signals received by island and seafloor sites and how the local structure influences and modifies these signals. A third consideration is to what extent the islands can be considered representative–in a seismic sense– of the neighboring region. Certainly the last two areas must be frequency dependent. Lower frequency and longer wavelength signals are likely to be less disturbed by the perturbation represented by the presence of the island. Measurements of the earth's free oscillations are likely to be as fruitful on islands as on the adjacent seafloor.

1.3 Short Term Technical Issues

Although the need for a wide variety of seafloor measurements is very clear, several technical issues must be resolved before these measurements can be made in a reliable manner. Presently, 1 Hz natural period geophones are the lowest frequency sensors which are routinely used in seafloor measurements and they have little sensitivity to earth noise below 50 mHz. Hopefully, presently available broad band seismometers can be adapted for use in the required pilot experiments. The full variability of seafloor measurements must be encompassed in order to understand fully the physical processes which cause seafloor noise. In order to make such measurements deployment times on the order of a year will be necessary. Although low data rate oceanographic instrumentation such as current meters can be deployed for such periods, no seafloor seismometer presently exists with this capability. Finally, timing is a difficult undertaking in an autonomous vehicle such as a seafloor seismograph. Millisecond accuracy over a year's period is probably feasible, but has not yet been accomplished.

1.4 Long Term Technical Issues

The two major issues which must be solved in the long term in order to make permanent seafloor seismic stations a practical reality are telemetry and power supply. Although the major scientific issues such as noise and island-seafloor comparability can be addressed by existing technologies, further progress depends upon progress during the coming decade along several additional lines.

1.4.1 Telemetry

The problem of data retrieval from a long-term, worldwide, broad band, deep ocean seismic array remains to be solved. This problem is aggravated by the high sampling rates and dynamic range that are required. For example, a three-component seismometer plus hydrophone sampled at 50 samples/s/channel with a 24-bit data word gives a data rate of 4800 baud or 52 MBytes/day. Acceptable options for data retrieval fall into two general categories: real time and delayed. Several real time options were discussed at the workshop, although none appear practical at this time. The most technically desirable option is to connect the sensors to land *via* dedicated, undersea (fiber optic?) cables. However, the costs for a suitable cable are probably prohibitive, ranging from $5 to more than $10/meter—tens of millions of dollars for the several thousand kilometers of cable that would be required for even minimal coverage of the oceans. A second option is to connect the sensors to a buoy which then transmits the data via a radio or satellite link. This option is currently unacceptable due to a combination of factors including high cost, low data transmission rates, high power requirements, short range and questionable reliability. A third option is to transmit the data to land via transoceanic communications cables that are being retired from commercial use. Dr. S. Nagumo presented the details of the Japanese POSEIDON Project which proposes to attach sensors to a section of the TPC-1 cable between Japan and Guam that will be retired in 1989. The primary disadvantages of this option are that it offers a severely restricted choice of sensor sites and a limited seismic bandwidth. Furthermore, the cost of establishing four permanent stations on this cable is of the order of $10,000,000.

The prospects for delayed data retrieval are challenging. These systems require some method of remote data storage with subsequent data retrieval by ship or by plane at fairly regular intervals. The most compact format currently available for data storage are helical scan magnetic tape recorders with a capacity of 2.2 GBytes. Although a month's data could be stored on such a medium, some form of data compression and segmentation would be necessary for a year's deployment if the number of recorders was restricted. Additional problems with this option include power supply, unproven technology necessary to get the data to the ocean surface and retrieve it and the high cost of a ship or plane to do the regular retrieval. A detailed description of specific scenarios for most of the data retrieval schemes described above can be found in Table 1.1.

ID	Scenario	Distance to Land	Power	Data Trans.	Real Time	Relative Cost	Estimated Duration	Notes
1A	One or more instr. linked by cable to island	100 km (Few)	Cont. from cable	Cont. by cable to island	Y	V. High	>10 Yr	Special cable deployed
1B	"	No limit	"	"	Y	High	"	Re-use abandoned cable*
2A	Instrument linked by E/M cable to surface buoy	<50 km	From buoy (solar, nuclear)	Cont. VHF to island	Y	High	<10 Yr	Buoy needs "annual" maintenance. Weather problem?
2B	"	No Limit	"	Burst by rented channel in comm. satellite	Y	V. High	"	Buoy needs "annual" maintenance. Direct. buoy antenna req'd.* High data rate to satellite req'd.*
3	Instrument linked by E/M cable to near surface *elevator* buoy	No Limit	Buoy (Batt?)	To ship/ aircraft of opport. in burst mode	N	High/ Moderate	<5 Yr	Buoy ascends and descends on command or at preset times. *Elevator* under development.*
4	Instrument linked by E/M cable to near surface buoy*	No Limit	Buoy Batt. change by ship	Burst mode via cable to ship	N	Moderate	1–2 Yr	Buoy ascends on command. Needs research ship plus regular visits. Ship resupplies buoy.
5	Instrument ejects pop-up package & Argos beacon.*	100 km (Few)	Sea floor batt.	Original seafloor record. medium	N	Low	1-2 Yr	Could use local craft for pick-up any time after package surfaces. Argos fix ±1 km.

Table 1.1: Telemetry scenarios for delivering data from seafloor or subseafloor stations. Asterisks (*) indicate those technologies which require additional development.

1.4.2 Alternative Sensors

A marine borehole geophysical observatory (MBGO) should include modules for measuring seismic displacements, seismic and tectonic strains, tilt, the magnetic and electric field, pressure and temperature. The individual sensors needed all exist in some form, and have been demonstrated to function reliably for long periods of time, decades in most instances, either in marine or terrestrial environments. Work is needed on the packaging of individual sensors, however, so that they can be installed together in a deep sea drill hole.

Strainmeters are an important component of an MBGO. When combined with accelerometers, strainmeters can be used to measure seismic phase velocity and azimuthal anisotropy in the oceanic crust and upper mantle. Deep ocean strainmeters are also needed to study the deformation of the oceanic lithosphere in tectonically active areas. Seismicity and deformation of the oceanic lithosphere associated with subduction in the northwestern Pacific region could be studied better if marine observatories including strainmeters were located east of Japan, for example.

A marine geophysical observatory would necessarily consist of two parts, a downhole sensor package and a control unit on the seafloor to house a power source and data recording and/or transmission equipment. It may be feasible and desirable to service the control unit at intervals, perhaps once per year, but it is unlikely that the downhole sensor package could be serviced and redeployed, for both technical and economic reasons. Including strainmeters in the sensor package mandates that they be cemented into the well bore. Cementing, rather than clamping, the sensor package also solves other potential problems. These include increased noise on the horizontal seismometers due to the presence of a convecting fluid in the well and the possible failure of a mechanical clamp during the lifetime of the observatory. There are no obvious technical difficulties connected with cementing an instrument package into a deep sea drill hole, but a high priority should be given to a preliminary experiment to test this capability.

1.5 Summary

The critical problems which face the establishment of permanent seafloor or subseafloor seismic stations can be summarized in a brief outline:

- **(Sub)seafloor noise**
  - Seafloor noise levels at low frequency
    - + Vertical
    - + Horizontal
    - + Pressure
  - Noise gain as a function of burial depth
    - + Low→high frequencies
    - + Vertical/Horizontal
  - Correlation of noise with currents
    - + Intermediate and high frequencies
  - Dependence of noise upon
    - + Wind
    - + Sea state
    - + Bottom type
    - + Proximity to shore, tectonic features...
    - + Shipping

- **(Sub)seafloor signals**
  - How do signal levels vary with depth of burial and in boreholes?
    - + Low→high frequencies

- **Comparison with island stations**
  - Are seafloor sites quieter than islands?
    - + Low→high frequencies

- **Wireline reentry capability must be developed and demonstrated**

- **Short term technical issues**
  - Can present broadband seismometers be adapted for pilot experiments?
  - Are one year deployments practical?
  - How can accurate timing at independent stations be achieved?

- **Long term technical issues**
  - Should a *permanent* network be serviced by ships?
  - How can accurate timing at independent stations be achieved over the long term?
  - How can real time telemetry be achieved?
    - + Acoustic
    - + Electromechanical cable
    - + Fiber optical vs coaxial telemetry
    - + Deep ocean cables
  - Is real time telemetry required?
  - Can a broadband seismometer be constructed for long term seafloor use?
    - + Micropower
    - + Is mechanical locking adequate?
    - + Small diameter (3 5/8") diameter
  - May be required for bare rock sites if mining technology drills are used.

2. Pilot Experiments

A series of Pilot Experiments should be conducted to address the scientific and engineering issues raised above. Three classes of experiments were recommended:

- Measurements on islands and *nearby* seafloor
  - Experiments should be conducted at more than one site
    - + Volcanic island
    - + Quiescent island
    - + Exploration and long-term deployments must be made at each site
  - Instruments should include:
    - + Shallow, buried sensors
    - + Seafloor sensors
    - + Environmental measurements
  - Site surveys should be conducted to characterize the sediments, crust and upper mantle
  - Sensors should be > 100 km from shore
  - Comparative island sites should be *optimally* sited (boreholes?)
  - Issues addressed include:
    - + Are earthquake signals simpler on the seafloor?
    - + Increased database on low frequency seismic noise
    - + Current effects on noise
    - + Effects on signal and noise of shallow burial

+ Comparison of island and seafloor noise levels

- Borehole Experiments
  - Experiments should be conducted at more than one site
  - Instruments should include:
    + Borehole broadband sensors
    + Seafloor broadband sensors
    + Subseafloor broadband sensors
    + Environmental measurements
  - Site surveys should be conducted to characterize the sediments, crust and upper mantle
  - Issues addressed include:
    + Are mechanical hole locks adequate for subseafloor broadband seismometers?
    + Effects on signal and noise on shallow and deep burial
    + Increased database on low frequency seismic noise
    + Current effects on noise

- Technology Experiments
  - Develop optimal broadband seafloor seismometer
  - Experiment with proposed telemetry methods including seafloor cables
  - Long term, independent timing
  - Seafloor power

2.1 Comparisons with Island Sites

The initial pilot experiments are designed to measure the noise field at long periods on the seafloor and compare seafloor noise and signals to those on a nearby island. Seafloor noise will be measured by instruments on the seafloor and buried ≈ 1m beneath the bottom sediment. Islands, while providing economical station sites, are noisy and may distort seismic waveforms.

Long period noise levels on the ocean bottom are unknown, except for measurements made the Point Arena OBS more than fifteen years ago. These measurements are possibly contaminated by ocean current interactions because of a large attached current meter. If sufficiently low noise levels can be achieved with sensors mounted on the seabed, the expense of borehole installations might be avoided. An understanding of the structure of the spatial noise field may also allow us to predict noise at depth.

The primary objectives of this experiment are:

- Measure broadband seafloor noise and compare it with simultaneous measurements on islands.

- Determine the effects on noise levels of:
  - Shallow sensor burial
  - Ocean current interactions
  - Bottom pressure variations
  - Wind
  - Wave height

Experiments to achieve these goals will be conducted in several different areas, with the ocean bottom instruments 150–200 km away from the island. The choice of sites should include active hotspot islands, such as Hawaii or the Galapagos, and inactive sites. The three to four OBSs will probably utilize broadband sensors such as the Guralp CMG-3 or the Teledyne KS 54000. One of the instruments will be buried at shallow depths beneath the seafloor, with the remainder sitting on the ocean bottom. Broadband differential pressure gauges will also be attached to the OBSs to record the ambient pressure field. Within the array, a current meter will be placed in a position to allow the recorded noise to be correlated with the strength of the bottom currents. The seafloor noise measurements can then be compared with those on the nearby island and later with noise levels observed in ocean bottom boreholes. The deployment period should be at least one month to allow the broadband sensors time to stabilize, after some test deployments of a few days. A full year's recording would be useful in order to sample thoroughly a number of different weather conditions.

2.2 Borehole Seismic Experiments

Recent comparisons of short period and long period ocean bottom seismograph (OBS) observations suggest that considerable signal to noise improvement may be achieved by implanting the sensors beneath the ocean floor to isolate them from noise sources. Unfortunately, no long period data are available to compare the performance of broadband borehole seismometers and OBSs. Accordingly, to facilitate the overall objective of comparing the performance level of island observatories, ocean bottom and borehole seismometers, the primary objective of the pilot phase experiments are to determine unequivocally the broadband (10 mHz-50 Hz) signal to noise ratio of borehole seismometers as a function of observation depth beneath the ocean floor. Observations should be made at several tectonically-contrasting sites (ridges, trenches, oceanic plateaus) in the deep ocean.

Measurements should be made in the unconsolidated oozes and lithified sediments as well as the volcanic and intrusive rock typical of the upper layers of the oceanic crust. Geophysical site survey information should be available to characterize the velocity structure beneath the borehole site. Ideally, simultaneous seismic measurements will be made at the borehole and OBS sites along with environmental observations of barometric pressure, wind, velocity, bottom current velocity and pressure, surface wave height and tides. This will provide correlation of the environmental noise with the seismic observation to allow prediction of likely signal to noise ratio performance for siting future ocean floor seismic observatories.

In order to minimize the potential drilling cost for implanting borehole seismometers, special care should be taken to acquire a series of observations depths down each borehole, especially in the sediment above the hard oceanic basement rock to determine the minimum depth required to increase significantly the broadband signal to noise ratio for various sub-bottom structures and environmental conditions. In fact, recent studies suggest it may be possible to achieve marked signal to noise ratio improvement for OBS-type instruments by simply implanting the sensor a few meters beneath the seawater/sediment interface.

Three possible modes of seismometer deployment are envisioned to attain our proposed pilot program objectives:

1) <u>Through Pipe</u> lowerings using a *slim line* version (3.5" diameter) of conventional broadband borehole seismometers such as the Teledyne-Geotech model 36000 attached to the well logging cable channel of the D/V JOIDES Resolution. These experiments would probably be of short duration with recording limited to the period while

the drill ship is over the borehole. However, using the *pipe stripping* technique, the logging cable can be attached to a tethered seafloor recording package which can be left on the seafloor for longer duration observations (months) as was done with the Hawaii Institute of Geophysics short period borehole seismometer. A further justification for a *slim tool* seismometer will be to allow for emplacement in the small diameter mining-type borehole planned for deep penetration into bare rock at rise axis sites. For the purposes of the pilot experiments, however, the added expense of the drilling ship and the technical challenges associated with the development of a *slim line* tool preclude the use of this approach for any but the high frequency band.

2) Strap-on lowerings on the outside of the ODP drill pipe, as was done in the previous DARPA borehole seismometer experiment aboard the D/V Glomar Challenger, should be considered. In particular, the experiment could use existing large diameter, broadband borehole seismometers. The experiment might be most readily accomplished aboard the D/V JOIDES Resolution using the existing television camera shuttle system which is routinely used as an aid in borehole reentry.

3) Wireline Reentry into existing ODP/DSDP boreholes using conventional research ships rather than a drill ship's pipe may also be available in the near future. For these experiments existing large diameter broadband seismometers could also be implanted and ship expenses could be minimized.

Figure 1. Noise power spectral density in pressure units for three seafloor sensors. The data for the solid and wide dashed curves were collected in the Pacific Ocean, off the coast of Baja California (at 4090 m depth) and off the Gulf of California (at 3210 m depth) respectively. The data for the closely dashed curve were collected in the Atlantic at the Nova Scotia Rise in 5000 m of water.

Figure 2. Acceleration amplitudes expected at an epicentral range of 30° from Earthquakes ranging in Mw from 5.0 to 9.5. The seafloor noise level curves are described in the caption for Figure 1. The amplitudes plotted are 1.25 times the RMS values over 1/3 octave bandwidths.

Figure 3. Noise recordings made on an island (RPN-Easter Island) compared with measurements made on continental sites located near the ocean (ESK-Eskdalemuir; PFO-Piñon Flat Observatory). Note that the high frequency content of the microseism noise decreased as the distance from the shore increases. The amplitude of the microseism peak also decreases in the same sense as the high frequencies.

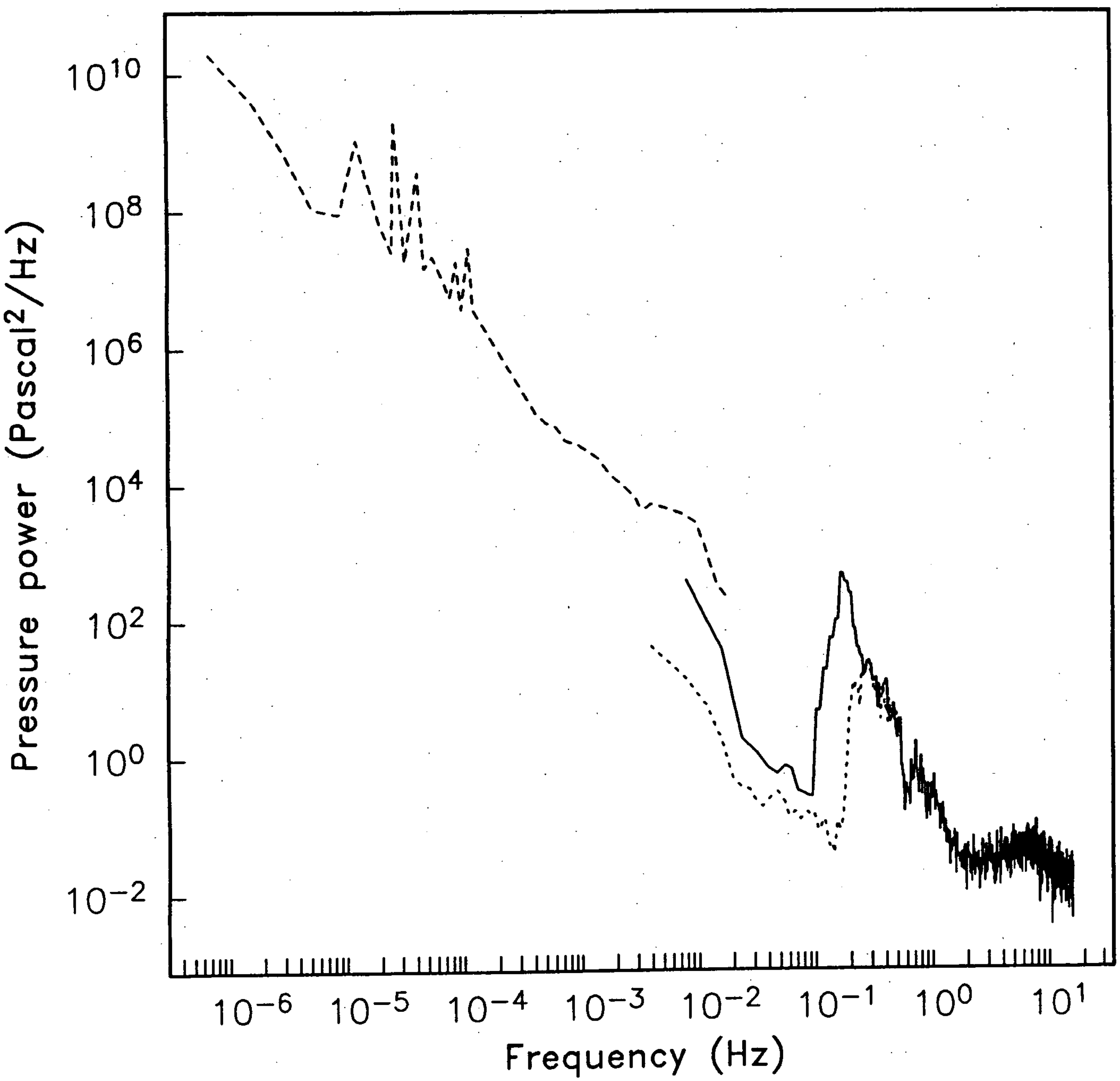
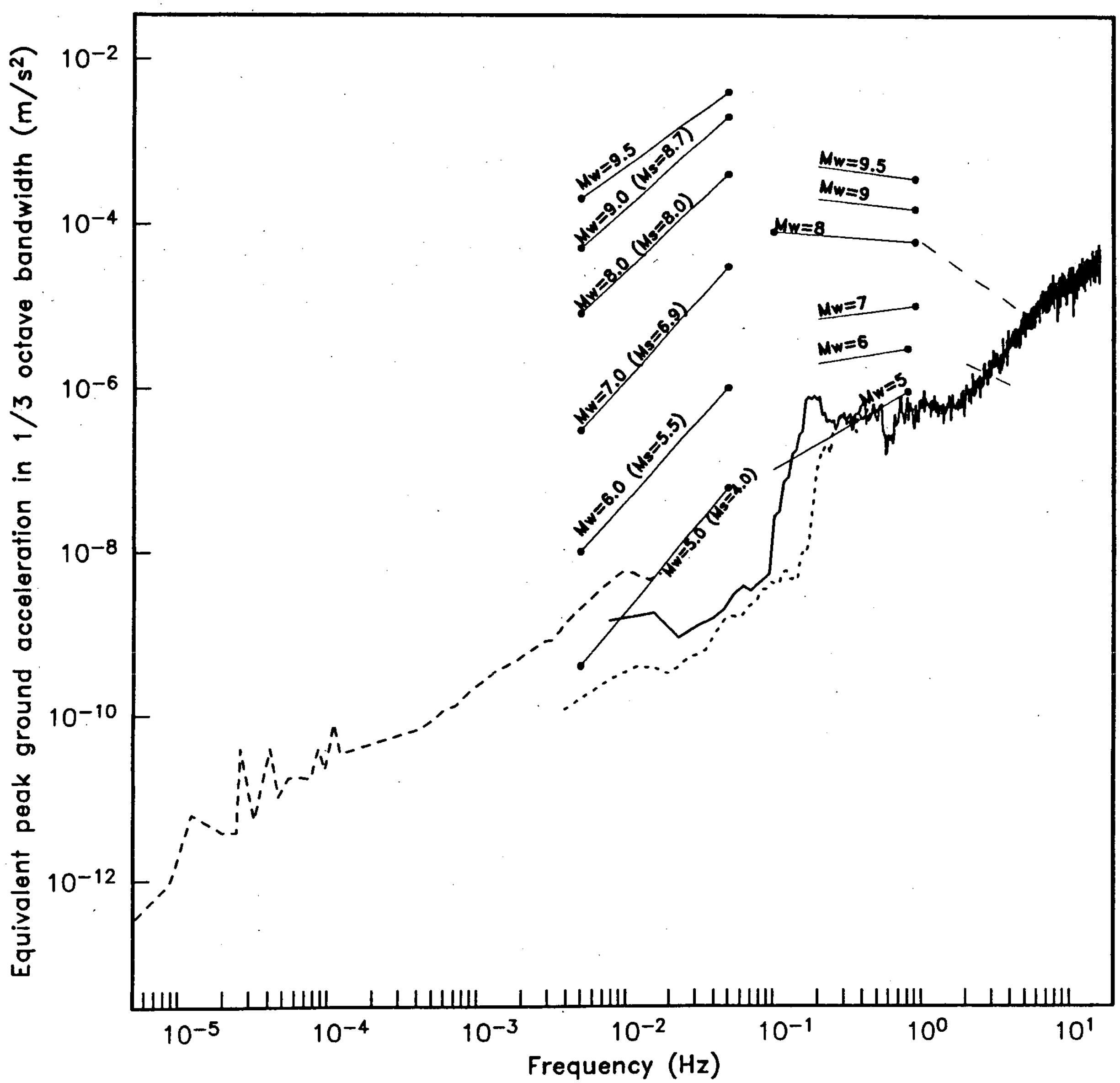

Pressure power (Pascal^2/Hz)
10^{10}
10^{8}
10^{6}
10^{4}
10^{2}
10^{0}
10^{-2}
10^{-6}
10^{-5}
10^{-4}
10^{-3}
10^{-2}
10^{-1}
10^{0}
10^{1}
Frequency (Hz)

Figure 1



Equivalent peak ground acceleration in 1/3 octave bandwidth (m/s²)
10⁻²
10⁻⁴
10⁻⁶
10⁻⁸
10⁻¹⁰
10⁻¹²
Frequency (Hz)
10⁻⁵
10⁻⁴
10⁻³
10⁻²
10⁻¹
10⁰
10¹
Mw=9.5
Mw=9.0 (Ms=8.7)
Mw=8.0 (Ms=8.0)
Mw=7.0 (Ms=6.9)
Mw=6.0 (Ms=5.5)
Mw=5.0 (Ms=4.0)
Mw=9.5
Mw=9
Mw=8
Mw=7
Mw=6
Mw=5

Figure 2

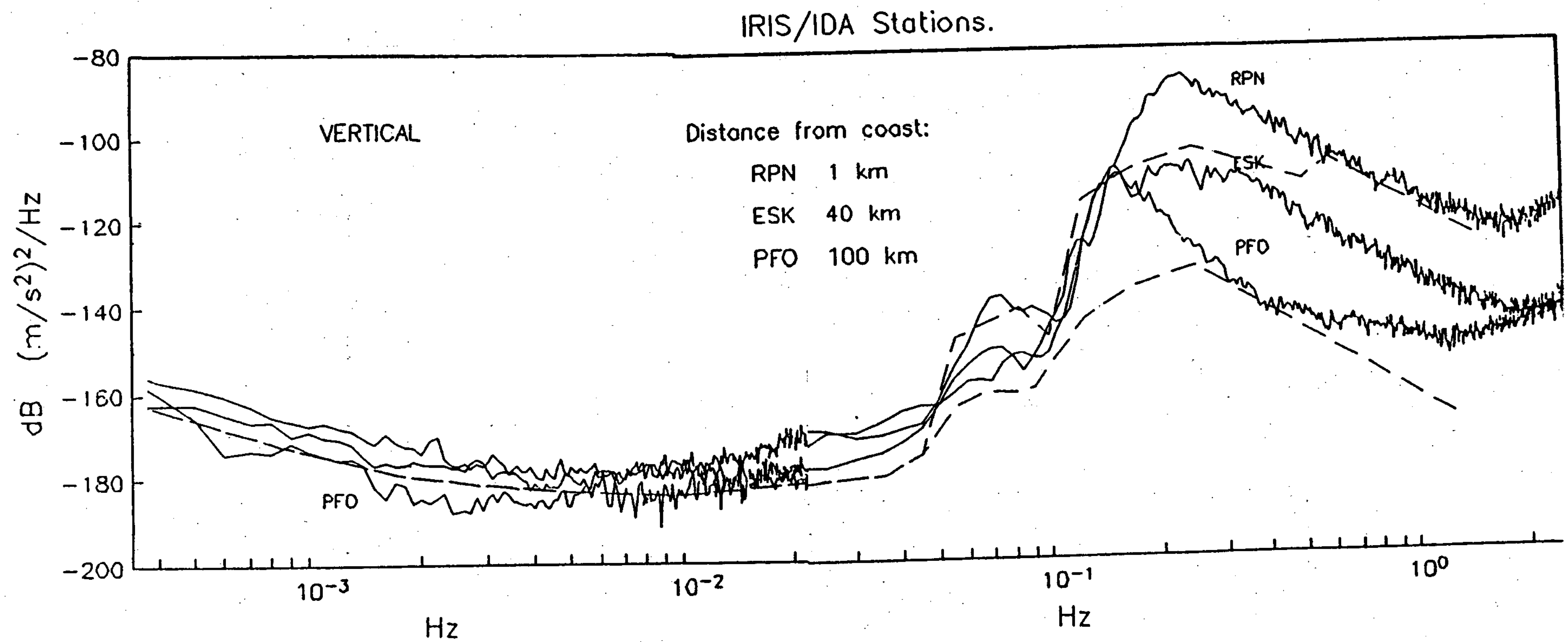

IRIS/IDA Stations.
VERTICAL
Distance from coast:
RPN 1 km
ESK 40 km
PFO 100 km
RPN
ESK
PFO
PFO
dB (m/s²)²/Hz
-80
-100
-120
-140
-160
-180
-200
10⁻³
10⁻²
10⁻¹
10⁰
Hz
Hz

Figure 3

A.3 RECOMMENDATIONS

Understanding the internal dynamics of the Earth - how it moves and evolves over geologic time - is the key to understanding many important geologic processes on the Earth's surface. Seismology concerns the excitation, propagation and recording of elastic waves in the earth. The excitation of elastic waves provides the most detailed information about the kinematics and dynamics of the earthquake rupture process, and their propagation is the richest source of information about the composition and state of the Earth's interior. To record these waves seismologists are necessarily limited to a finite set of stations at or very near the surface of the Earth.

Recently, the pace of progress in mapping and modelling the properties of the deep interior of the Earth has quickened. Thirty years elapsed between the discovery of the fluid core by Oldham in 1906 and the discovery of the inner core by Lehmann; it took until 1971, another 35 years, for a satisfactory demonstration that the inner core is solid. Only two years separate the publication of the first three-dimensional maps of the upper mantle and the appearance in the literature of two independent arguments in favor of anisotropy of the inner core.

The availability of data from global digital networks is the principal reason for this dramatic increase in the rate of progress, even though those that contributed the early data have been very sparse and did not explore the full range of the seismic spectrum. The earth sciences community within the U.S. and abroad realize the scientific opportunities that would be provided by a denser, more uniform coverage of the Earth's surface in the state-of-the-art seismographic stations. This is the reason for initiatives such as GSN-IRIS in the United States, GEOSCOPE in France, MEDNET in Italy or CANDIS in Canada. Ten countries, either operating or planning deployment of broad-band digital seismographic stations, now form the membership of the Federation of Digital Seismographic Networks. The member countries are: Australia, Canada, China, France, Germany, Italy, Japan, U.K., U.S.A. and U.S.S.R.

However, the coverage of the Earth's surface cannot be complete and, therefore, the scientific goals fully achieved, without placing permanent stations on the ocean bottom. The cover of this report shows a world map, with its surface divided into 128 equal area regions ("squares") of, roughly, 2,000 x 2,000 km dimensions. Those "squares" that cannot contain a seismic station because of the absence of any land-mass, including small islands, are shaded darkly or are left blank. They represent major continuous zones in the eastern Pacific, South Atlantic and Indian Ocean. Without the deployment of geophysical observatories at most of these "squares" our sampling of the Earth's structure and monitoring of its tectonic activity will remain incomplete and aliased.

We recommend that plans be made to establish a global network of approximately 15-20 permanent broad-band ocean floor seismic observatories. A solid base of scientific justification exists to support this ambitious goal. However, before it is realizable, a number of preliminary step must be taken.

A most important and specific recommendation of the workshop is that **during the next 2-3 years a number of pilot experiments should be carried out** to establish reliable solutions to the instrumental and environmental problems associated with the design of a permanent global network. Key problems are knowledge of the optimal regional siting of observatories to provide improved signal-to-noise ratio data, determination of the best sensor emplacement strategy to compromise between station cost and data quality (e.g., what depth below the seafloor? In sediments or volcanics?), design of a reliable, remotely operable and rugged broad-band sensor, and construction of adequate low-power data storage/telemetry and timing systems. The specific objectives of these first pilot experiments should be to:

- understand sources, propagation mechanisms and environmental controls on ocean floor noise in the band 3 mHz to 50 Hz

- determine the dependence of noise spectra upon the depth of burial of sensors beneath the ocean floor in a range of appropriate tectonic regimes

- compare both signal and noise data from seafloor and subseafloor broadband sensors with that from nearly island sites.

- prove the operational reliability of sensors, data recording and/or telemetry schemes, power sources and timing systems during long-term (>1 year deployments.

Clearly such a series of experiments would provide important acoustic and seismological results quite independent from the long-term goals of this program. They should be located to optimize the usefulness of the structural and tectonic data they collect. Important advances can be expected from single, broad-band, long-term stations deployed at the following sites:

- Nazca-Pacific-Cocos ridge-ridge-ridge triple junction. A station at this triple junction could sample upper mantle structure along three spreading centers ranging from intermediate to very fast spreading rates using surface waves and body waves.

- Outer Rise seaward of a trench. A station on the outer rise could be used to study the source characteristics of outer rise and trench earthquakes, determine the structure of the downgoing slab with body waves and is the ideal location for tsunami monitoring and warning.

- On the Hawaiian swell. A station on the crest of the swell sufficiently far from the islands could serve as an excellent site on 'normal' seafloor, isolated from rapidly varying structures that perturb the seismic waveforms.

- Near the South America-Antarctic-Africa triple junction. This would fill one of the largest gaps in global coverage of seismic stations and allow the study of the upper mantle beneath ridges with very close to intermediate spreading rates.

Because of logistic considerations, and the opportunity for comparisons with island site data, **we specifically recommend that the first pilot experiments be carried out near the Hawaiian islands**. A the time of writing this report, a drilling proposal is under review by JOIDES that will provide a suitable drill hole, with re-entry cone, within 100 n.m. of Oahu that should at the earliest opportunity be instrumented with appropriate sensors and recording systems to begin these studies.

We recommend that a steering group be formed, co-sponsored by JOI and IRIS, with the following mandate.

- to stimulate and coordinate efforts by groups of investigators to carry out the necessary pilot experiments to permit the informed design of an effective global network.

- to begin a process of long term planning for the emplacement of this network that includes substantial involvement of the international community.

- to ensure effective communication between IRIS, USGS and the global seismology community, and JOI-USSAC, JOIDES and the marine sciences community (with specific concern for the new global initiatives e.g., RIDGE).

- to work with funding agencies to plan realistic goals and time tables for progress that can be supported by the available resources.

- to serve as a focal point for information concerning progress of non-U.S. initiatives in deep ocean seismic observatories.

- to provide periodic reports for the community and for the funding agencies that describe progress in these endeavors and better define the emerging needs for resources and technology.

The member of this steering group should be endorsed by JOI and by IRIS and should consist of approximately five members of the U.S. academic community.

In addition to the primary recommendations described above, workshop participants recognized **three auxiliary efforts for which broad support was expressed.**

- The development of a routine wireline re-entry capability is critical to being able to carry out the pilot experiments efficiently and will result in considerable cost savings in the emplacement and servicing of the eventual global network.

- Broadband seismic stations should be established on a greater cross-section of available island sites in order to begin the process of reducing severe spatial aliasing problems and to learn more about the ocean environment.

- In the design of the planned observatories consideration must be given to the measurement of other parameters (e.g., strain, magnetic field) that will also enhance our understanding of earth structure. Section B6 contains several examples of measurements of other parameters or variables.

SECTION B

PAPERS CONTRIBUTED TO THE WORKSHOP

1. Scientific Goals
2. Existing and Planned Global and Regional Networks
3. Island Station and Ocean Floor Noise
4. Marine Downhole Seismic Experiments
5. Instrumentation
6. Other Measurements

Section B1: Scientific Goals

1. The Earth's Large-Scale Three-Dimensional Structure: John Woodhouse and Adam Dziewonski
2. Ocean Floor Stations and the Measurement of Anisotropy of the Oceanic Lithosphere: Toshiro Tanimoto
3. Ocean Floor Stations and the Measurement of Regional Surface Wave Dispersion, Anisotropy and Attenuation: Brian J. Mitchell
4. The Application of Ocean Floor Stations to Tsunami Prediction: Gordon D. Burton
5. Resolution of Changes in Slab Dip via Body Wave Inversion: Importance of Pacific Ocean Basin Seismic Stations: Larry J. Ruff
6. Ocean-Floor Seismic Stations and Source Mechanism Studies: Hiroo Kanamori

THE EARTH'S LARGE-SCALE THREE-DIMENSIONAL STRUCTURE

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Introduction

Plate tectonics, which, over the last 20 years, has provided the framework for understanding large-scale geological processes, describes the motion of large, rigid sections of the Earth's crust and uppermost mantle. The plates are transported by mantle convection currents, but a definitive understanding of mantle convection has not yet been reached. Seismic tomography offers the opportunity to investigate the interior of the convective system by mapping, in three dimensions, the seismic wave velocity variations. These are related to temperature and composition and, in particular, to the density variations which provide the driving force of mantle convection.

Figures 1*a-d* show a composite of models of the upper mantle S-velocity (M84C: Woodhouse and Dziewonski, 1984) and the lower mantle P-velocity (L02.56: Dziewonski, 1984), and a schematic representation of the anisotropic properties of the inner core (Woodhouse *et al.*, 1986; Morelli *et al.*, 1986). Lighter shades indicate lower than average wave velocities at a given depth and darker shades indicate higher than average velocities. These figures illustrate the scale lengths of global heterogeneity which have been resolved. In the upper mantle, the model is expanded up to degree 8 in spherical harmonics and as a cubic polynomial in depth, corresponding to nominal resolving lengths of approximately 2500 km horizontally and 150 km vertically. In the lower mantle the model is expanded up to degree 6 and as a quartic polynomial in depth, corresponding to resolving lengths of roughly 3000 km and 400 km respectively. These models represent, therefore, the result of a spatial filter applied to the true state of heterogeneity in the Earth. Such filtered versions of reality would, perhaps, be of limited interest if heterogeneity on smaller scales were dominant, but it appears that the spatial spectrum of heterogeneity contains very strong long wavelength components. In the upper mantle, for instance, S-velocity variations in spherical harmonic degrees up to 5 are as large as $\pm 4\%$ in the upper 150 km and are much larger than the expected signature of subducted slabs in this range of wavelengths. Similarly in lower mantle P-velocity spherical harmonic degrees 2 and, to a lesser extent, 4 are dominant terms and have amplitudes of the order $\pm 0.5\%$ (See Figs. 2-5).

The study of lateral heterogeneity is of great significance to seismology. For example, in the investigation of earthquakes lateral heterogeneity - like an imperfect lens - can distort the image of an event. Estimates of location, fault length, and the pattern of stress release can be false if the medium is inadequately known. Even the introduction of corrections for longwavelength lateral heterogeneity can result in shifts in inferred epicenters by as much as 20 km and changes in origin times of more than 1 second.

Outside the field of seismology, the recent results on the Earth's three dimensional structure have an impact on other fields of Earth sciences. Several examples of such linkage follow.

Mantle convection. Under the assumption that seismic anomalies are proportional to density perturbations, they provide constraints on the modelling of mantle convection and on the viscosity distribution in the mantle (Richards and Hager, 1984; Forte and Peltier, 1987, 1988; Hager and Clayton, 1988).

Petrology and geochemistry. Models of seismic anomalies in the upper mantle show the oceanic ridges to be the dominant regions of low velocity at shallow depth. The long wavelength models tomography models have the potential to provide integral constraints on petrological and thermal models of the ridge systems. Deep high velocity anomalies are associated with the continental shields, apparently confirming the hypothesis of 'continental roots' (Jordan, 1975, 1978). About 80% of hot spots occur over regions of lower than average wave velocities near the core-mantle boundary (see Fig. 5). There also appears to be a correlation between a band of low velocity anomalies near the core-mantle boundary, in the latitude band from 10°S to 30°S, and the occurrence of a large scale isotopic anomaly (Dupal anomaly; Hart, 1984; Castillo, 1988).

Geomagnetism. The regions at the core-mantle boundary where the magnetic field changes with time most rapidly coincide with low velocity seismic anomalies which, presumably, represent regions of elevated temperature in the lowermost mantle. The inference has been made that the thermal state of the lowermost mantle determines the stability of convection patterns, and hence of the geomagnetic field, in the outer core (Bloxham and Gubbins, 1987).

Gravity. It was early recognized (Dziewonski *et al.*, 1977) that there was a strong correlation between P-velocity anomalies in the lower mantle and the long wavelength geoid undulations, but that the correlation had the opposite sign from that expected if high velocities correspond to high densities - as would the case if temperature fluctuations were responsible for the seismically observed anomalies. It has been shown, however, (Pekeris, 1935; Richards and Hager, 1984; Hager *et al.*, 1985) that density anomalies embedded in a fluid mantle induce deflections of the free surface and of the core-mantle boundary which can change the sign of the inferred geoid perturbations; thus lower mantle heterogeneity has been identified as a major cause of long wavelength geoid anomalies.

Mineral physics. An example of an *in situ* measurement made possible through seismic tomography is the ratio of relative perturbations in the shear and compressional velocities: $d\ln v_S/d\ln v_P$. The tomographic results yield a ratio much higher than determined, at relatively low pressures, in the laboratory (Anderson *et al.*, 1968). It appears that under the temperature and pressure conditions appropriate for the lower mantle, the shear modulus is much more sensitive to changes in temperature than the bulk modulus. Similar observations for the upper mantle have been interpreted as being associated with partial melting (Hales and Doyle, 1967).

Geodesy and astronomy. From the analysis of data on the Earth's rotation, obtained by the VLBI (very long baseline interferometry) technique, it has been determined that the flattening of the Earth's core departs from its equilibrium value by 400±100 meters (Gwinn *et al.*, 1986). While the seismically determined core mantle boundary topography (Morelli and Dziewonski, 1987; has a range of ±6 km, the component corresponding to excess flattening is very small, and error estimates are such that the seismic results are consistent with the geodetically determined flattening.

Tomographic models and their implications

Studies of the large-scale three-dimensional structure of the Earth have been carried out using various kings of seismological data, spanning more than three orders of magnitude in frequency (1 Hz - 0.0005 Hz). These are *(i)* large collections of P (and PDP, PKIKP, PcP) travel times, *(ii)* measurements of phase and group delays and amplitude anomalies of surface waves and measurements of the locations of spectral peaks of fundamental modes, interpreted asymptotically, *(iii)* complete waveforms of mantle waves, used as data in a least squares inversion *(iv)* complete waveforms of long period body waves and *(v)* complete spectra of split multiplets in the Earth's free oscillation spectrum. [See Dziewonski and Woodhouse (1987), and Woodhouse and Dziewonski (1988) for recent reviews.]

Studies of class *(i)* have illuminated lower mantle P-velocity structure (Dziewonski et al. 1977; Dziewonski, 1984); Clayton and Comer, unpublished), and those of classes *(ii)* and *(iii)* have led to models of upper mantle S-velocity (Masters et al. 1982; Nakanishi and Anderson, 1982, 1983, 1984; Woodhouse and Dziewonski, 1984; Nataf et al. 1984, 1986). With the addition of classes *(iv)* and *(v)* (Woodhouse and Dziewonski, 1986; Tanimoto, 1987; Giardini et al. 1987, 1988) it has become possible to constrain lower mantle S-velocity structure and thus to obtain models of the same region of the Earth using different classes of data. This is very valuable in that it provides a check on the various modelling techniques and also allows the comparison of heterogeneity in different structural parameters in the same region. Using one kind of data alone, it is often difficult to completely rule out the possibility that systematic errors or deficiencies in coverage, which are inherent in the data, degrade or corrupt the resulting models. We find, however, that the application of different techniques is yielding a coherent picture of the Earth's global heterogeneity, reinforcing the conclusions drawn from each data set alone.

From such studies some of the major features of the Earth's three-dimensional structure are becoming clear and the process of interpretation is underway; it should be borne in mind, however, that the interpretation of the 'geographical' information provided by seismology is a matter of considerable complexity and we may expect that many of the conclusions to be drawn from the tomographic models lie in the future.

In the upper mantle the strongest low velocity features are associated with the fast-spreading oceanic ridges, together with regions in the western Pacific, characterized by back-arc volcanism (see Fig. 2). High velocities are associated with the continents, and in particular with the continental shields, extending to depths in excess of 300 km (Woodhouse and Dziewonski, 1984; see Jordan, 1978). While the latter can be understood in terms of the upwelling of hot material as the plate move apart, the deep extensions of the continents present some difficulty since it may be expected that vigorous convection would rapidly destroy them. Probably they are compositionally different, cooler and of higher viscosity than the sub-oceanic mantle (Jordan, 1988). By translating the observed velocity anomalies into density anomalies and performing a dynamical calculation of fluid flow in a mantle of assumed viscosity structure, these models have been used to infer the expected fluid motion at the surface (Forte and Peltier, 1987). Comparison with the poloidal component of the observed motion of the plates shows fairly good agreement at very low degree and highly significant correlations at higher degrees and thus it has become possible to begin to understand the internal forces which drive plate motion. An important qualitative conclusion from such analysis is that the magnitude of the observed seismic anomalies is of the order expected in a convecting system having the viscosity, temperature derivatives and flow rates which characterize the mantle.

Much interest interest in geodynamics centers on the trajectory in the mantle of material subducted at convergent plate boundaries. This is because it bears on the question of how well the mantle is mixed, which is of great importance in understanding the Earth's chemical evolution. While the debate on this issue will doubtless continue for some time to come, the most natural inference from both global (Dziewonski, 1984; Woodhouse and Dziewonski, 1986) and regional (*e.g.*, Creager and Jordan, 1984, 1986; Grand, 1987) seismological studies is that cold subducted material penetrates the boundary, although a significant increase in viscosity may act to reduce the velocity and to increase the cross-sectional area of the descending flow. Present day subduction takes place principally around the rim of the Pacific; models of the lower mantle show that a very long wavelength pattern of high velocities around the Pacific persists throughout the lower mantle (Dziewonski *et al.*, 1977; Dziewonski, 1984; Woodhouse and Dziewonski, 1986; Morelli and Dziewonski, 1986; Giardini *et al.*, 1987) and, moreover, that its morphology bears the closest resemblance to what may be expected due to subduction in the upper part of the lower mantle. These high velocity features include the rim of the Pacific and a region stretching from Indonese to the Mediterranean which marks the Tethys convergence zone. Figures 3*a*, *b* show independent models of P-velocity inferred from travel times (Morelli and Dziewonski, 1986) and S-velocity obtained by the inversion of long period body waveforms (Woodhouse and Dziewonski, 1986), at the depth 1300 km. It will be noted that the anomalies are displaced outwards from the Pacific relative to the current loci of subduction, and that regions such as North America and southern Eurasia, which have been stronger convergence zones in the past than they are at present, are strongly represented in the models. In North and Central America these models are in good agreement with the regional model of Grand (1987) which shows, with somewhat higher resolution, a high velocity N-S trending anomaly in the depth range 800-1200 km. At its southern end, in the Caribbean, this coincides with an anomaly identified by Jordan and Lynn (1974) and by Lay (1983). Jordan and Lynn (1974) and Grand (1987) have suggested that the anomaly represents material subducted beneath the North American continent, and the fact that it forms a part of a global feature skirting most of the subduction zones of the Pacific tends to substantiate this.

This pattern blends, with increasing depth, into long wavelength pattern of high velocities around the Pacific, largely contained in spherical harmonics of degrees 2 and 4, which continues to the core-mantle boundary. Figures 4*a*, *b*, *c* show the degree 2 and 4 components of the pattern at the depth 2300 km, from three independent seismological studies. This feature, which fills a major fraction of the Earth's volume, is undoubtedly of fundamental significance in understanding the Earth's long term dynamics; it remains to be determined, through geodynamical modelling, whether this feature *explains* or whether it is *explained* by the fact that subduction zones have tended to be in fixed locations for long periods of Earth history. That is: is the tomographically observed anomaly to be attributed to material subducted in the same locations for long periods of time, or are the locations of convergent plate boundaries governed by a long-lived thermal anomaly in the lower mantle?

This anomaly has been detected in both compressional and shear wave velocities (see Fig. 4), and yields a ratio of relative perturbations in v_s and v_p in the lower mantle in the range 2-2.5. Such values, which are much larger than has sometimes been assumed, roughly correspond to the case that perturbations in shear modulus dominate those in bulk modulus. In addition, by assuming that density anomalies, proportional to wave velocity anomalies, act as internal loads in a viscous, fluid mantle, it has been shown that the observed low degree geoid undulations are caused by lower mantle heterogeneity and that lower mantle viscosity is substantially higher than that of the upper mantle (Hager et al., 1985).

Just outside the core the pattern of high and low velocities is such that approximately 80% of hot spots at the surface are above regions of lower than average velocity in the lowermost mantle, lending support to the hypothesis that hot spots are the surface manifestation of plumes rooted in the deepest part of the mantle (see Fig. 5).

The shape of the core mantle boundary, in spherical harmonic degrees up to 4, has been determined using the travel times of both reflected and transmitted waves (Morelli and

Dziewonski, 1978), yielding consistent results where coverage is adequate. The resulting model exhibits a pattern of depressions encircling the Pacific, having an amplitude of approximately $\pm$ 6 km, which has been shown to be consistent with the stresses induced by density anomalies inferred from tomographic models of the lower mantle (Forte and Peltier, 1988). However, studies incorporating rays which have high angles of incidence (almost grazing) have shown a different pattern (Creager and Jordan 1987). This apparent discrepancy is probably a further manifestation of the complex heterogeneity in the lowermost mantle and, possibly, in the outermost core.

Finally, while no well documented heterogeneity has been reported in the fluid outer core - and, indeed, is not to be expected owing to the inability of an inviscid fluid to maintain such heterogeneity - the inner core has been found to be characterized by an anisotropic, crystalline structure (Woodhouse et al., 1986; Morelli, et al., 1986; Shearer et al., 1988) in which there is preferential alignment of the high velocity axes parallel to the Earth's rotation axis. This enigmatic observation is possibly evidence low degree convection in the inner core (Jeanloz and Wenk, 1988).

Discussion

An important goal of seismology is to obtain more detailed representations of the Earth's internal heterogeneity. The major improvements in seismic instrumentation that are underway, together with advances in techniques of analysis and in computational power, will enable considerable progress to be made. We believe, however, that only if instruments are deployed on the ocean floor will it be possible to greatly improve the tomographic models on a global, rather than on a regional basis. The lack of coverage by seismic instruments in the oceans leads to inaccuracy in seismic models, particularly in the Pacific region and in the Southern oceans. This shortage of information is particularly regrettable since the oceans, in general, but the Pacific in particular, contain the most significant global tectonic process - the fastest spreading ridges, the most active hot spots, the oldest ocean floor, the majority of subduction zones. The ability to determine how these surface manifestations of mantle convection express themselves at depth will be of crucial importance in understanding the dynamics of the Earth as a whole.

Even a small number of instruments on the ocean floor can considerably enhance our ability to determine deep earth structure, since they provide control points of the mesh of intersecting paths. Placing instruments in a sparsely sampled area does not only provide a proportionate number of additional paths, but it also enhances the value of all the other available data which are sensitive to the structure of the undersampled region.

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FIGURE CAPTIONS

Figures 1a, b, c, d. Three dimensional sections of models M84C and L02.56, together with a schematic illustration of the anisotropic properties of the inner core. The depth of the section, in km, is indicated. In b the upper mantle is shown to the depth 550 km with the depth scale exaggerated by a factor of 5. Plate boundaries are indicated.

Figure 2. Relative S-velocity perturbations in the model U84L85/SH of Woodhouse and Dziewonski (1986), based upon the analysis of SH waveforms; the depth is 150 km, characteristic of the uppermost mantle. The model contains spherical harmonics up to degree 8. Plate boundaries are indicated.

Figure 3a. The P-velocity travel time model V3.I of Morelli and Dziewonski (1986) at depth 1300 km. The model contains spherical harmonics up to degree 6. Note that the upper captions on the scale (±0.5%) relate to a.

Figure 3b. The S-velocity travel time model U84L85/SH of Woodhouse and Dziewonski (1986) at depth 1300 km. The model contains spherical harmonics up to degree 8. Note that the lower captions on the scale (±1.0%) relate to b.

Figure 4a. P-velocity model V3.I, depth 2300 km, degrees 2 and 4 only; scale represents the range ±0.3%; see caption to Fig. 3a.

Figure 4b. S-velocity model U84L85/SH depth 2300 km, degrees 2 and 4 only; scale represents the range ±0.6%; see caption to Fig. 3b.

Figure 4c. S-velocity model 'Model 1' of Giardini et al., 1987), based upon free oscillation data, degrees 2 and 4 only; depth 2300 km, degrees 2 and 4 only; scale represents the range ±0.6%.

Figure 5. P. velocity model V3.I, depth 2750 km, all degrees 1-6; scale represents the range ±1.0%; hot spots are indicated; see caption to Fig. 3a.

a)

b)

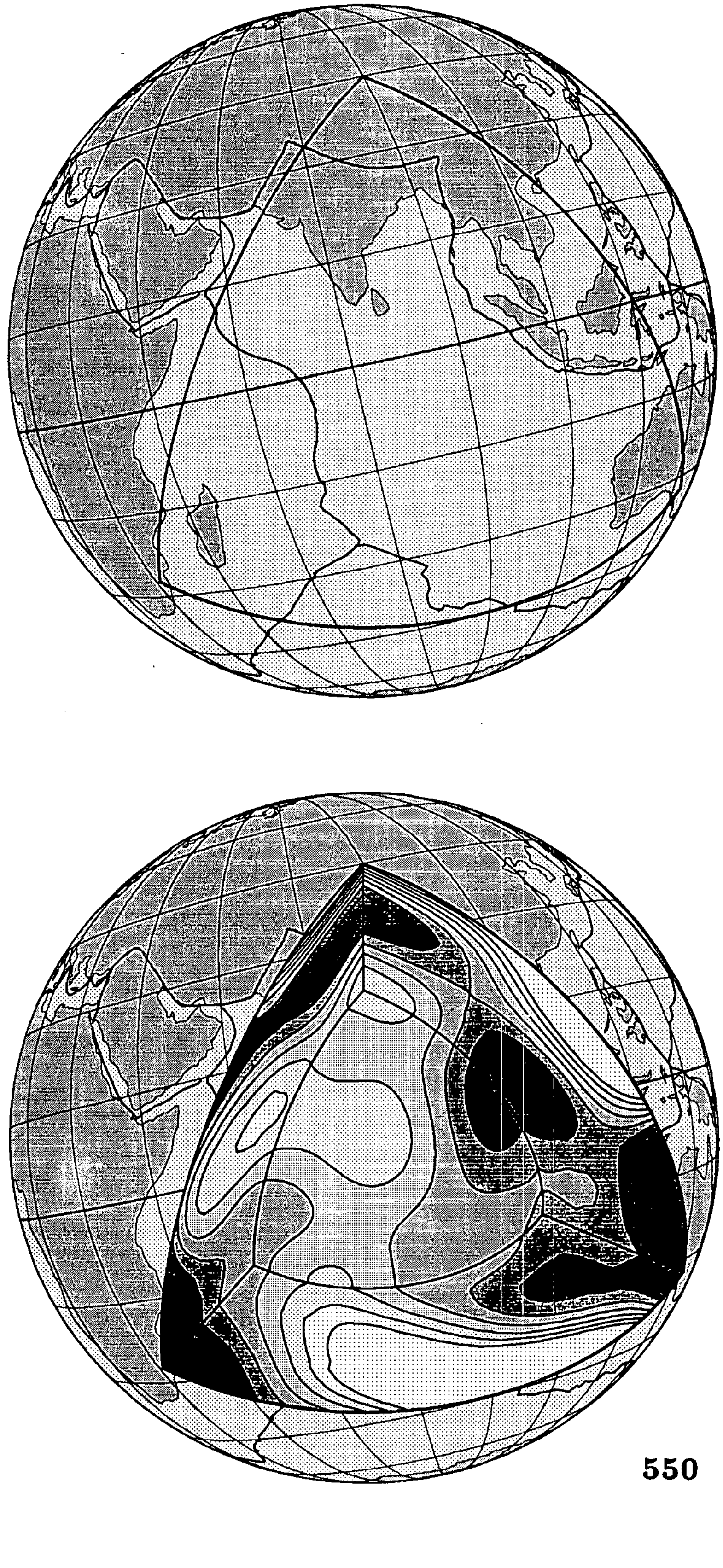
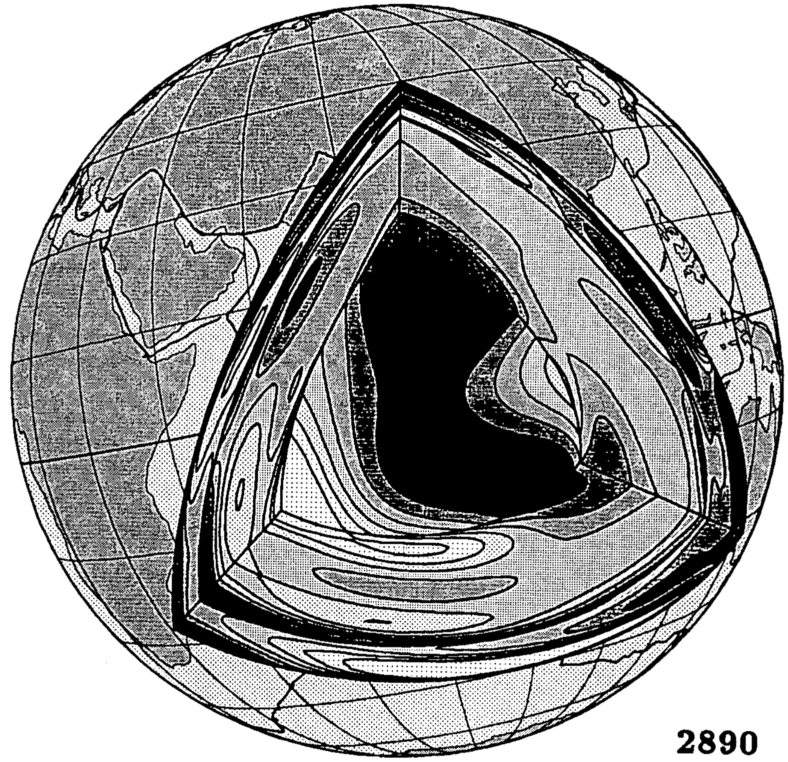


Fig. 1.

c)



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d)

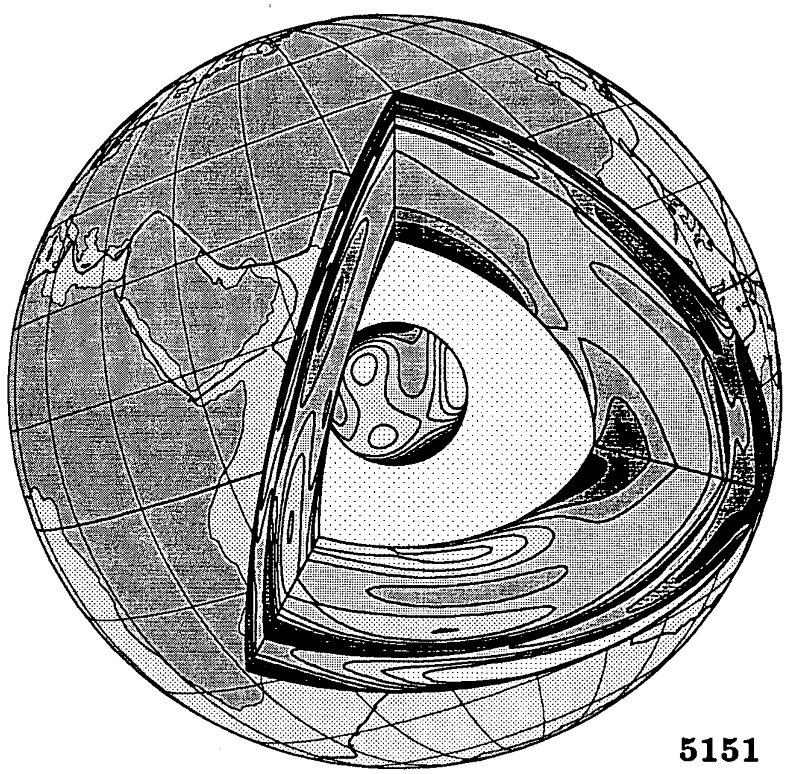

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Fig. 1. cont.

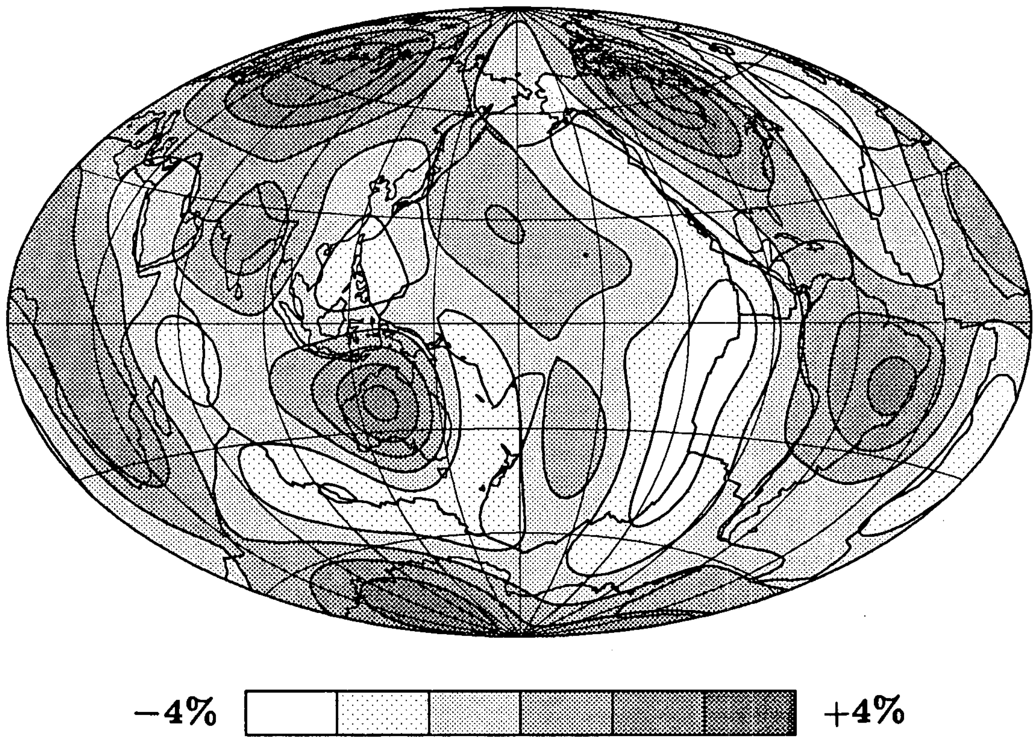

Fig. 2.

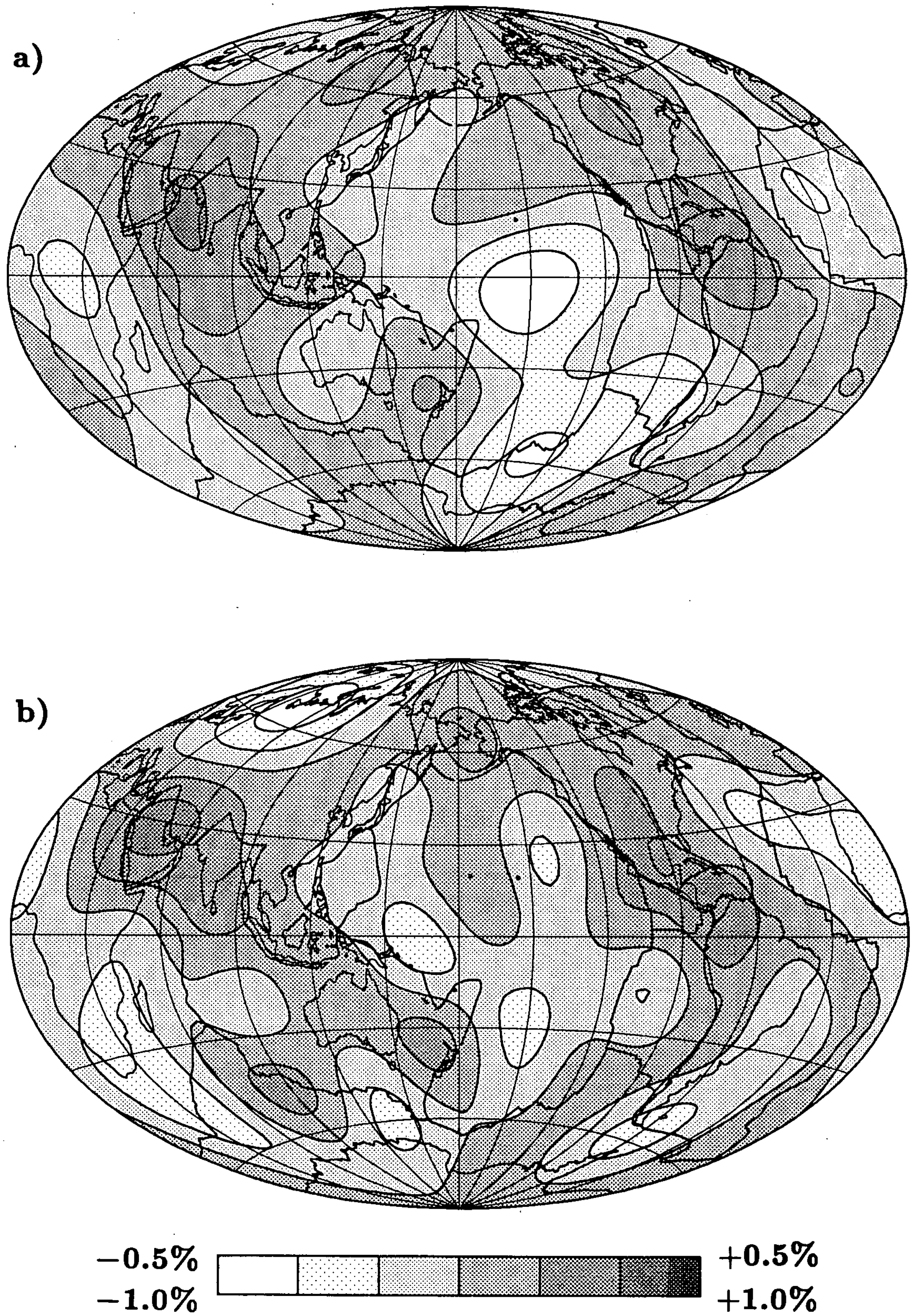

Fig. 3.

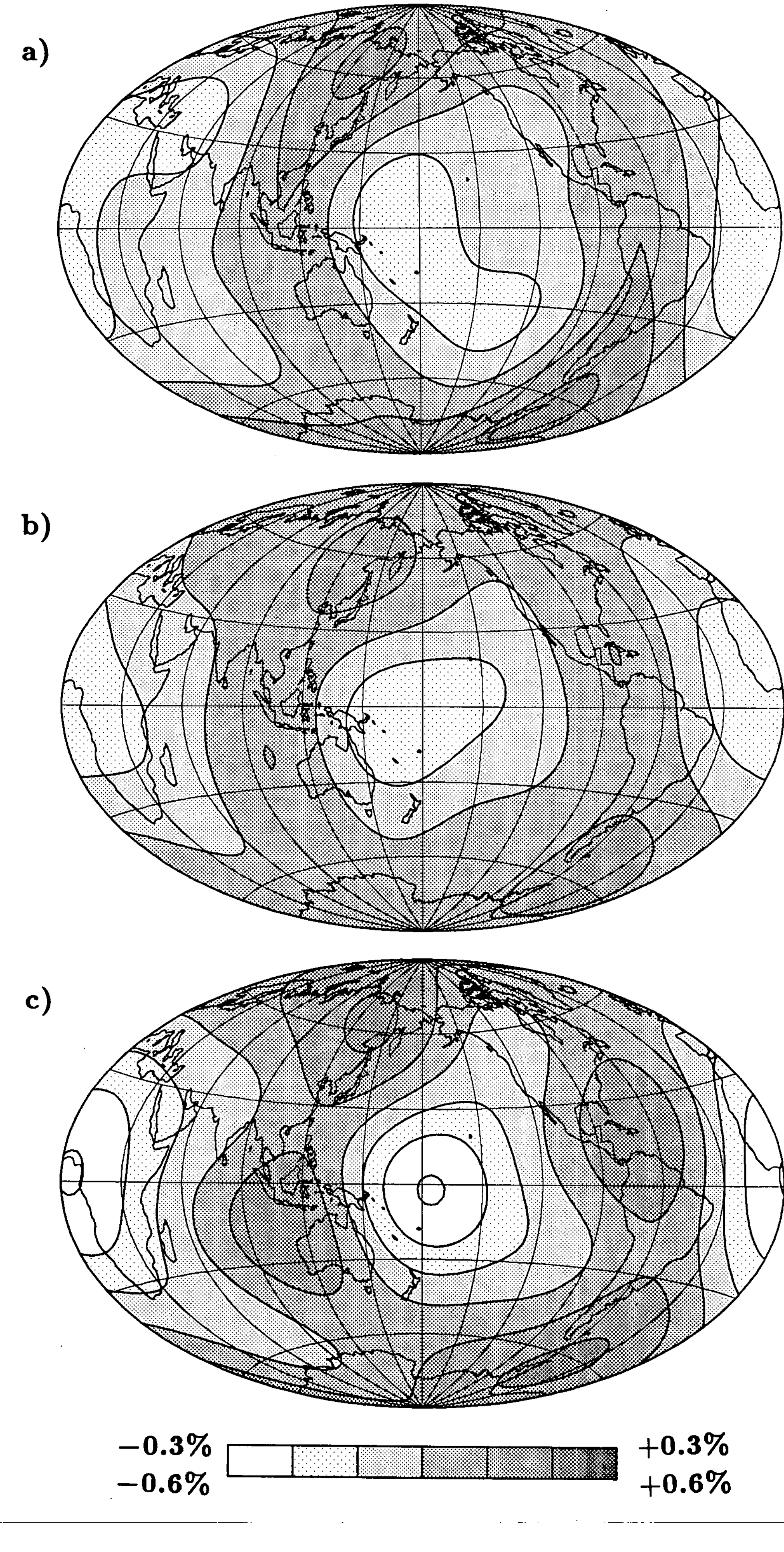

Fig. 4.

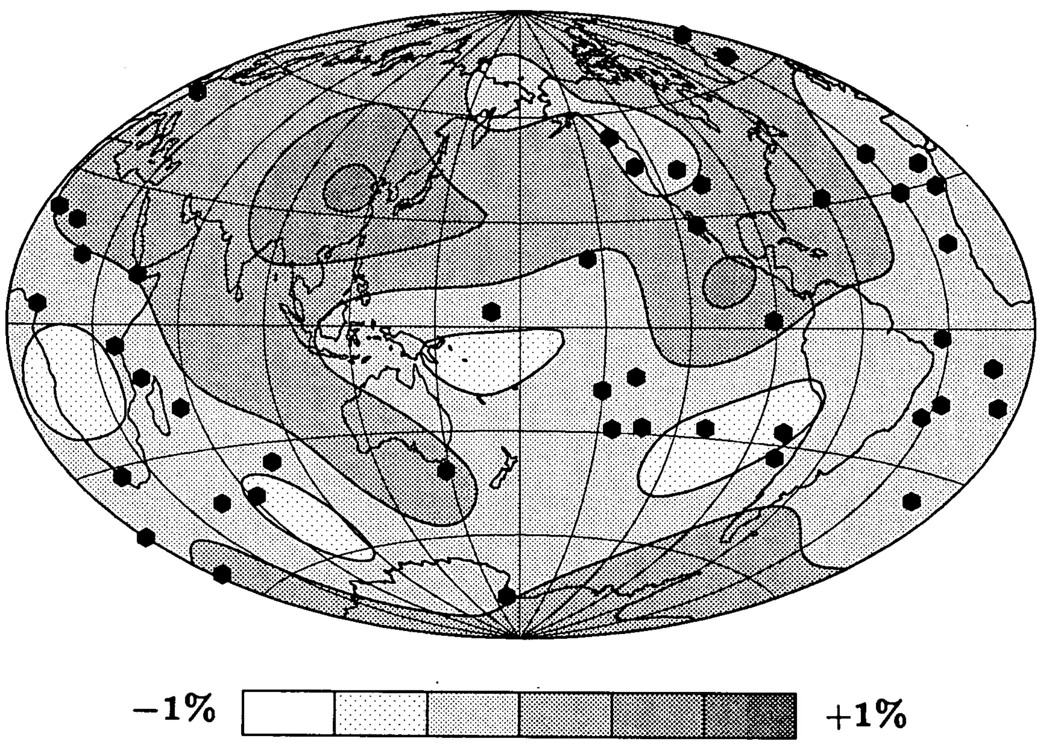

Fig. 5.

OCEAN FLOOR STATIONS AND THE MEASUREMENT OF ANISOTROPY OF THE OCEANIC LITHOSPHERE

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Introduction

During the last five or six years, we have observed the power and the success of the digital seismograph networks. There are some clear deficiencies, however, due mainly to their locations on land. In the following, I first discuss some of the deficiencies using our study of the Pacific Ocean (Zhang and Tanimoto, 1988) and then discuss future possibilities of detecting anisotropy in the oceanic lithosphere.

Limitations of the current digital network

Figure 1 shows the path coverage used for a study (Zhang and Tanimoto, 1988) in the Pacific ocean and its adjacent area. Approximately 120 earthquakes between 1980 and 1985 recorded by GDSN (Global Digital Seismograph Network) were used for this study. Although one could improve this study slightly by adding other network data (e.g. IDA and GEOSCOPE) and a few more years worth of data since 1986, I believe most of the arguments in the following remain to be true.

There are two features which one might notice in Figure 1. They are (1) the northern half of the Ocean is well covered while the coverage in the southern half is sparse and (2) there are many more paths going in the NE-SW direction than in the NW-SE direction.

One can quantify those features and get more insights by the singular value decomposition analysis (SVD); we divide the structure into 10°x10° blocks and constructed a matrix A whose ij component gave the ray length in the j-th block for the i-th ray. If we examine the eigenvector matrix V as in $A = U\Lambda V^T$ (singular value decomposition), we can get the better understanding of the limitations of the data set.

Fig. 2a shows the eigenvector associated with the largest singular value, which is the most robust feature in the solution. Black and white triangles indicate negative and positive numbers in Fig. 2a-2c, but the sign of the patterns are arbitrary since they are eigenvectors. Fig. 2a demonstrates that the most robust feature in the solution is the average velocity in the northern half, which is to be expected since the path coverage in the northern half is excellent in Fig. 1. Fig. 2b shows the eigenvector for the second largest singular value. It shows one cycle of sinusoidal modulation in the NW-SE direction while it is almost uniform in the NE-SW direction. This is closely related to the fact that there are many more paths going in the NE-SW direction than in NW-SE direction. In general, this feature persists for most eigenvectors associated with large singular values. Fig. 2c is for the sixth largest singular value which show more oscillation in the NW-SE direction. Under such circumstances any anomalies tend to be elongated in the NE-SW direction because of inadequate path coverage.

Based upon the above observations, we can complement the current network and obtain more balanced resolution by (1) installing seismic (OBS) stations in the southern area and (2) also installing OBS more in NE-SW direction than in NW-SE direction so that we can chop up the elongated patterns such as those in Figs. 2b and 2c.

Anisotropy in the lithosphere

Anisotropy just below Moho was established by refraction studies in the late sixties. How deep the anisotropy persists below that depth in a question which has not been answered. The depth extent of anisotropy could give us some clues in understanding the evolution and dynamics of the lithosphere, since alignment of crystal structure may occur as the lithosphere cools and thickens with age.

There are obviously two approaches, one using body waves and the other surface waves. Body wave studies will require a large number of seismic stations before one can grasp the overall picture but will yield high resolution results. On the other hand, it is relatively easier to obtain the large scale, big picture by the surface wave approach while the resolution cannot be sharp. In the following, I restrict the discussion to surface waves.

For surface waves, the first step in getting anisotropy is to detect the directional dependence of surface waves. Theoretically, velocity should show 2ψ ($\cos 2\psi$ and $\sin 2\psi$) and 4ψ ($\cos 4\psi$ and $\sin 4\psi$) dependence for a weakly anisotropic media, where ψ is the azimuth (ordinarily measured clockwise from north). In practice, coefficients for 2ψ and 4ψ terms must be obtained simultaneously with lateral heterogeneity, which obviously requires a huge amount of data. Figs. 3a and 3b show an attempt to resolve the 2ψ coefficient, Fig. 3a for $\cos 2\psi$ and Fig. 3b for $\sin 2\psi$, using the same data set given in Fig. 1. Resolution in those figures is considerably worse than the case when only lateral heterogeneity is investigated. It seems that a much larger data set is required to resolve azimuthally dependent velocity. On the other hand, one can perhaps state that if we complement the current network by OBS as (for example) suggested at the end of the last section, it will be only a matter of time to resolve azimuthally dependent velocity.

Lastly, we briefly discuss how the coefficients for 2ψ and 4ψ terms can be used to infer the anisotropy in the lithosphere. In a weakly anisotropic sphere, eigenfrequency perturbation is given by

$$\frac{\delta\omega}{\omega} = \frac{1}{2\omega^2 I}\left(h + C_2 \cos 2\psi + S_2 \sin 2\psi + C_4 \cos 4\psi + S_4 \sin 4\psi\right)$$

where ω is the eigenfrequency, I is the normalization for a mode, h is the heterogeneity perturbation, C_2, C_4, S_2, and S_4 are azimuthal terms. The observables are h, C_2, S_2, C_4 and S_4. In Figs. 4a and 4b, we show the theoretical variation of C_2 and C_4 for four different models. We assumed the transversely isotropic symmetry but the symmetry axis is in the horizontal plane. Five elastic constants, C, A, L, N, F, are assumed to have the relations $\varepsilon_1 = (L-N)/A = 0.018$, $\varepsilon_2 = (C-A)/A = 0.073$ and $\varepsilon_3 = (C+A-2F-4L)/A = -0.022$, which determine the results completely. Four different models have the same anisotropy in (a) the upper 40 km (upper half of the lithosphere), (b) the upper 80 km (entire lithosphere), (c) the upper 220 km and (d) the upper 400 km. Curves for C_2 are given by solid lines while dash lines are used for C_4. In general, Rayleigh waves possess large 2ψ terms while Love waves have large 4ψ terms, at least for the assumed anisotropy. Since the curves show some differences as a function of frequency, it seems possible to put constraints on the depth extent of anisotropy. This is of course under the assumption that we know the anisotropy of the upper mantle material. Experimental determination of anisotropy for upper mantle material seems equally important in this endeavor.

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Figure Captions

Fig. 1: Path coverage of Love waves for the Pacific Ocean. Data is from GDSN between 1980 and 1985.

Fig. 2a: Eigenvector of the largest singular value, which gives the average velocity in the northern half.

Fig. 2b: Eigenvector of the second largest singular value.

Fig. 2c: Eigenvector of the sixth largest singular value.

Fig. 3a: Resolution kernel for the coefficient of cos 2ψ at Hawaii.

Fig. 3b: Same with 3a except for sin 3ψ term.

Fig. 4a: Theoretical variation of the coefficients of cos 2ψ (solid) and cos 4ψ (dash) for Rayleigh waves. Models a, b, c, and d have different depth extent of anisotropy (see text).

Fig. 4b: Same with 4a except for Love waves.

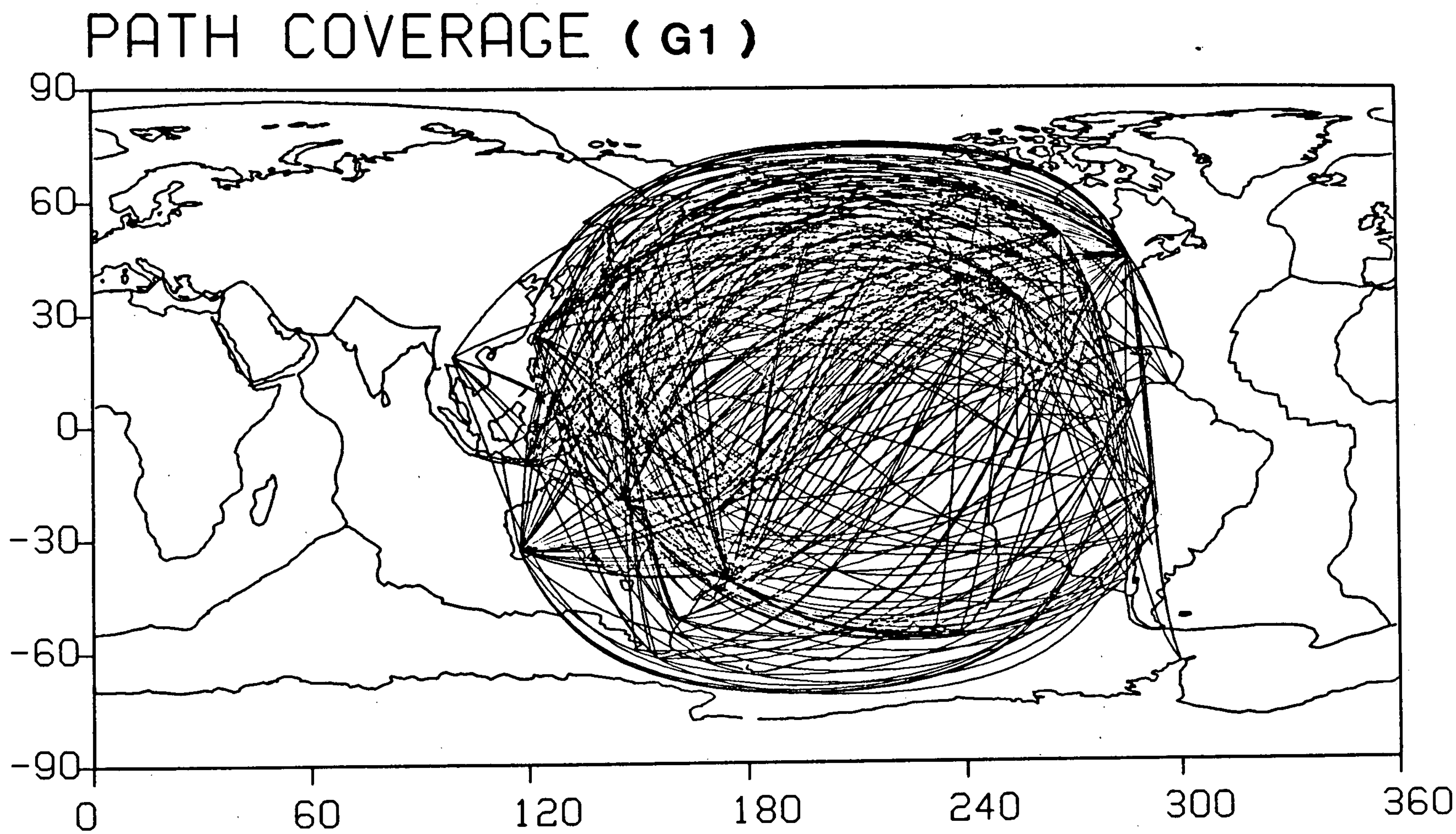

Figure 1: Path coverage of Love waves for the Pacific Ocean. Data is from GDSN between 1980 and 1985.

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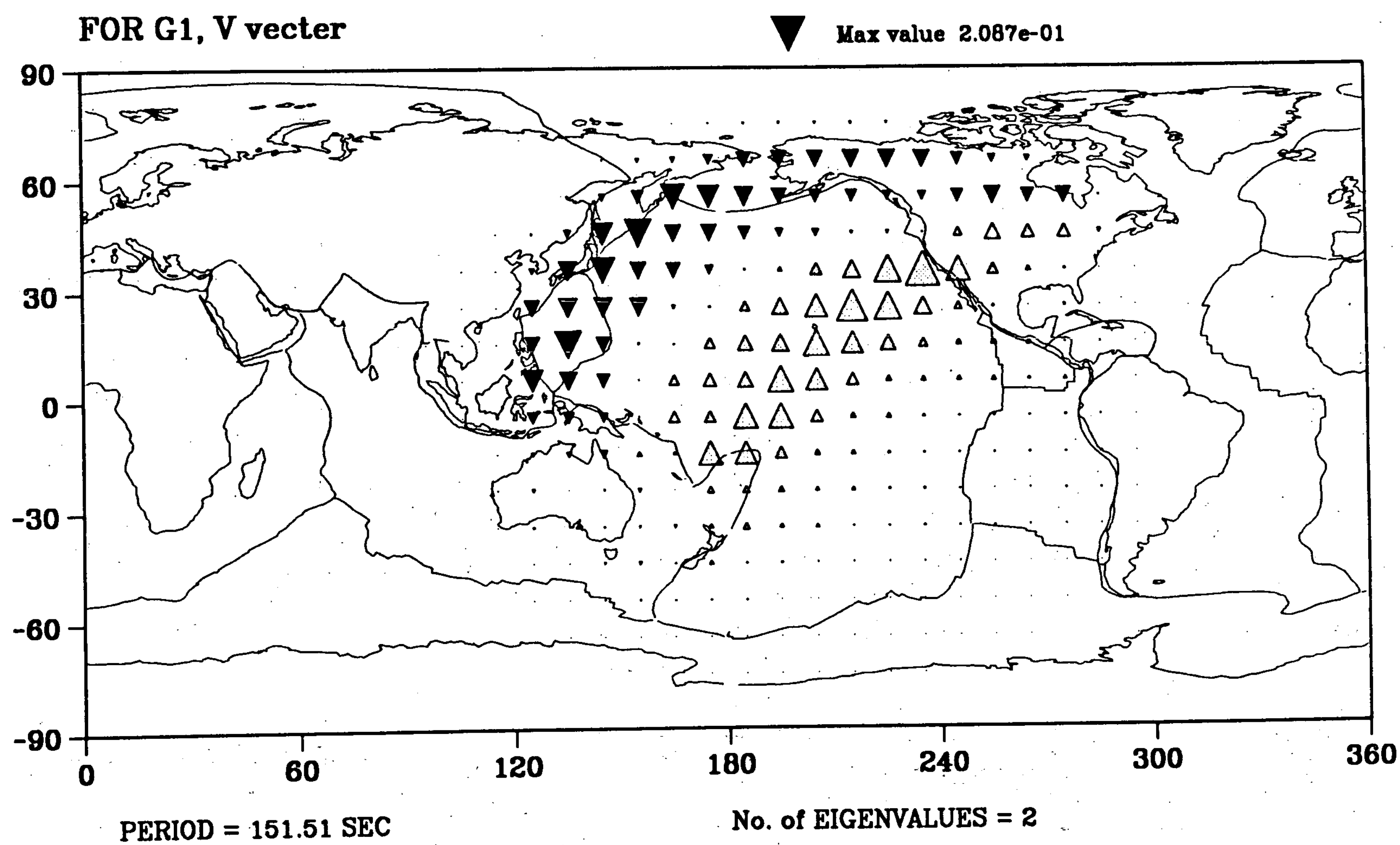

Figure 2b: Eigenvector of the second largest singular value.

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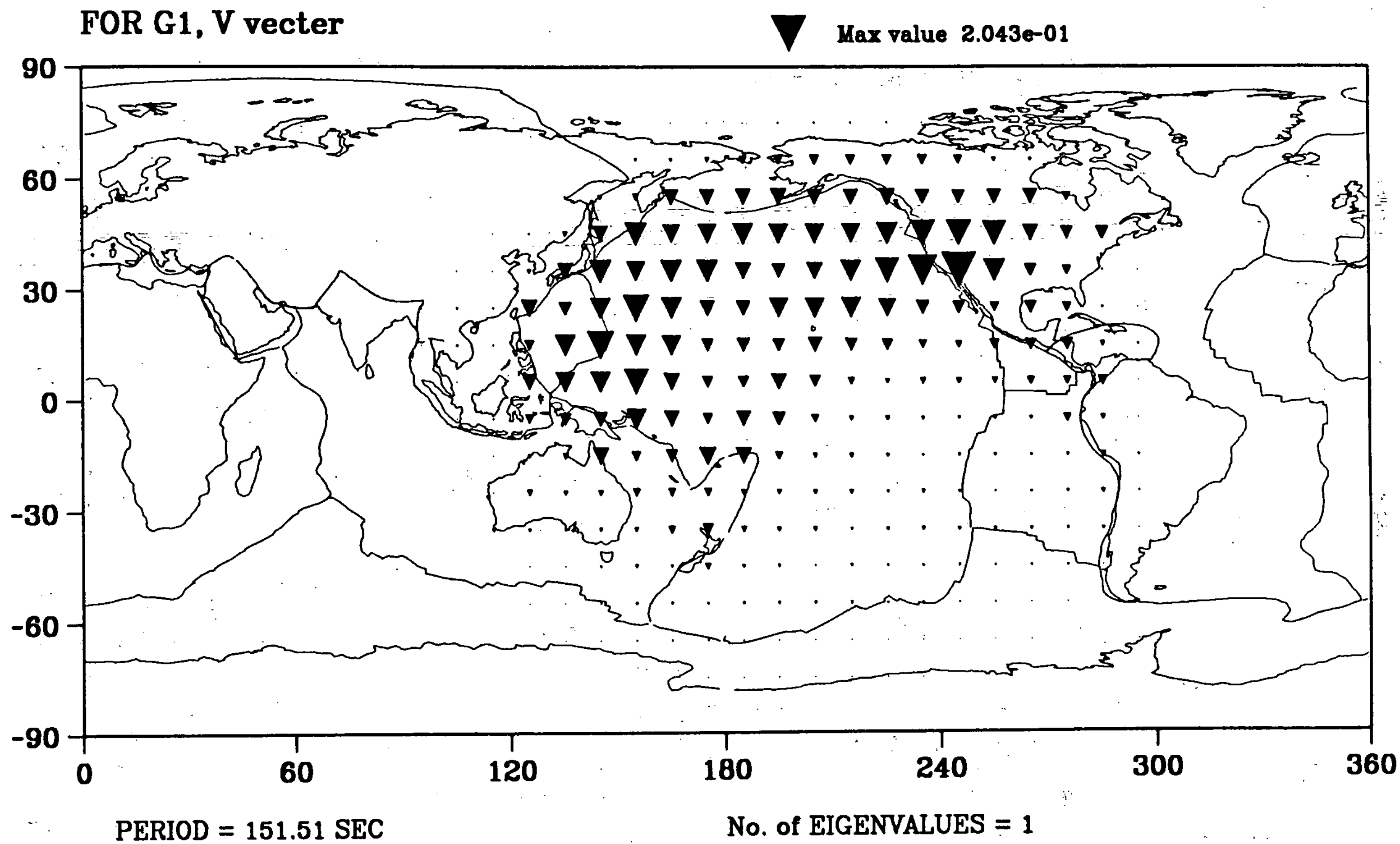

Figure 2a: Eigenvector of the largest singular value, which gives the average velocity in the northern half.

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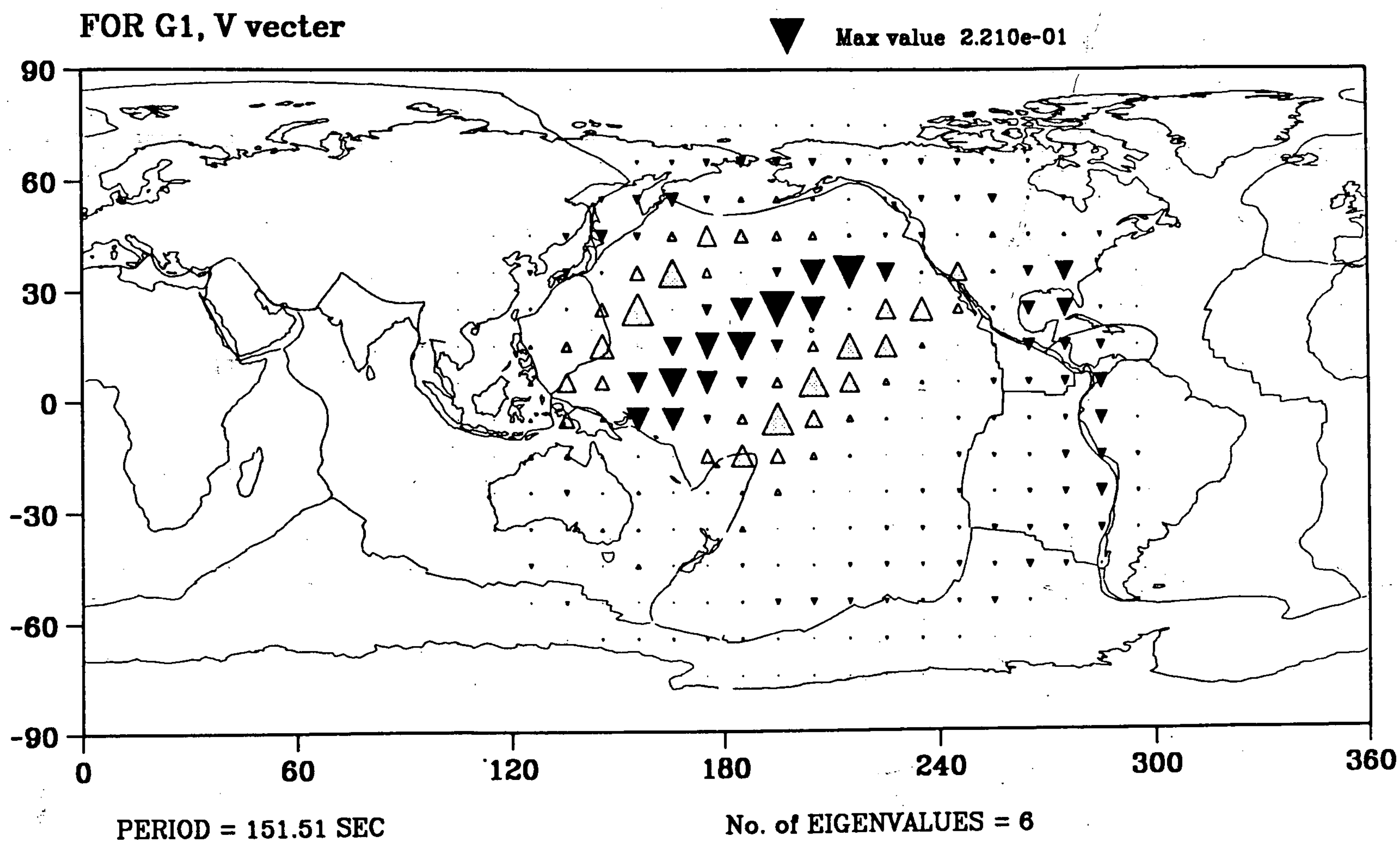

Figure 2c: Eigenvector of the sixth largest singular value.

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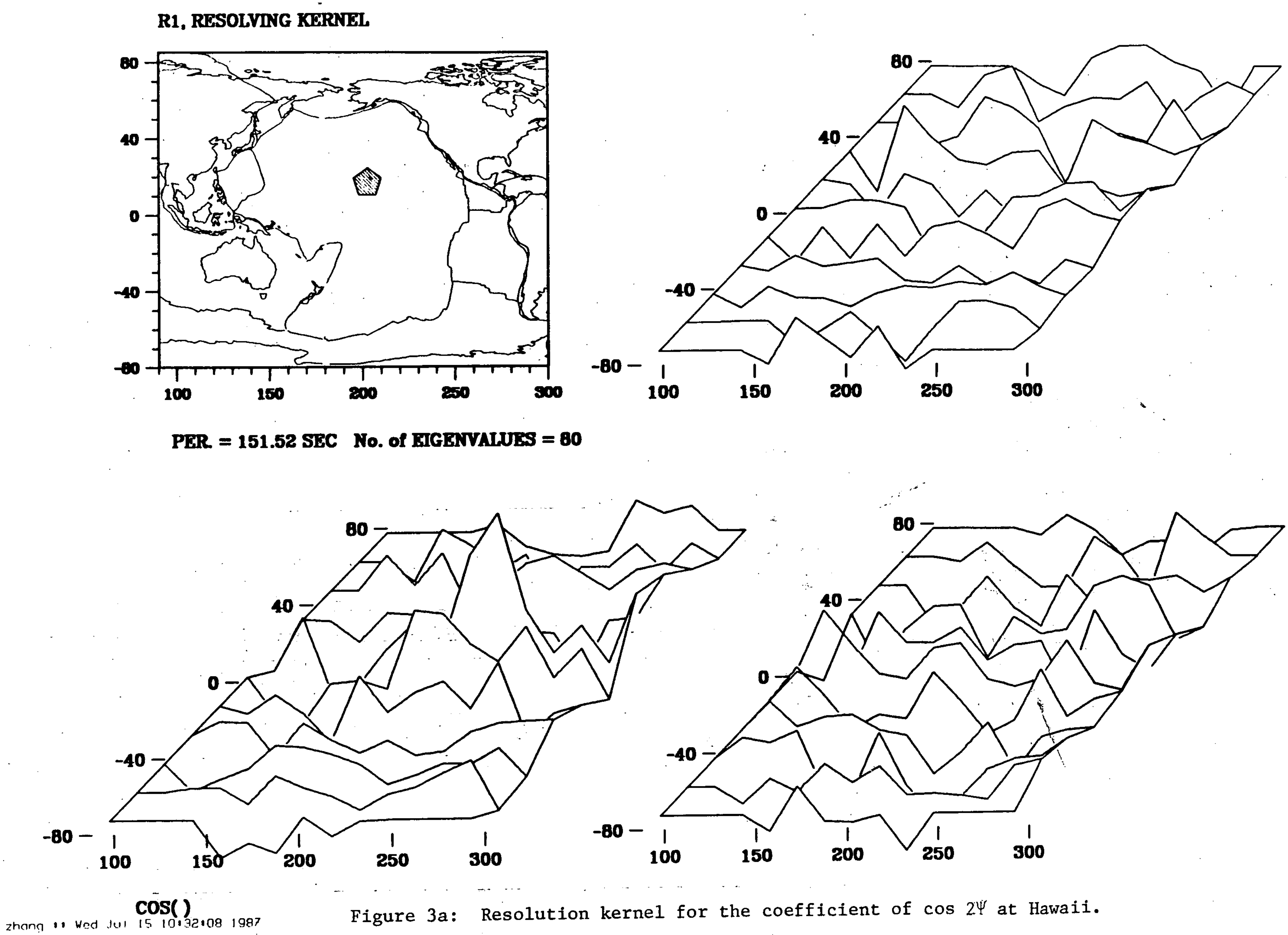
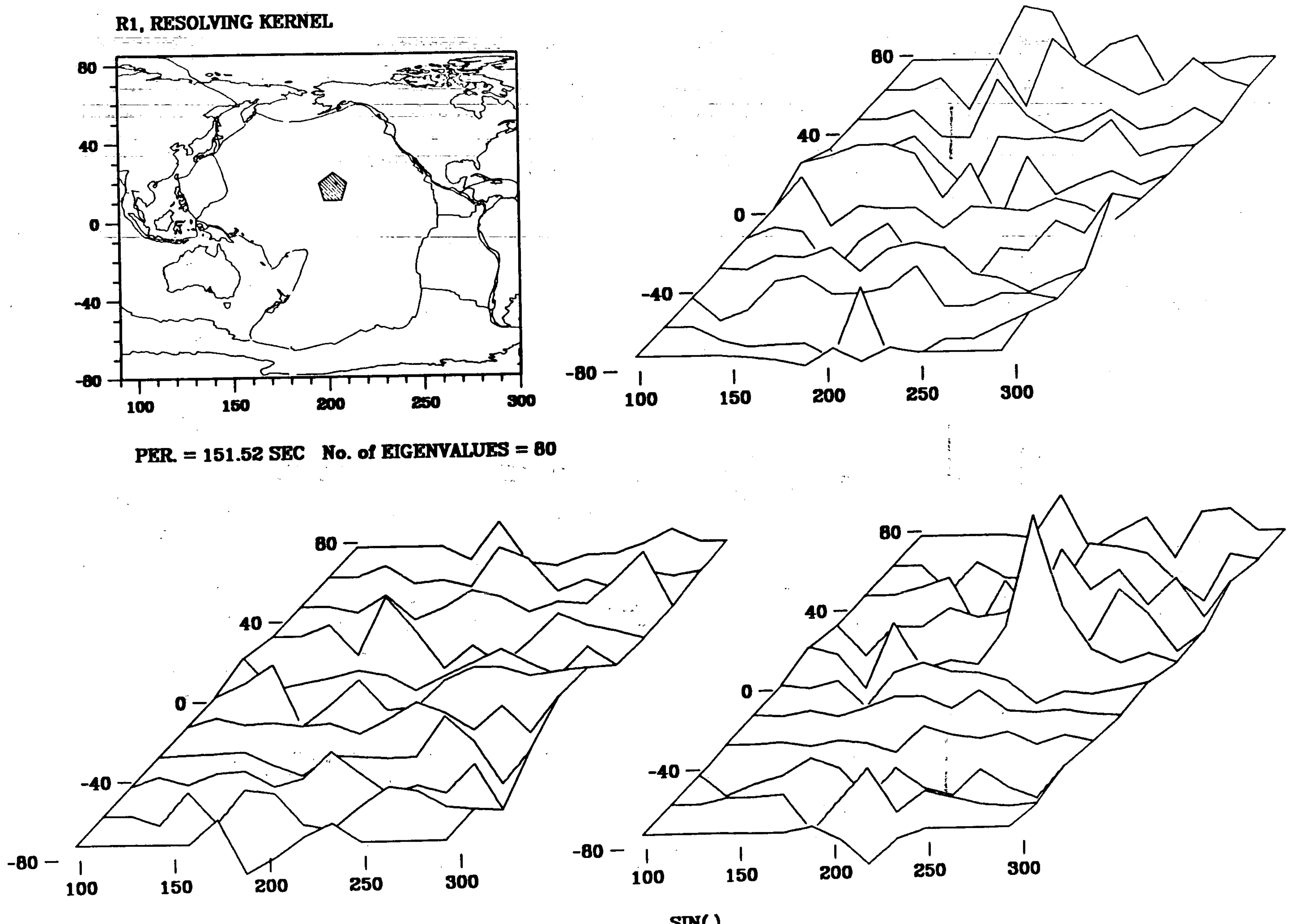


Figure 3a: Resolution kernel for the coefficient of cos 2Ψ at Hawaii.



zhang :: Wed Jul 15 10:41:47 1987 Fig. 3b: Same with 3a except for sin 3Ψ term.

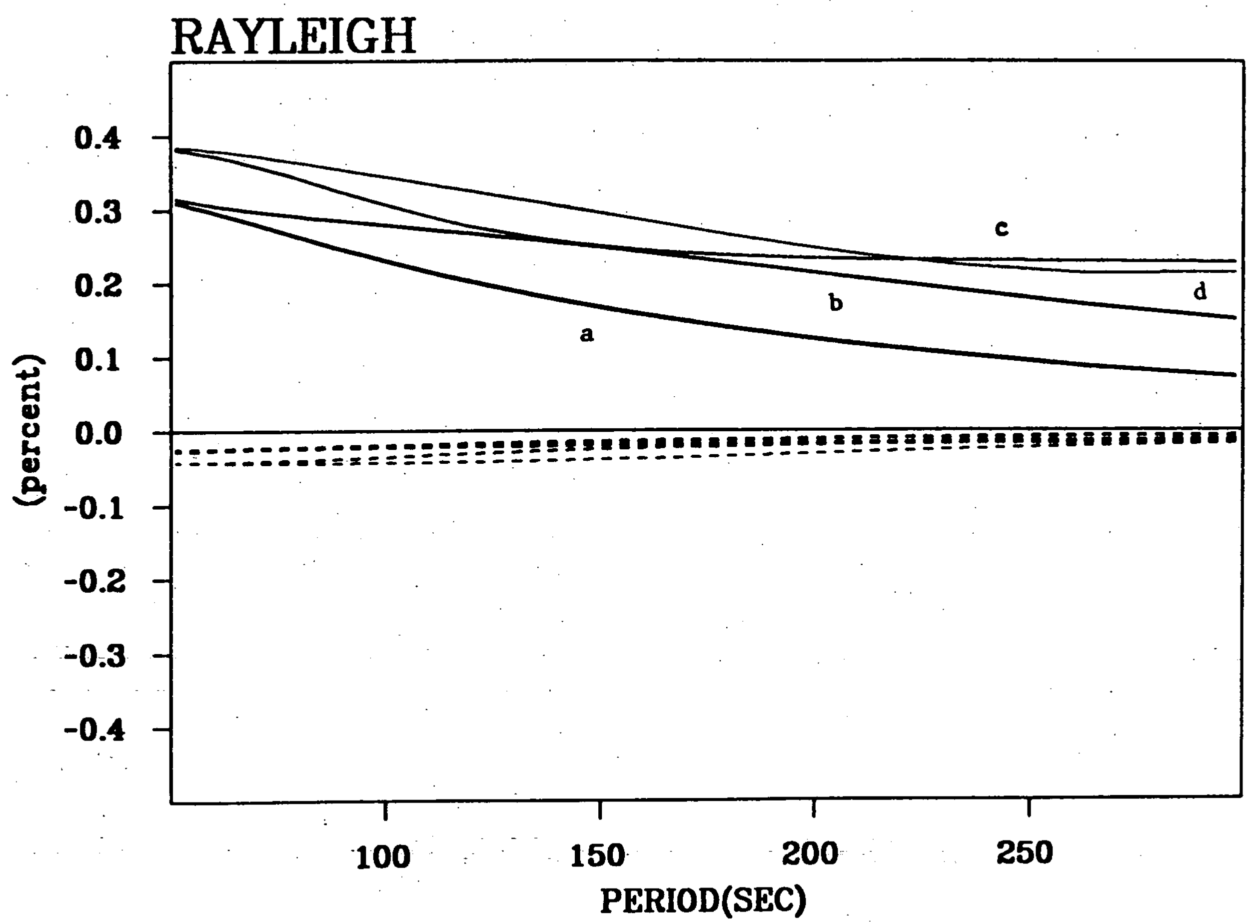


Figure 4a: Theoretical variation of the coefficients of cos 2Ψ (solid) and cos 4Ψ (dask) for Rayleigh waves. Models a, b, c, and d have different depth extent of anisotropy (see text).

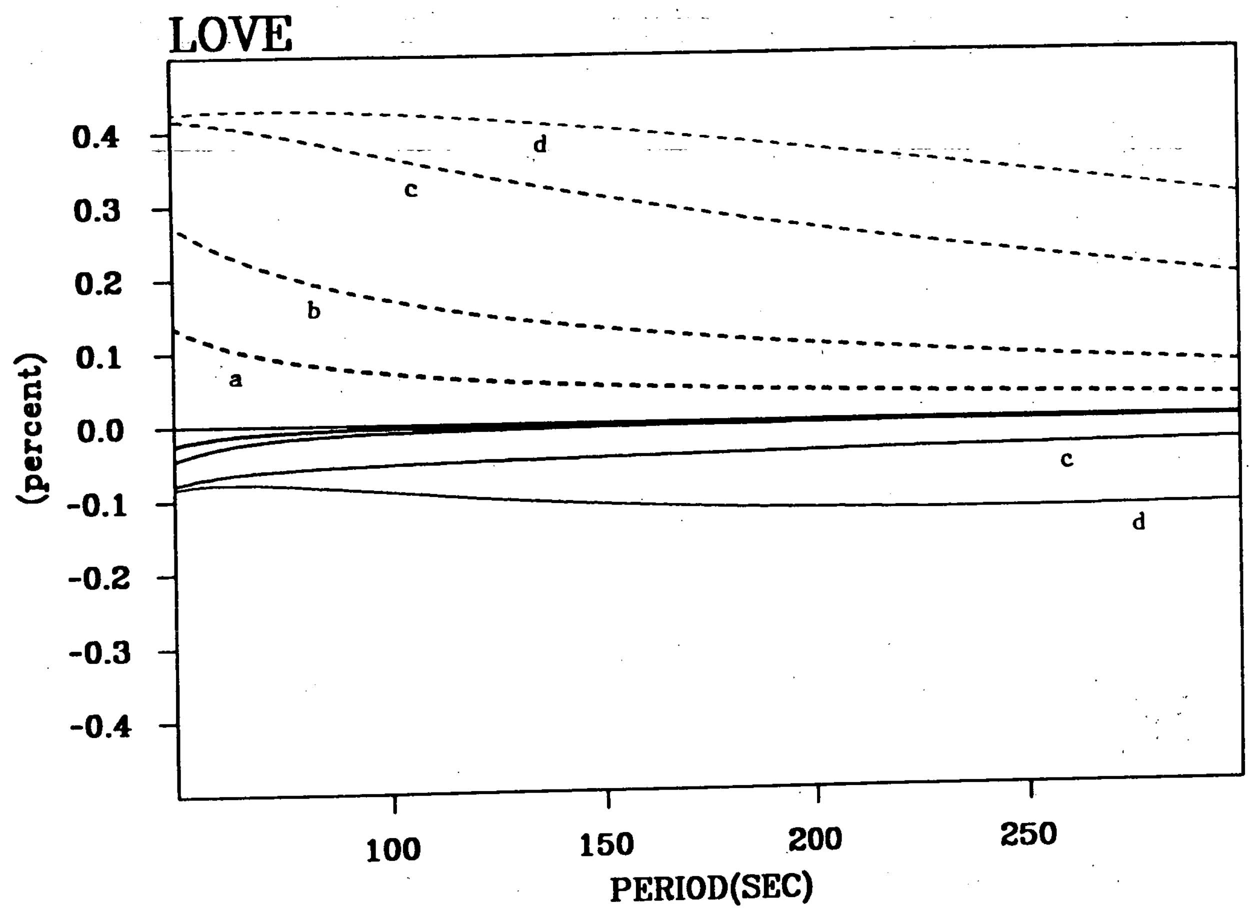


Figure 4b: Same with 4a except for Love waves.

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OCEAN FLOOR STATIONS AND THE MEASUREMENT OF REGIONAL SURFACE WAVE DISPERSION, ANISOTROPY, AND ATTENUATION

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Introduction

Studies of seismic surface wave attenuation since the mid-70's have revealed several complexities in the elastic and attenuative properties of the oceanic crust and upper mantle. One of these is an increase in lithospheric thickness with age (Kausel et al., 1974; Forsyth, 1975a; Yoshii, 1975; Yu and Mitchell, 1979). Although surface waves are generally consistent with this overall pattern across the Pacific, there are exceptions where velocities are not consistent with an age-velocity relationship (Mitchell and Yu, 1980; Nishimura and Forsyth, 1985). Another complexity is the presence of anisotropy, the magnitude and direction of which may vary with sea floor age. Anisotropy has now been observed in both the oceanic upper mantle and oceanic crust. Attenuation of seismic surface waves also vary systematically across ocean basins with older regions displaying less attenuation than younger regions. A brief review of work in these areas is given in the following paragraphs and benefits of these studies, especially for regional studies of anisotropy and attenuation from ocean bottom bottom seismographs, are discussed.

Anisotropy in the Oceanic Upper Mantle

Since the early work of Hess (1964), several refraction studies (e.g. Raitt et al., 1969); Shearer and Orcutt, 1985, 1986) have confirmed that anisotropy exists and that the fast direction of compressional waves is aligned with the direction of sea floor spreading at the time of formation of the lithosphere. This azimuthal anisotropy has also been observed using surface waves, first in young oceanic regions near the east Pacific Rise (Forsyth, 1975a), and later in older regions of the Pacific (Yu and Mitchell, 1979; Okal and Talandier, 1980; Mitchell and Yu, 1980). Differences in the magnitudes of anisotropy reported by Forsyth (2%) and Yu and Mitchell (<1%) suggest that the magnitude of anisotropy may vary with age of the sea floor.

Anisotropy has also been inferred for the Pacific upper mantle from the inability to explain both Love and Rayleigh waves by the same velocity model (Forsyth, 1975b; Schlue and Knopoff, 1977; Yu and Mitchell, 1979; Montagner, 1985), and from observations of anomalous surface wave particle motion (Kirkwood and Crampin, 1981a, b).

More recent studies of anisotropy using surface waves have employed higher modes, synthetic seismograms, or regionalizations using globe-circling paths. These include work by Anderson and Dziewonski (1982), Journet and Jobert (1982). Leveque and Cara (1983), Nakanishi and Anderson (1983, 1984), Tanimoto and Anderson (1985), and Nataf et al. (1986).

The presence of anisotropy in the oceanic upper mantle brings additional degrees of freedom into the process of modeling that region. Anderson and Reagan (1984), for instance, found that it was possible to satisfy Pacific surface wave data with a lithosphere which is much thinner than that inferred from isotropic inversions. To better constrain the thickness of the lithosphere it will be important to use higher-mode data (Okal and Jo, 1983; Lerner-Lam and Jordan, 1983; Leveque and Cara, 1985).

In studies where Love and Rayleigh wave data cannot be explained by the same model, it has been common to invert the data as if structures were isotropic but with different SH and SV velocities. This procedure is not strictly valid for anisotropic media (Crampin, 1976; Kirkwood, 1978; Anderson and Dziewonski, 1982; Mitchell, 1984) where surface waves do not, in general, separate into two independent families of Rayleigh and Love modes, but instead form a family of generalized modes having three-dimensional particle motion. In addition, compressional-wave velocities have a greater effect on surface wave dispersion in anisotropic media than they have in isotropic media. A further difficulty when using surface waves to study anisotropy is that it is difficult to resolve it to scales which will enable it to be related to details of the spreading history of the Pacific if changes in spreading direction occur.

Anisotropy in the Oceanic Crust

The crust is thought to grade downward beneath sediments from pillows to dikes to a gabbro layer. Stephen (1981) reported the first observations of anisotropy in the upper oceanic crust using borehole data in the western Atlantic. Later observations of anisotropy in the upper oceanic crust have been reported by White and Whitmarsh (1984), Shearer and Orcutt (1985, 1986), Little and Stephen (1985), and Stephen (1985). Anisotropy in the upper crust is usually attributed to preferentially oriented vertical fractures which were formed in the early stages of crustal development. White and Whitmarsh (1984) and Stephen (1985) reported that the cracks are oriented at an oblique angle to the spreading direction whereas Shearer and Orcutt (1986) indicate that they are parallel to the spreading ridge. These differences could be real and be related to laterally variable stress fields and crustal accretion processes which are likely to occur along ridges.

Christensen (1972) reported evidence for possible anisotropy in the lower oceanic crust from an examination of available seismic refraction results. He attributed that anisotropy to the preferred orientation of amphibole. Christensen and Salisbury (1975), in a review of seismic data, petrology of oceanic dredge samples, and laboratory measurements of seismic velocity pertaining to the lower oceanic crust concluded that layer 3 is likely to be composed of hornblende metagabbro underlain by normal gabbro. They noted that the deeper levels of layer 3 thicken for 40 m.y. and attribute that thickening to off-ridge intrusion fed from underlying anomalous mantle. Lewis and Syndsman (1979), using seismic refraction data in the Cocos plate, also conclude that the crust thickens with age, but propose that this thickening and coincident reduction in velocity which they observed, are produced by serpentinization of ultramafic rocks in the uppermost mantle.

The orientation and magnitude of seismic anisotropy is important in understanding the mode of formation and evolution of the oceanic crust. Several models for the upper crust (e.g., Cann, 1974; Kidd, 1977; Rosencrantz, 1982) and lower crust (e.g., Christensen, 1972; Christensen and Salisbury, 1975; Lewis and Snydsman, 1979) have been proposed. If the orientation of the anisotropy can be determined we may be able to distinguish between possible models which have been proposed.

Recent Surface Wave Observations

Our recent work suggests that the magnitude and direction of anisotropy varies with age in the Pacific. In young regions (0-20 m.y.), the fast direction of Rayleigh waves parallels the East Pacific Rise, a result which suggests that cracks in the upper crust are dominating surface wave velocities there. In the region of the Pacific between 20 and 50 m.y. in age the fast direction of Rayleigh waves parallels the trends of fracture zones and in the region 50-100 m.y. in age it is oblique to those zones. In the region >100 m.y. the fast direction is oriented in a more northerly direction. Our largest measured anisotropy is about 1.5% and occurs at periods greater than 50 s in the region 20-50 m.y. in age. In regions of the Pacific greater than 50 m.y. in age the average magnitude of anisotropy we observe is 0.5% or less, but these small values may simply mean that we have averaged over regions of the Pacific where the directions of anisotropy have changed with change in spreading directions.

Surface Wave Attenuation

Surface wave attenuation measurements have indicated that the anelastic properties of the oceanic upper mantle vary with age, Q values increasing with age of the lithosphere (Canas and Mitchell, 1978). Studies of surface wave attenuation will probably continue to require fairly long paths. This is necessary because surface waves attenuate very slowly except in regions of very low Q. The primary difficulty in these measurements is the large errors, both random and systematic, which are associated with surface wave attenuation measurements. It is believed that these uncertainties are caused by focussing, multipathing, scattering, and mode conversion which are produced by lateral variations in structure along the path of travel of the waves.

The Need for Ocean Bottom Stations

In the review of results described above, we have been restricted to averages over large regions using island stations. There are likely to be regional variations in seismic velocity, anisotropy, and attenuation which occur over scales which are much smaller than those presently amenable to study. These may occur because of anomalous processes associated with downgoing slabs or hot spots, by changes in spreading direction (Menard, Atwater, 1968), and first-order differences among spreading centers (Bratt and Purdy, 1984).

In order to study these small-scale regional variations and develop models for the formation of the crust and upper mantle we must develop the capability to study the seismic properties of the oceanic crust and upper mantle in some detail. This will require developments in two areas. First, we need to develop tools to study velocity structure and anisotropy on a near-regional scale at relatively high frequencies. B. Mandal and I have recently developed computational methods for computing complete synthetic seismograms for fairly generalized anisotropy. Thus, for the first time, we will be able to study the effects of anisotropy on amplitudes, signal durations, and wave forms at short distances where adverse effects of lateral heterogeneity can be minimized. Second, we will need better station coverage of the oceanic regions. This is necessary both because we need to sample several portions of the oceanic crust and upper mantle at relatively short distances and because our use of short paths and small travel-times will require that the earthquake hypocenters be located quite well.

Islands, on which present seismic stations are situated, are anomalous structures, as far as surface waves are concerned, and may overlie deep-seated anomalies which could affect surface waves at very long periods. They are likely to adversely affect results of both anisotropy and attenuation studies, the former because of changes in travel-times by anomalous structures which may be misinterpreted as anisotropy, and the latter because anomalous amplitudes may be produced by lateral changes in structure. Thus it would be advantageous for studies using surface waves to have a number of stations which are located on the sea floor in regions where velocity structure is reasonably laterally homogeneous.

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THE APPLICATION OF OCEAN FLOOR STATIONS TO TSUNAMI PREDICTION

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In many ways, Hawaii occupies an unique location within the Pacific Basin. Second only to Japan, it is the area most frequently impacted by destructive tsunami, with these generally propagating across great distances of ocean to reach the Islands. An analysis of historical tsunami data indicated that every tsunami in the Pacific Ocean that has been destructive outside the near source region has also been destructive in Hawaii. Conversely, any tsunami that has not been destructive in Hawaii has either been non-destructive or destructive only in the near-source region. South America, Alaska, and the northwest Pacific have historically been the source of destructive Pacific-wide tsunami. Tsunami generated in Central America, the northeastern Pacific, the western Pacific, and the southwestern Pacific may have been locally destructive, but posed no threat to Hawaii, and, by correlation, with other distant locations in the Pacific.

Organizational Structure of the Tsunami Warning Service

The National Weather Service operates and administers two tsunami warning centers. The Pacific Tsunami Warning Center (PTWC), located at Ewa Beach, Hawaii, has the threefold responsibility as the U.S. National Tsunami Warning Center, as the operational center for the International Tsunami Warning System in the Pacific, and as the Hawaii Regional Tsunami Warning Center. The Alaska Tsunami Warning Center, located at Palmer, Alaska, has responsibility as the Alaska, Canada, and U.S. West Coast Regional Tsunami Warning Center.

Data Acquisition Network

As operational center for the Tsunami Warning System of the Pacific, PTWC relies on a network of international seismological stations which provide critical seismic data either in real-time or in near real-time. These stations are listed in the recently published Eleventh Edition of the Communication Plan for the Tsunami Warning System. Through a cooperative effort with USGS/NEIC, PTWC records selected mainland and Alaskan seismic stations in real-time via a commercial satellite data circuit connecting PTWC with USGS/Menlo Park, with landline data communications to NEIC in Golden, Colorado.

For monitoring sea level data, PTWC depends upon an international data exchange with a network of tide observers throughout the Pacific Basin. Many tidal stations provide sea level data upon receipt of tide queries from PTWC. In addition, over the last three years the Pacific Tsunami Warning Center has participated in a joint effort in establishing a satellite sea level network at remote stations in the Pacific Basin. This network continues to expand, but presently includes 31 stations in South America and the Southwest Pacific, with sea level data transmitted automatically via the Geostationary Operational Environmental Satellite (GOES). This Pacific Satellite Sea Level Network now provides PTWC with the operational capability of receiving and analyzing sea level data from remote stations within 3-5 minutes of data transmission and, as such, constitutes a greatly improved data communications capability as well as tsunami evaluation capability.

Tsunami Forecasting

The evaluation of an earthquake with regard to location and size provides the initial forecast as to the probable tsunamigenesis of an event. Using present forecasting procedures as the National Tsunami Warning Center and the operational center for the Tsunami Warning System of the Pacific, a Tsunami Watch is issued for any coastal event exceeding a Richter magnitude 7.5 (or 7.0 for the Aleutians), based on the probability that a tsunami may have been generated. The Watch includes all areas that might be impacted within a 6-hour tsunami travel time, with urgent action being suggested for those areas within a 3-hour tsunami travel time. When wave confirmation has been received that a potentially destructive tsunami has been generated, a Tsunami Warning is issued.

It is well known that a tsunami is a very complex wave system. Though extensive benefits have been achieved through numerical modeling techniques for tsunami generation, propagation, and coastal runup, the most reliable forecasting technique is in the determination of tsunami travel-times and the computation of estimated times of Arrival (ETA's) for particular locations. For a Tsunami Warning, the Pacific Tsunami Warning Center routinely computes ETA's for 62 selected tidal stations around the Pacific.

In addition to predicting tsunami arrival times, applied research efforts are continuing at PTWC to better predict the generation of a tsunami using only an evaluation of the earthquake location and size. This is being accomplished through an intensive and detailed study on an area-by-area basis of each region of the Pacific. This includes an analysis of the tectonics of each area as well as documentation of the historical data available.

The ultimate goal is to develop a predictive evaluation capability for coastal impact throughout the Pacific. Though it is recognized that coastal impact is extremely variable and unpredictable, even over short distances, it is still possible to categorize tsunami within certain ranges of coastal impact. Just as operational procedures are based on some threshold value of earthquake size, tsunami thresholds can also be developed based on the predicted threshold impact along a coastal area. For example, if a particular threshold is established for Hawaii, such as a maximum of 0050 cm or 0100 cm, then tsunami activity below that threshold does not constitute a destructive potential and coastal evacuation is unnecessary.

In summary, the applied research being conducted at the PTWC relating to a tsunami discrimination scheme is focused at establishing the tsunami potential of each area of the Pacific Ocean, both with regard to near-source and far-field impact. Quantitative threshold values can then be established for minor tsunami that pose no threat, and for those that pose only a near-source threat. Rather than attempting to predict maximum tsunami impact, PTWC is attempting to establish minimum thresholds above which a tsunami poses a destructive potential regardless of how much maximum tsunami activity may exceed that threshold. Present operational procedures require that if an earthquake exceeds a Richter magnitude of 7.5 (Ms), a Tsunami Watch is issued regardless of how much the earthquake exceeds that threshold. In like manner, PTWC is attempting to better define the tsunami threshold for which a Tsunami Warning should be issued, as well as better define those areas of the Pacific for which the tsunami is predicted to be destructive.

Operational Requirements for an Ocean Bottom Station

On an operational basis, earthquake detection is the first phase involved in the provision of tsunami warning services. This is immediately followed by earthquake evaluation in terms of location, size, and ground displacement. Actual tsunami detection and evaluation are the final operational requirements.

Since PTWC's fundamental operational requirement is for a rapid response, both seismological data and oceanographic data must be available from a network of stations around the Pacific Basin on a real-time or near real-time basis. This requirement would also apply to any ocean bottom stations for the data to be useful for operational applications. Broad band seismological instrumentation provides the promise of an improved evaluation of earthquake parameters, including size and ground displacements. This information has important applications in relating the probability of tsunamigenesis associated with an earthquake.

In addition to requiring real or near real-time data, it is also strongly recommended that multiple sensors be deployed for any ocean bottom installation. Deep ocean pressure gauges should be an integral part of the system to provide sea level data to monitor tsunami activity. To determine wave directionality, multiple pressure gauges must be deployed over an extended grid.

The final consideration is that of location of an ocean bottom observatory. Any data obtained will either be applicable to near-source evaluation or far-field evaluation. For near-source applications the station must be located within a potential tsunami source region, namely an active seismic zone near the coast. Japan provides an example, and the Japan Meteorological Agency has deployed two networks of ocean bottom seismometers along the south coast of Honshu in anticipation of the forthcoming Tokai earthquake and tsunami.

For application to far-field evaluation, i.e., teleseismic earthquakes or distant tsunami, any location within the Pacific Basin may be considered, but particular areas have various characteristics. The southwest Pacific is most active in terms of the number of large earthquakes, but with a low probability for generation of destructive Pacific-wide tsunami. South America poses the problem of large earthquakes and destructive tsunami, but the frequency of occurrence is low. The north and northwest Pacific are not only zones of major earthquake activity, but also source regions for destructive distant tsunami.

Hawaii is a special place, and special consideration should be given to the Island of Hawaii. The following items are of interest:

a.) Though the frequency of local tsunami generation is low, it is an active seismic area with undersea volcanism occurring off the southeast coast at Loihi volcano.

b.) Both PTWC and USGS's Hawaiian Volcano Observatory could use the data on an operational basis as well as provide operational support. Numerous research applications could also be supported.

c.) Data communications could be either via satellite transmission from an existing stationed buoy or via submarine cable similar to Japan's system.

d.) The Big Island offers the advantage of monitoring frequent local seismic activity as well as being in an optimal location to monitor teleseisms, with all earthquakes in the Pacific Basin being within 103 degrees of Hawaii.

e.) A station in Hawaii can monitor more tsunamigenic earthquakes and, given the inclusion of deep ocean pressure gauges, record more tsunami than any other location in the world.

It is therefore recommended that initial deployment of an ocean bottom observatory be considered for the north Pacific, with Hawaii being considered as the site for the installation.

RESOLUTION OF CHANGES IN SLAB DIP VIA BODY WAVE INVERSION: IMPORTANCE OF PACIFIC OCEAN BASIN SEISMIC STATIONS

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Abstract

Underthrusting earthquakes are generated at the coupled interface in subduction zones. There may be a geometric kink in the subducting slab at the downdip edge of the coupled zone, hence it is important to resolve changes in focal depth and fault dip for under thrusting events. Teleseismic P and SH phases can be inverted to determine the moment tensor; good resolution depends on good sampling of the focal sphere. For the circum-Pacific subduction zones, the proper sampling of a critical sector of the focal sphere requires seismic stations in the Pacific Ocean basin.

Introduction

This brief report presents just one argument for the advantages of seismic stations scattered throughout the Pacific Ocean basin. This argument is motivated by a particular geophysical problem: resolution of changes in slap dip at the downdip edge of the coupled zone. Although this report does not present a comprehensive argument for oceanic seismic stations, it may be useful to characterize their importance for a specific seismological study. The outline of this report is as follows: introduce the notion of slab kinks; discuss focal mechanisms and body waves; show the "circum-Pacific problem"; generate moment tensor resolution matrices with and without the Pacific Ocean stations.

Slab Kinks

It is well recognized that subducting slabs do not maintain a constant dip angle from the trench axis to their deepest extent. While the slab dip may change smoothly in some cases, Ruff & Kanamori (1983) suggested that abrupt changes in slab dip at about 40 km depth may occur in many subduction zones (see figure 1). The slab kink is expressed in the Wadati-Benioff zone seismicity, but since this kink may be associated with the change from interplate to intraplate seismicity, it is important to directly verify the slab kink by finding two or more underthrusting events that span the kink at a particular locale. There should be a correlation between deeper focal depth and steeper fault dip. This geophysical problem prompts a detailed discussion of fault dip resolution.

Underthrusting Focal Mechanisms & Body Waves

Underthrusting events in subduction zones typically show fault dips from 10° up to 40° or so. Slip vector orientation depends on the details of arc strike and convergence direction, but typically cluster about pure thrust. Figure 2 shows two "typical" underthrusting mechanisms with a fault dip of 20° (a) and 35° (b). With the assumption of little to no oblique slip, the dip of the shallow fault plane is directly given by the dip of the steep auxiliary plane. Given the steep take-off angles of teleseismic body waves, P wave first-motions are mostly sensitive to the orientation of the steeply dipping plane. Thus P wave observations in the stippled area of the focal sphere in Figure 2 offer the most information concerning fault dip. Of course, waveform inversion uses more information than just first-motions, but before we enter the realm of waveform inversion, lets look at the geographical location of the stippled region for circum-Pacific subduction zones.

The Circum-Pacific Data Gap

The focal mechanism in Figure 2 is the "generic" underthrusting mechanism for any subduction zone. The stippled region on the focal sphere represents the traditional teleseismic range (distance of 30° to 90°) with respect to take-off angle and an azimuthal range of 100° centered about the direction perpendicular to arc strike. This focal sphere region is then easily translated into a geographic region for any particular subduction zone. Figures 3 through 7 show the geographic location of the stippled focal sphere patch for several circum-Pacific subduction zones. It is simply a geographical fact that the Pacific Ocean basin is large (more than 90° across) and thus the Pacific Ocean basin occupies the teleseismic distance band over a large azimuthal range for a vast number of subduction zone earthquakes. Needless to say, there are not many seismic stations in the Pacific Ocean. Just as frustrating, the occasional record that one obtains usually shows a high noise level. Our research group has performed detailed P wave studies of several large underthrusting events that rim the Pacific (e.g. Beck & Ruff, 1987), and we are always scrambling to find just one or two high-quality phases on the oceanic side of the focal sphere. Hence there is a intrinsic "data gap" for body wave studies of circum-Pacific earthquakes. Not only does the "data gap" degrade resolution for rupture process studies, but it can also affect moment tensor results.

P & SH Inversion for Moment Tensor Components: Resolution

A glance at Figure 2 shows that Pacific Ocean stations are critical to fault dip determination based on first-motion studies. How important are these stations for waveform inversion studies of focal geometry?

Waveform inversion for the moment tensor is now a well-established procedure for obtaining high-quality unbiased estimates of the focal mechanism (some examples are: Dziewonski et al., 1981; Kanamori & Given, 1981; specifically body wave inversion: Sipkin, USGS bulletins; Ekstrom & Dziewonski, 1986). Lets restrict our attention to teleseismic P and SH waveform inversion; this type of inversion seems well-suited for the simultaneous resolution of focal depth and mechanism. The circum-Pacific "data gap" is characterized by lack of data in the stippled region of the generic focal mechanism in Figure 2. Thus a resolution analysis for this one generic problem is applicable to many events that rim the Pacific.

Under the assumption that the focal depth and source time function are known, observed body waves can be linearly inverted to determine the moment tensor components. With no isotropic part, the moment tensor have five independent unknowns that can be arranged in several ways (see Aki & Richards, 1980). The forward problem is then: $Af=d$, where the data vector, d, consists of the observed seismograms stacked as one long column vector, A is the matrix of Green's functions convolved with the source time function, and f is the model vector with five components. Even though this problem appears to be highly overdetermined, some components of f may not be resolved. The normal equations for the least squares solution to the above problem are simply: $A^TAf=A^Td$, where A^T is A transpose and we define $B=A^TA$ as the 5x5 "inner product" matrix. Resolution difficulties are characterized by an ill-conditioned B matrix. To obtain a stable inversion result, it may be necessary to truncate the eigenvector expansion of B.

To calculate and analyze the B matrix it is only necessary to generate the set of Green's functions for a hypothetical distribution of seismic stations. A somewhat typical irregular station distribution is used (plotted on the focal sphere in Figure 2) with a total of six P waves and two SH waves. The focal depth is 30 km, and a source time function is chosen to mimic a "small" event, i.e. magnitude 6 to 6.5. The key point concerning the latter assumption is that ideally we want to invert seismograms that contain "higher" frequencies so that the depth phases (i.e. *pP* and *sP*) are distinct, yet the seismograms should <u>not</u> contain "higher" frequencies such that the assumed Green's functions are inadequate. The parameters used in this experiment produce Green's function with fairly distinct depth phases, hence we are near the high resolution limit.

The moment tensor component arrangement that I shall use consists of five double-couple earthquakes. With the "north" direction chosen to coincide with the "arc strike" direction in Figure 2, the order and double-couple type are as follows: (1) 45° dip-slip with NS striking nodal planes; (2) 45° dip-slip with EW striking nodal planes; (3) strike-slip with NS/EW vertical nodal planes; (4) 9° dip-slip with NS striking nodal planes; (5) 90° dip-slip with EW striking nodal planes. Given the above definitions, a pure dip-slip mechanism that strikes NS, such as those plotted in Figure 2, corresponds to nonzero values for only components #1 and #4. We shall return to this specification later.

Two B matrices are calculated. One uses all eight stations plotted in Figure 2 (i.e. assumes that high-quality Pacific Ocean stations are available), the other one just uses the four "landward" stations. As expected, there are resolution difficulties for both cases; the condition number for B with all eight stations is 43.2, while it is 267.1 for the B with four stations. For the eight station case, there is only one small eigenvalue, and it corresponds almost 100% to component #3 - the pure strike-slip component. If we eliminate this contribution to B, the remaining eigenvalues then give a condition number of 11.95 - the other components of f would be reliably determined. On the other hand, the two smallest eigenvalues of the B with only the landward stations must be eliminated so that the remaining eigenvalues yield a condition number of 12.86. Unfortunately, the eigenvectors associated with the two smallest eigenvalues make significant contributions to all four dip-slip components of the moment tensor. The elimination of the two smallest eigenvalues for the landward B case has significant impact on resolution.

From basic linear analysis, the model estimate, $\underline{f}$, is related to the "true" model f by: $\underline{f}=Rf$, where $R=B^{-1}B$ is the resolution matrix and B^{-1} is the matrix used as the B inverse. B^{-1} is calculated for the two different B matrices. For the eight station case, only the small eigenvalue associated with the strike-slip component is eliminated. Thus, the resolution matrix is essentially the identity matrix except the third component is zero (see Figure 8). For the four station case, B^{-1} is constructed with the two smallest eigenvalues eliminated; the impact on the resolution matrix is much more severe (Figure 8). For this latter case, it is clear that for the same level of stability, the moment tensor components are poorly resolved. To further illustrate this point, we can look at a particular case: specify the "true" f to be (1,0,0,1,0), which represents a pure dip-slip fault with a NS strike and fault dip of 22.5°. The "resolved" model for the eight station case is then given by the construction of the resolution matrix with the above f. The resultant model is plotted in the lower part of Figure 9 - the fault dip of 22.5° is recovered from the inversion (ignoring data variance of course). On the other hand, contracting the above "true" model with the resolution matrix for the four station case results in the "resolved" model plotted in the lower part of Figure 10. Not only are there spurious non-zero values for the other moment tensor components, but the ratio of f_1 to f_4 is no longer 1. If we were to ignore the spurious values of the moment tensor to extract a pure dip-slip mechanism with a NS strike, then the dip of the fault plane would be 33.7°, quite a significant deviation from the input 22.5°. Thus we can clearly see that body waves from the stippled region of the focal sphere should be included in the inversion if we are to obtain a stable and well-resolved value for fault dip.

Conclusions

The reliable determination of fault dip for circum-Pacific earthquakes is important for many different reasons; the geophysical application developed in this report is related to the slab kink near the downdip edge of the coupled zone. It is a geographic fact that seismic stations installed in the Pacific Ocean basin would offer the best coverage of a critical sector of the focal sphere for most subduction zone events. A formal resolution analysis shows that a particular station distribution, frequency pass-band, and inversion stability level, P and SH waves from the oceanic sites are essential to resolve fault dip.

Acknowledgments

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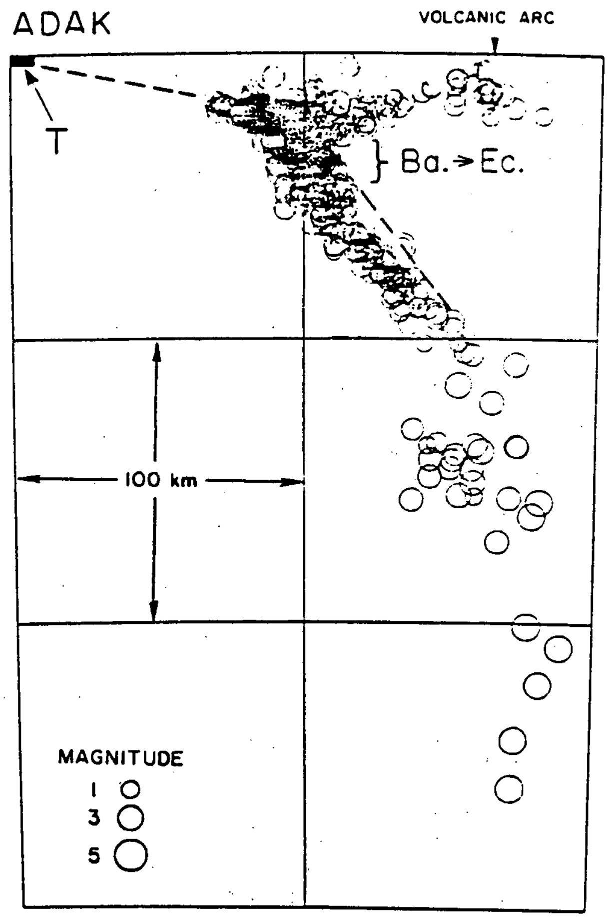

Figure 1. Cross-section of Wadati-Benioff zone seismicity in the central Aleutians (after Engdahl, 1977). The distinct kink at 30 to 40 km is quite apparent in the seismicity trends. Documentation of a change in fault dip of underthrusting events would corroborate the slab kink. Ba.>Ec. refers to the basalt to eclogite phase change.

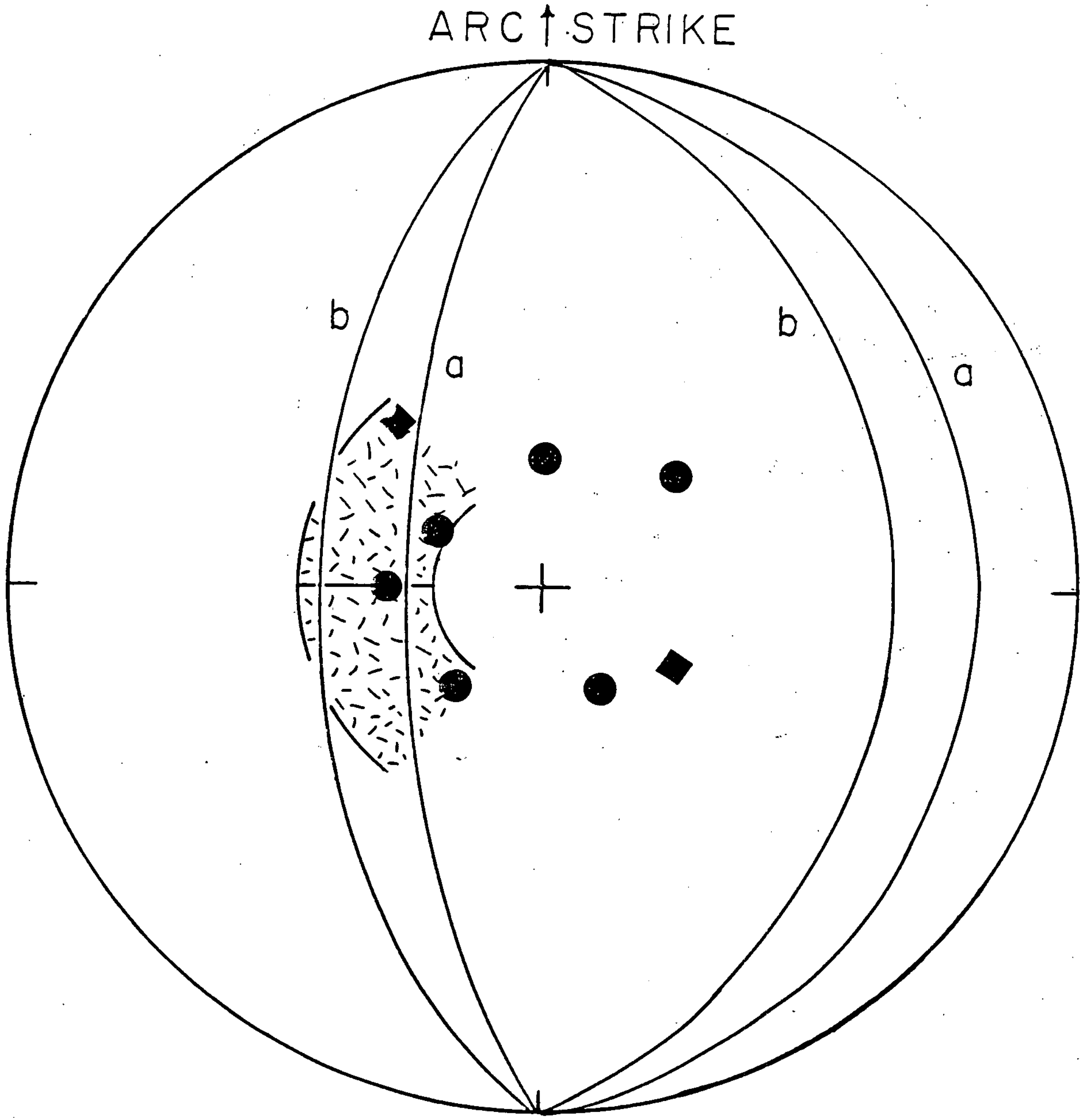

Figure 2. Generic focal mechanisms for underthrusting events . North coincides with the strike of the arc system, two different fault dips are shown, (a) dips 20°, while (b) dips 35°, with pure dip-slip. *P* wave observations in the stipled region on the focal sphere are quite sensitive to fault dip. This stipled region represents the teleseismic distance range and an azimuthal range of 100°. Station locations for the resolution study are also plotted, circles are *P* waves while squares are *SH* waves.

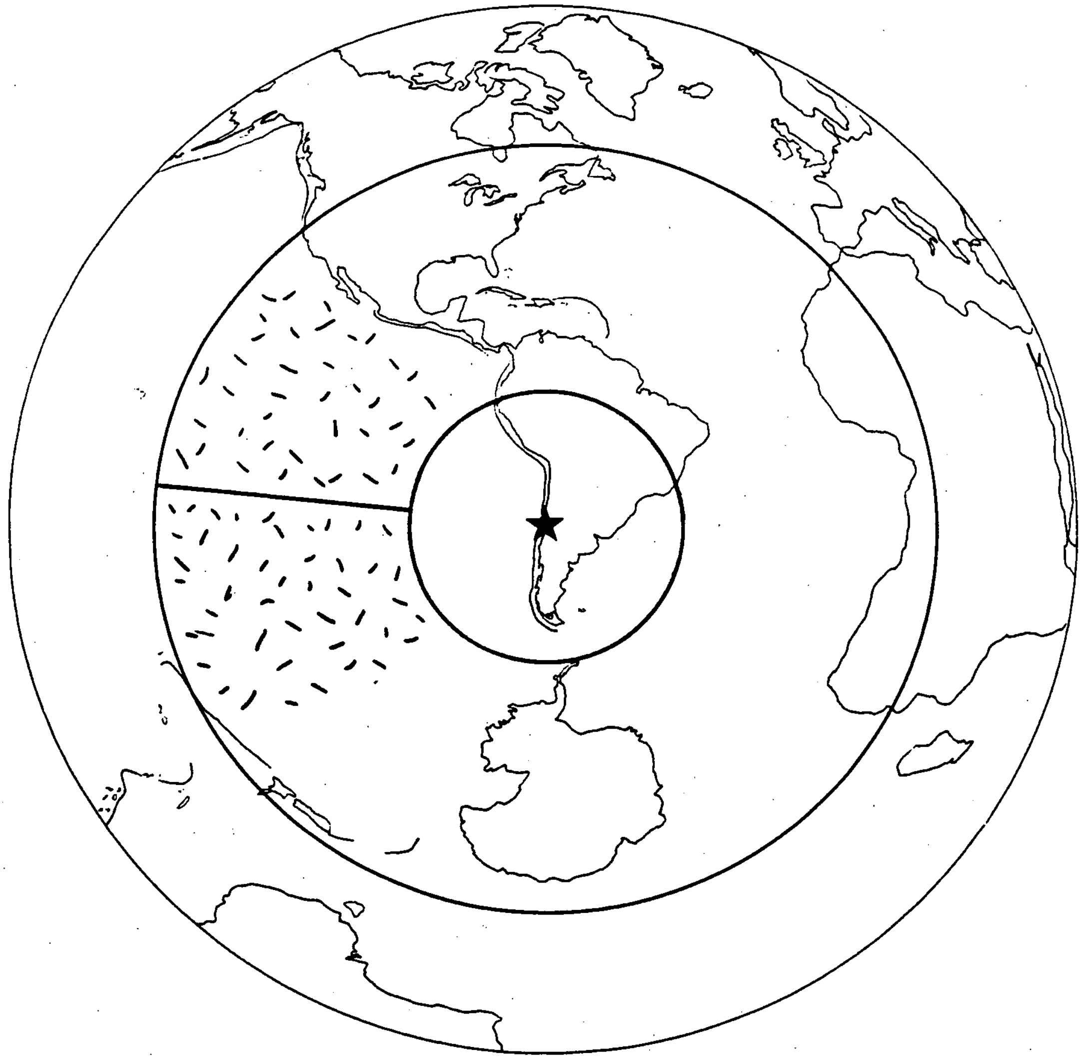

Figure 3. Azimuthal equal-distance global map centered on the central Chilean subduction zone. The outside map border is at a distance of 120°, the two heavy concentric circles show the distance range of 30° to 90°. The stipled region on the focal sphere in Figure 2 plots as the stipled geographic region for an underthrusting event in central Chile. If there are seismic stations located in the above stipled area, then the steeply dipping nodal plane will be well-determined.

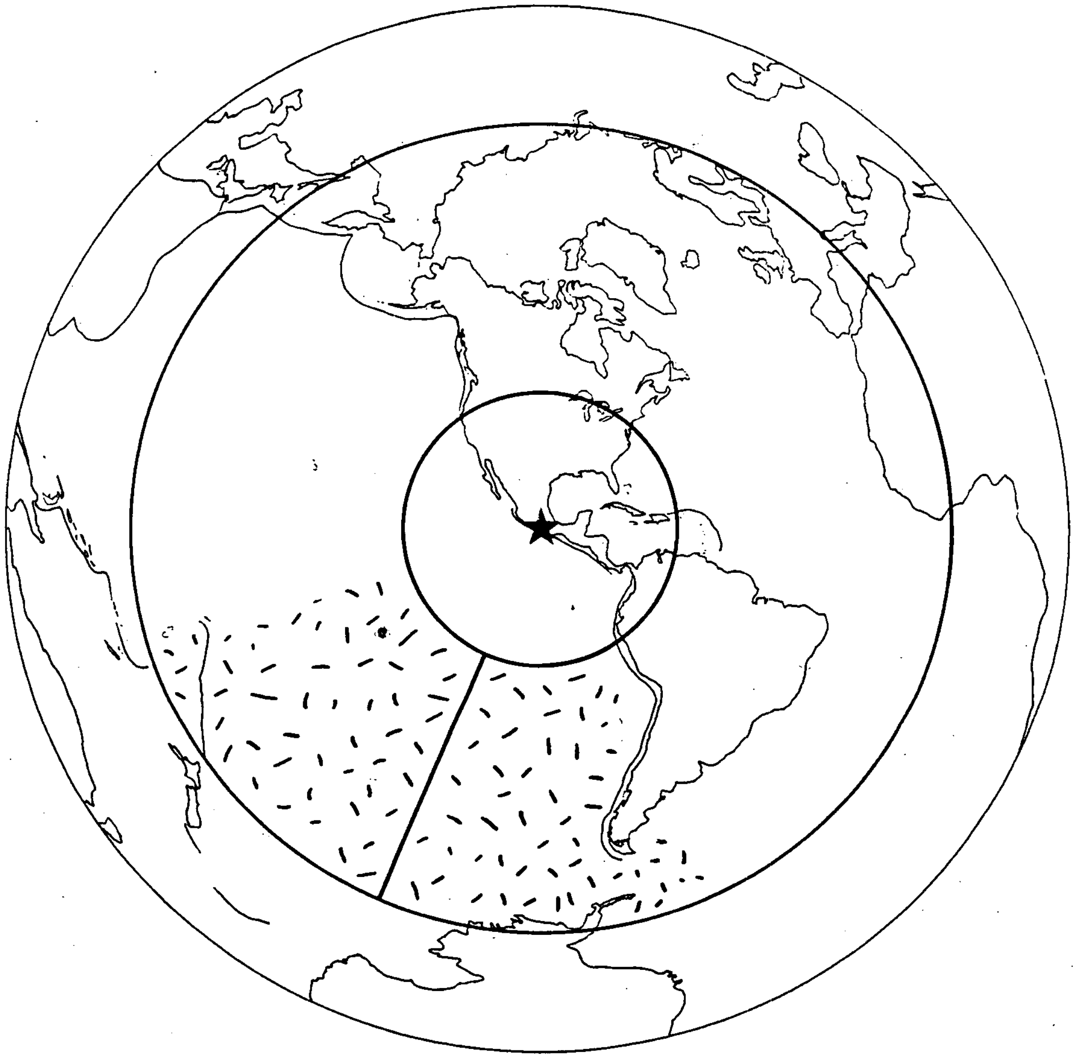

Figure 4. Same as Fig. 3, except centered on the Mexico subduction zone.

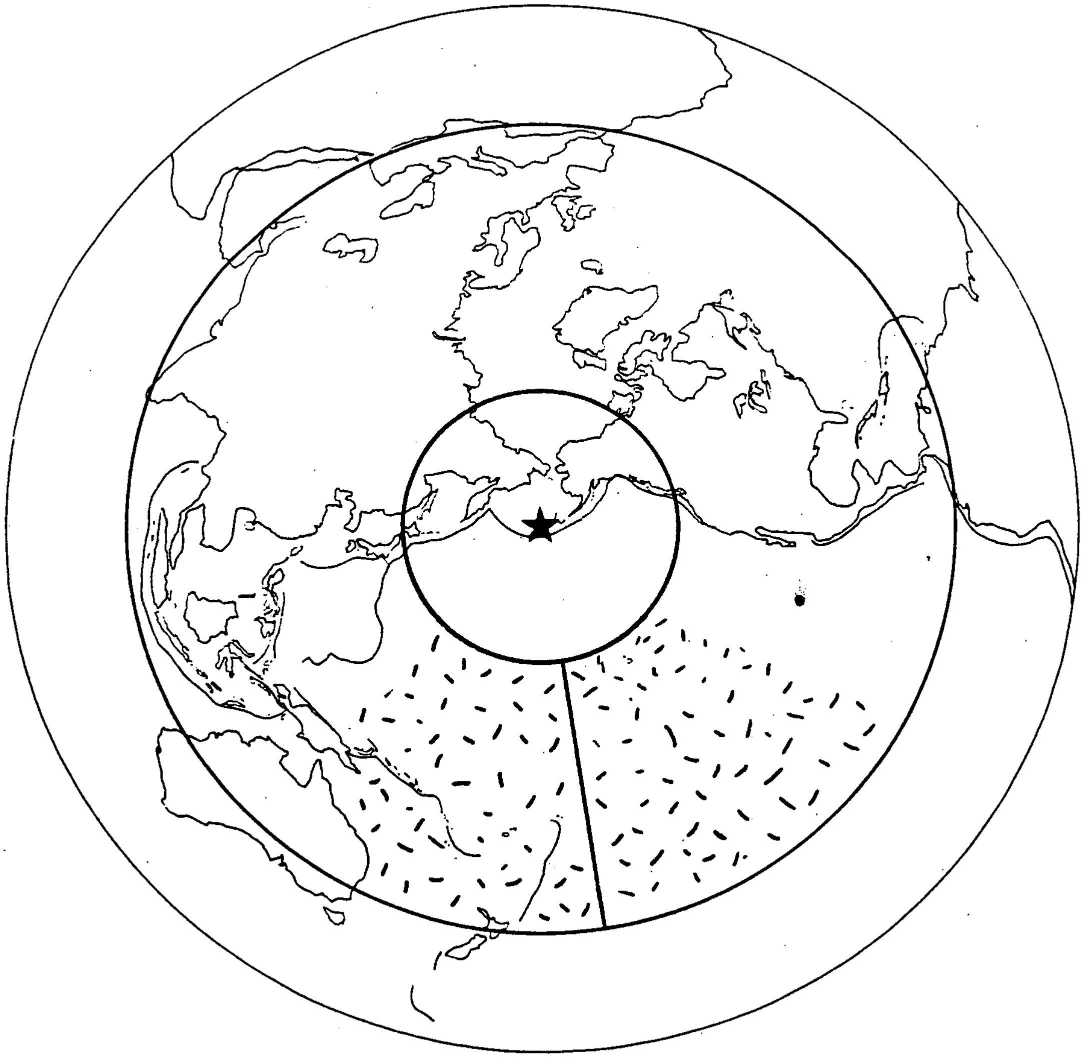

Figure 5. Same as Fig. 3, except centered on the central Aleutians subduction zone.

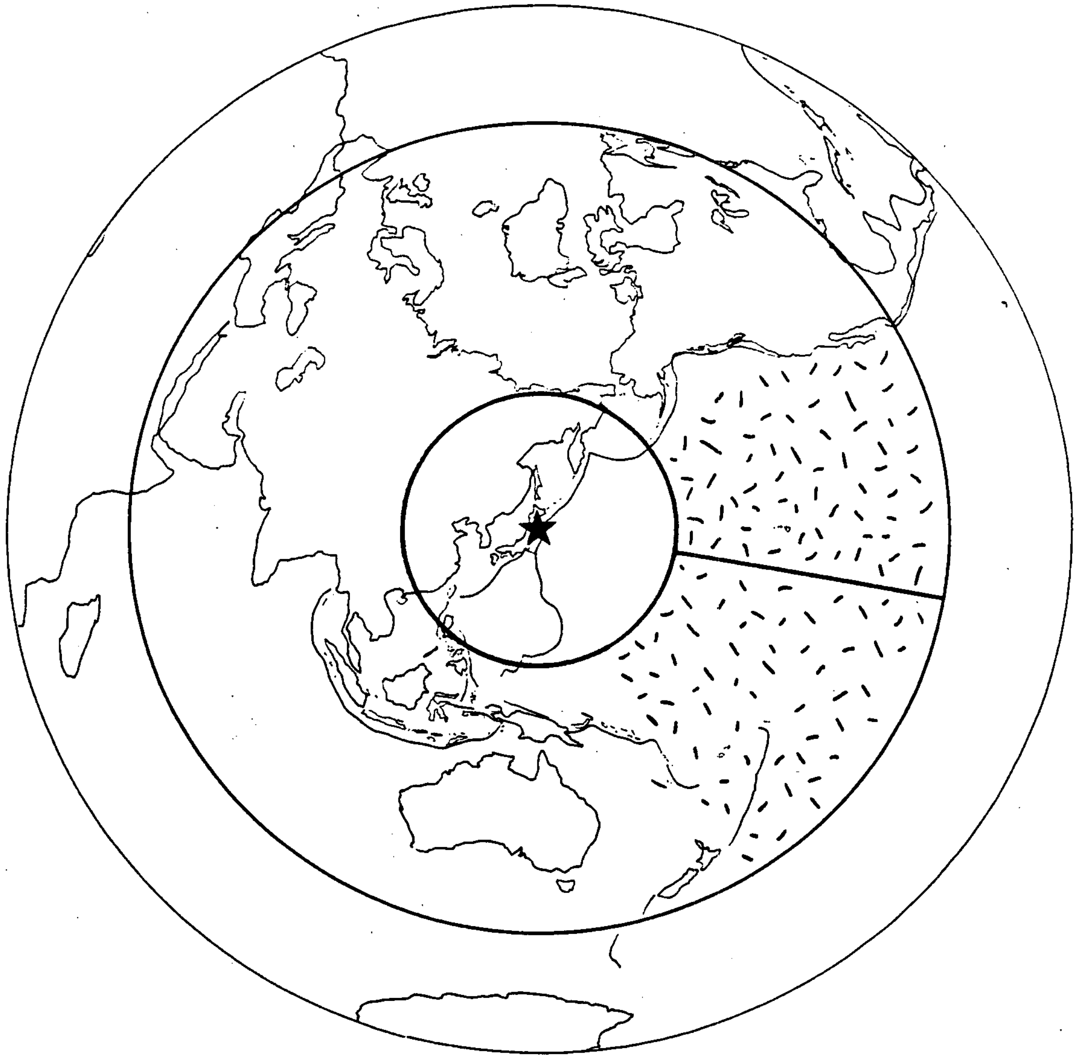

Figure 6. Same as Fig. 3, except centered on the Northern Honshu subduction zone.

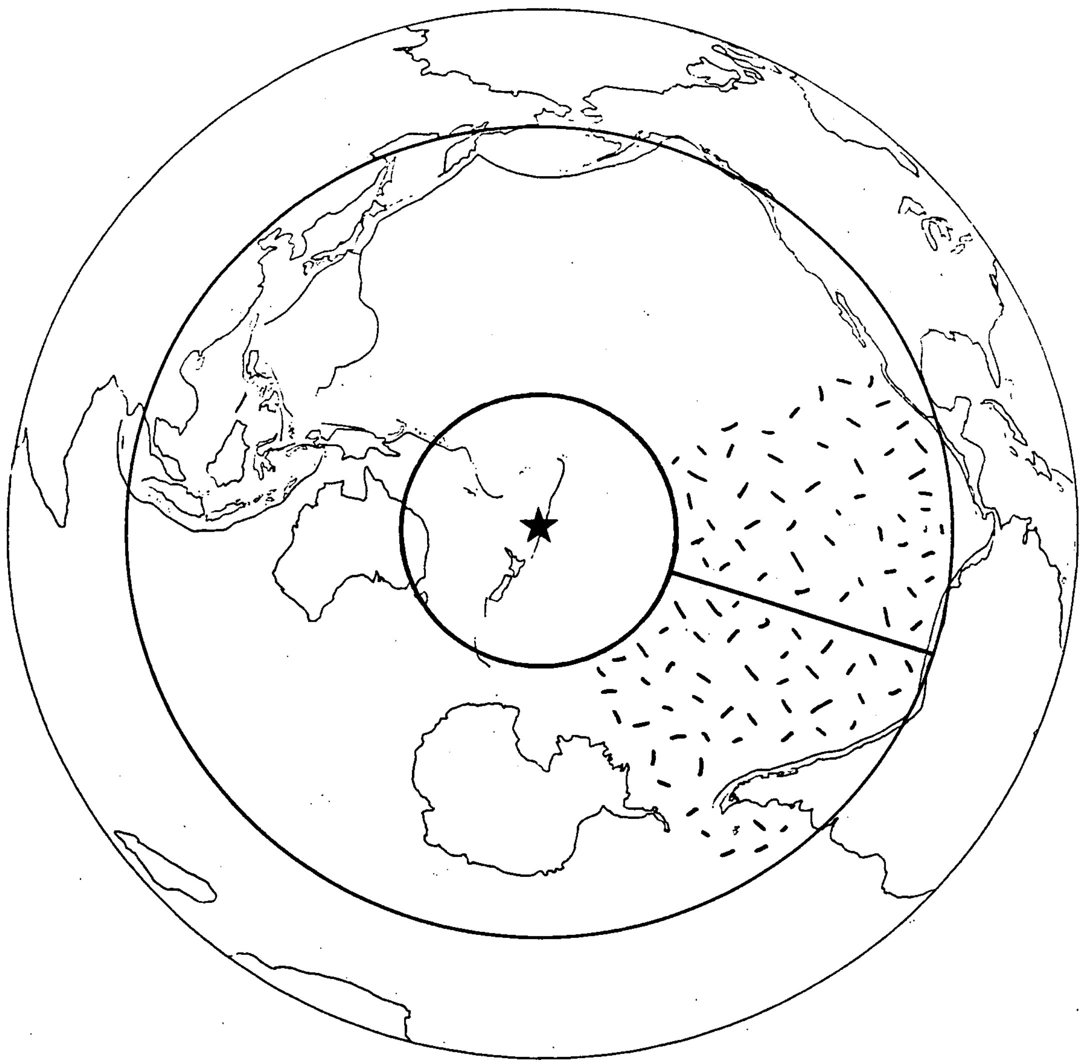

Figure 7. Same as Fig. 3, except centered on the Kermadec subduction zone.

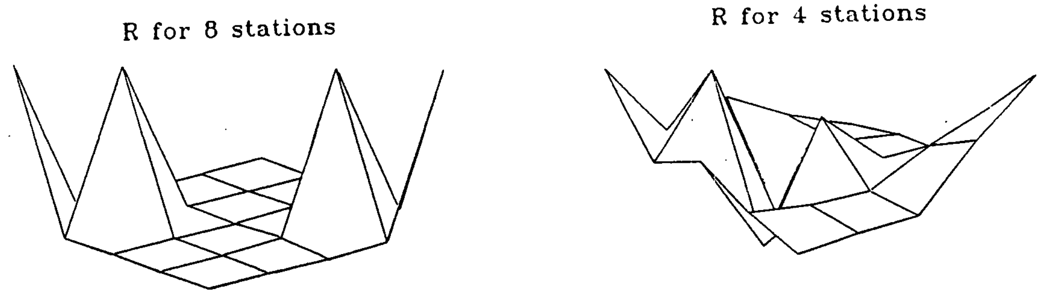

Figure 8. Perspective mesh plots of the resolution matrices for two different station distributions. The dimension of the matrices is 5 × 5, and the main diagonal of the matrices runs from left to right. The matrix on the left is for the case when all eight phases plotted in Figure 2 are used. Only the smallest eigenvalue corresponding to the pure strike-slip component (component #3) is eliminated from the inverse matrix. The other diagonal values are essentially 1. The matrix on the left is for the case when there are no oceanic stations, two eigenvalues are eliminated to construct an inverse with stability comparable to the first case. The off-diagonal values are now large.

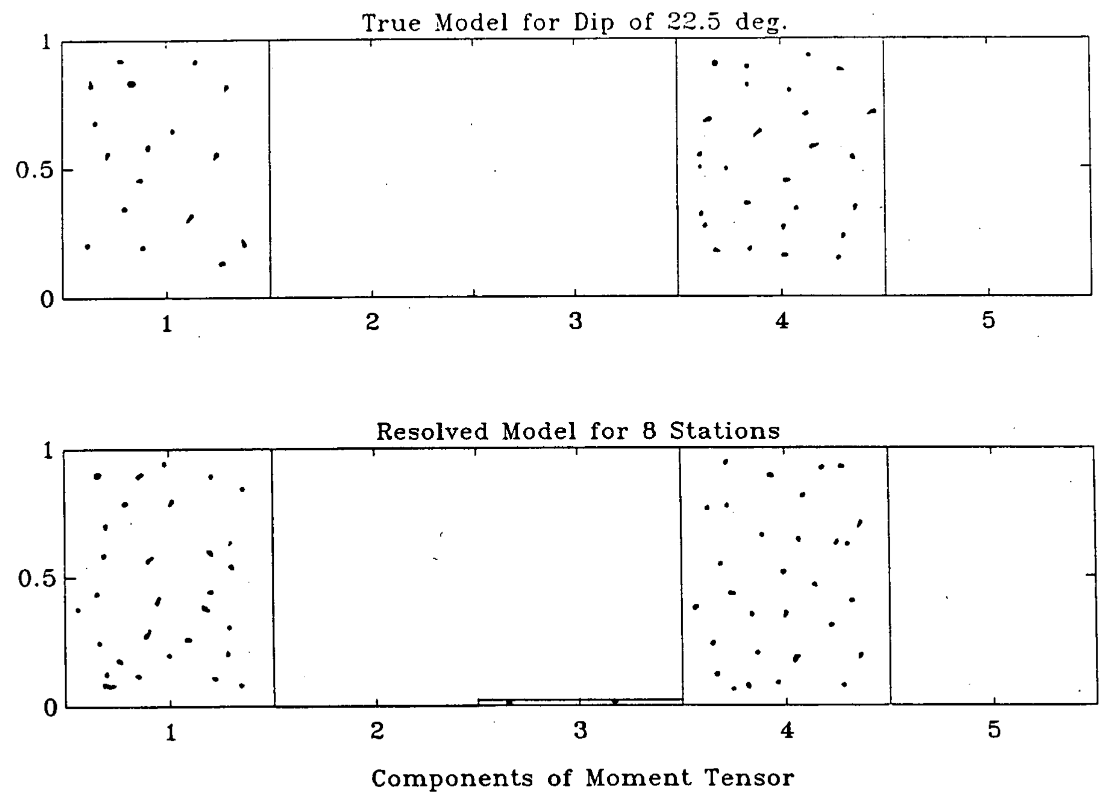

Figure 9. Numerical values of the five moment tensor components. The specified "true" model at the top represents a pure dip-slip earthquake with a fault plane striking NS with a dip of 22.5°. If all eight stations are available, then the "resolved" model that can be obtained by waveform inversion recovers the "true" model.

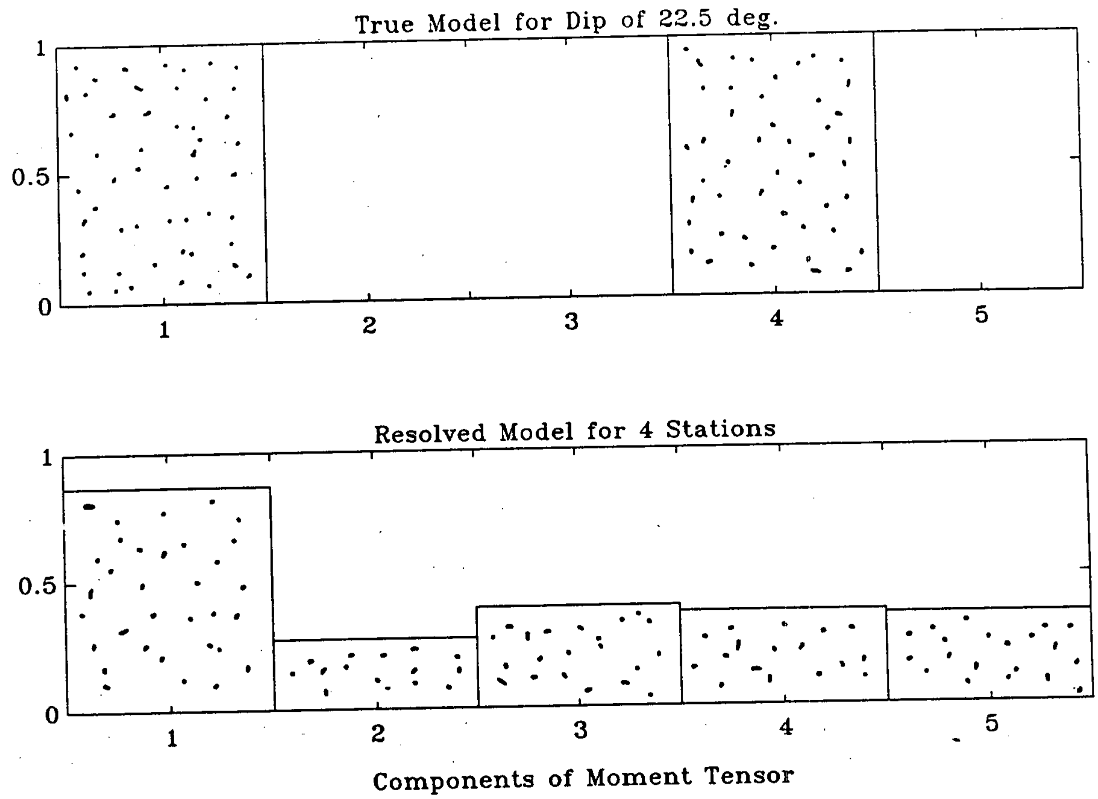

Figure 10. Same as Figure 9, except that the "resolved" model is calculated for the case when only the landward four stations are available. The ratio of components #1 to #4 is now distorted, the calculated dip for the NS striking pure dip-slip fault is now 33.7°. Stations that are oceanward from subduction zones are crtitical for fault dip resolution.

OCEAN-FLOOR SEISMIC STATIONS AND SOURCE MECHANISM STUDIES

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In general, the importance of ocean-floor stations is less critical for source mechanism studies than for earth structure studies, because, for source studies, what we really need, in most cases, is a good azimuthal coverage, but not necessarily a good path coverage. For most global problems, land stations will provide an adequate azimuthal coverage. As many studies have demonstrated, as far as the source can be approximated by a centroid (point source), land stations alone provide an adequate data. However, there are cases where ocean-floor stations would play a key role. One obvious case is a study of local or regional phenomena in oceanic areas which cannot be accomplished by land stations at all. Another case concerns the events which cannot be explained by conventional models, and require a detailed study using near-field data. Recent studies have shown that some events exhibit very unusual seismic radiation suggesting that they are not ordinary earthquakes. In these problems, the spatial extent and the geometry of the source are very important. Far-field data from land stations do not have sufficient resolution to determine them.

The following are a few examples where ocean-floor seismic stations would play a key role.

1. Improvement of source location and mechanism determinations.

Installation of ocean-floor stations improves the location (particularly the depth) accuracy and the resolution of source mechanism determination. Outstanding problems include the depth extent of oceanic intra-plate earthquakes such as outer-rise events. A particularly difficult problem is the determination of the vertical extent of very large events such as the Sanriku earthquake of 1933 and the Sumbawa earthquake of 1977. The problem has been debated for some time, but has not been resolved yet satisfactorily. Whether these events are relatively shallow, like many other outer-rise tensional events, or they involve failure of the bulk of the oceanic lithosphere has an important implication for the state of stress in the lithosphere. Likewise, the depth extent of large compressional outer-rise events, such as the 1981 Chilean earthquake, is another important problem.

2. Study of Unusual Events.

Although most earthquakes are caused by faulting, recent studies suggest that some events which were generally thought to be an ordinary tectonic earthquake may be caused by some other processes like large scale slumping and magmatic injection. Examples include, the 1929 Grand Banks, Canada, earthquake (slump), the 1946 Unimak Is. Aleutian Is. earthquake (slump), the 1986 Sanriku, Japan, earthquake (slump), and the 1984 Torishima earthquake (magma injection).

Whether the earthquake is an ordinary earthquake or a large scale submarine landslide has an important implication for estimation of long-term seismic hazard, because the repeat time of tectonic earthquakes (faulting) and submarine landslides are controlled by entirely different processes, plate motion and sediment transport respectively. In case of the Grand Banks earthquake, proper understanding of its cause is particularly important in view of commencement for hydrocarbon exploration along the continental margin of Atlantic Canada.

The Torishima event exhibited very unusual radiation patterns of seismic body and surface waves, and it is almost impossible to explain the data in terms of an ordinary double couple mechanism. The most likely cause of this is magmatic injection into the oceanic sediment. It is possible that many events like this may have escaped notice in the past because of the lack of adequate data.

The difficulty in the determination of the mechanism of these unusual events is that far-field data are not adequate to resolve the details of these events. Addition of near-field data, even a few, including ocean-floor instrumentation would greatly improve the resolution.

Tsunami Warning

Detecting seismic waves prior to the arrival of tsunamis is the basis of tsunami warning. Since the tsunami travel time to a coastal area from a nearby tsunami source is about 10 to 30 minutes, a very rapid warning system is required. To accomplish rapid warning, a use of near-field stations including those at ocean-floor is essential. A numerical experiment for the source-site geometry of Alaska and the Aleutians suggests that long-period seismic data from nearby stations can be effectively used for tsunami warning. It is true, however, that installation of pressure gauges is more suitable for direct determination of tsunamis. Nevertheless, it is desirable to understand the nature of tsunamigenic earthquakes using the broadband seismograph network which, when methodology has been developed, will enable us to make predictions of regional variations of tsunami amplitudes and arrival times.

Technical Aspects

Since we have not had any broad-band record from ocean-floor stations, there is no well established method regarding how to utilize near-field data for resolving the details of seismic source. However, broad-band near-field data recently collected by a few land stations do indicate that these data can be very effectively used for source studies.

Section B2: Existing and Planned Global and Regional Networks

1. The Federation of Digital Seismographic Networks: M.J. Berry
2. Siting Plans of the Federation of Digital Broadband Seismographic Networks: E.R. Engdahl
3. The Geoscope Program: Barbara Romanowicz
4. POSEIDON: Japan's Proposed Research Program in Global Seismology: POSEIDON Working Group, Y. Hamano, presenter
5. United States National Seismograph Network: Robert P. Masse, John R. Filson, and Andrew Murphy
6. The Role of Ocean Bottom Observatories in the Global Seismic Network: Rhett Butler
7. MEDNET - The Italian Broad-Band Seismic Network for the Mediterranean: E. Boschi, D. Giardini, A. Morelli, G. Romeo and Q. Taccetti

THE FEDERATION OF DIGITAL SEISMOGRAPHIC NETWORKS

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During the 1970's a number of institutions developed digital seismographs and deployed these in sparse global networks to address specific scientific or technical problems. The very long period (100 sec to 1 hour) International Deployment of Accelerographs (IDA) network was an important example providing much useful information on gross earth structure and the nature of earthquake sources, (Agnew et al., 1976). Other examples include the SR0 network and the more recent RSTN, deployed to further research in discrimination between earthquakes and explosions, (Peterson and Orsini, 1976). Although developed for a specific purpose these networks have also proved invaluable in providing information on earth structure and earthquake sources. Using the SRO network as a major component, the United States Geological Survey has now operated its global Digital Seismograph Network (GDSN) for over a decade, and has made the data from the GDSN generally available to the research community.

In the 1980's, with the development of stable well calibrated very broad-band seismometers with high linear dynamic range by Wielandt and Streckeisen (1982), and parallel developments leading to the availability of mass storage devices and powerful economical computers, several institutions began to plan for and to deploy broad-band digital networks. Most notable was the early development of the GEOSCOPE network by the Institut de Physique du Globe with 8 stations in place by the end of 1984, and with plans for more than doubling the network later, (Romanowicz et al., 1984). Also at this time, a consortium of U.S. institutions, IRIS, developed a comprehensive plan for a global network of 100 stations and an associated data centre to ensure that the data from the network would be readily available to its members, (IRIS, 1984). Several other countries, recognizing the significance of the new technology, began to plan for new broad-band seismographic networks on their territories, and some banded together to develop ambitious plans for regional networks, e.g., the European ORFEUS network.

During 1985 it became clear to many seismologists that there was an important opportunity for the seismological community to coordinate its efforts in order to develop an optimum global seismograph network. Under the sponsorship of the Interunion Commission on the Lithosphere, representatives of 20 institutions met in Karlsruhe in the spring of 1986. During two days of meetings the participants agreed to found a Federation of Digital Broad-Band Seismographic Networks, with the following statement of purpose.

Formation of a "Federation of Digital Broad-Band Seismographic Network"

> The international seismological community recognizes new opportunities within its field for improved understanding of the internal structure and dynamical properties of the Earth provided by recent developments in seismographic network technology.
>
> The developments include greatly improved broad-band seismographic systems that capture the entire seismic wave field with high fidelity, efficient and economical data communications and storage, and widely available, powerful computing facilities It also recognizes that rapid access to seismic data from arrays of modern broad-band digital instruments, wherever they might be, is now possible.

In view of the above, and to take advantage of existing and developing global and regional networks it is considered that a Federation should be formed to provide a forum for:

- developing common minimum standards in seismographs (e.g. bandwidth) and recording characteristics (e.g. resolution and dynamic range);
- developing standards for quality control and procedures for achieving and exchange of data among component networks;
- coordinating the siting of additional stations in locations that will provide optimum global coverage.

The Federation will welcome the participation of all institutions committed to the development of broad-band seismographs and willing to contribute to establishment of an optimum Global system with timely data exchange.

In the following August, the Federation held its founding meeting in Kiel, when it elected officers and set up four working groups to consider:

- digital broad-band seismographic specifications
- siting plans
- data collection and exchange formats
- data .

The Federation met again, briefly in San Francisco in December 1986, and during the Assembly of the IUGG in August 1987. In December 1987 the working group on data formats met in Albuquerque and the Federation will hold its 1988 annual meeting in June in Blanes, Spain.

Digital Broad-Band Seismograph Specification

The Federation has adopted as its primary objective the development of a "global" network of very broad-band high dynamic range digital seismographic stations having similar system response. The 'global' response is shown in Figure 1 and represents what is currently possible with state-of-the-art technology. Seismographs built to these specifications can faithfully record the signals from small M_L3 earthquakes at 1° and great M_W9 events at 30° without distortion.

Twelve very broad-band (VBB, 5 Hz - 250s minimum) 'global' and 96 broad-band (5 Hz - 20s) seismographic stations are now operating around the globe. In addition, members of the Federation have reported plans and proposals for a further 91 VBB stations by either converting existing stations or by installing new ones. Thus the community can look forward to a substantial global network of stations meeting the 'global' standard within approximately five years.

The Federation recognizes that for many 'regional' studies, stations with more limited capability will be entirely adequate, and the response for such stations is also shown in Figure 1. While regional stations will not faithfully record the longer period (greater than 250 sec) signals from large earthquakes, they will nonetheless be entirely adequate for recording body waves and intermediate period surface waves. Besides providing excellent high quality data for regional studies, they will provide a higher density distribution of stations than would otherwise be possible with the higher cost global stations.

Siting Plans

The distribution of existing planned and proposed Federation stations of both global and regional type is shown in Figure 2. Undoubtedly not all the planned and proposed stations will be deployed; however, it is now clear that the seismological community can confidently anticipate a network of broad-band stations distributed about the earth by the early 1990's. The distribution of course is not and will not be ideal. In particular the majority of stations are necessarily on the continents leaving the oceans, a significant proportion of the earth's surface, unmonitored. Stations sited on oceanic islands will provide much useful data especially in the West Pacific, but large areas will remain unmonitored unless ocean-bottom systems are deployed. Of the continental areas, the territory of the USSR is notable for its lack of stations contributing to the Federation network.

Data Formats

In 1985, the International Association of Seismology and Physics of the Earth's Interior, Commission on Practice formed a working group on Digital Data Exchange to develop a 'standard' to enable the efficient exchange of digital data from the digital seismograph networks then rapidly developing, (Halberg et al., 1988). Shortly after this, as explained above, the Federation was formed and, by agreement with the Commission on Practice, accepted responsibility for developing the standard. A preliminary meeting in August 1987 of the Federation Working Group on Data Collection and Exchange Formats considered a draft that had been developed by the U.S. Geological Survey and IRIS for distributing data from their own and cooperating digital stations. In the light of these discussions and others, Halbert et al. of the U.S.G.S. developed a complete draft for discussion in December where representatives of eight Federation Member Networks critically reviewed the standard in detail. The Standard for the Exchange of Seismic Data (SEED) will be considered for adoption by the Federation in Blanes, Spain, on June 19, 1988. It is hoped that it will be accepted widely and become de facto the international standard for the exchange of digital earthquake data. Some organizations plan to use the formats of the standard right at their seismograph stations as well as for archive and exchange purposes.

Data Centres

It is highly probable that by the mid-1990's there will be well over 100 broad-band digital stations operating around the globe - each with a system response meeting at least the minimum specification of the FDSN. The task of collecting, archiving and distributing all of these data to the international research community will be a major task, that may be beyond the resources of any one institution. The FDSN has therefore decided that the only practical approach is to coordinate the activities of existing and developing data into a network that will allow a sharing of costs and workload for the benefit of the research community. The conceptual model is shown in Figure 3.

National Centres and Institutions are responsible for the installation and operation of their seismograph stations, and for the collection and archiving of the complete waveform data from their own network. By agreement with the FDSN they will contribute full and event waveform data from a selection or from all their stations to a specified Global Network Data Centre. In addition to contributing their data to the FDSN, they will also be responsible for distributing FDSN data to their particular research community. Examples of National Centres include those operated by Federation Members in Australia, Canada, China, FRG, Italy and the UK.

Global or Major Regional Centres may install and operate global or regional networks of seismograph stations. They will act as regional data for member national networks. They will be responsible for the collection and archiving of their own data and that contributed to them by member (national) networks. They will collate their data and make full wave form available to the Federation Archive, and event wave-form data available to the Federation Event-Data Centre. In addition to contributing their data to the FDSN, Global Data Centres will be responsible for distributing FDSN data to National Centres and depending upon circumstances to individual seismologists.

Federation Data Centres

The U.S. Geological Survey (World Data Centre A - National Earthquake Information Centre) has agreed to act as the Federation Event-Data Centre, and is now accepting digital data from the global (GEOSCOPE and IRIS), major Regional Data Centres (ORFEUS), and some national networks. Event data will be distributed in the SEED format on CD-ROM's through the Federation data network.

In the future the Federation will consider the need for a Federation Archive that will collect and archive complete waveform data from all Federation stations. The task is of such a magnitude however that careful consideration will have to be given to the nature of the service provided by such an Archive. Clearly unlimited access by the international research community will likely not be possible. However, the rapidly developing technologies of mass digital storage and data communication will probably be of significant assistance, making a reasonable compromise solution practicable.

Summary

Since its formation in the spring of 1986, the Federation of Digital Seismographic Networks has attracted 11 Members representing approximately 20 institutions. It has adopted standards for the system response of Federation seismographic stations and for the data formats to be used in the exchange of earthquake data. The siting plans of member networks have been adjusted by detailed discussion between members, resulting in a significantly improved collective global network. An event-data centre has been established by the U.S. Geological Survey on behalf of the Federation. Thus the Federation has met its initial goal of coordinating the efforts of its Members so as collectively to deploy and to operate an optimum global network of digital broad-band seismograph stations. The stage is now set for Members to deploy a significant number of seismographic stations meeting Federation standards in the near future. If current plans and proposals are executed, the Federation's Global Network should be comprised of at least 90 stations by the early 1990's.

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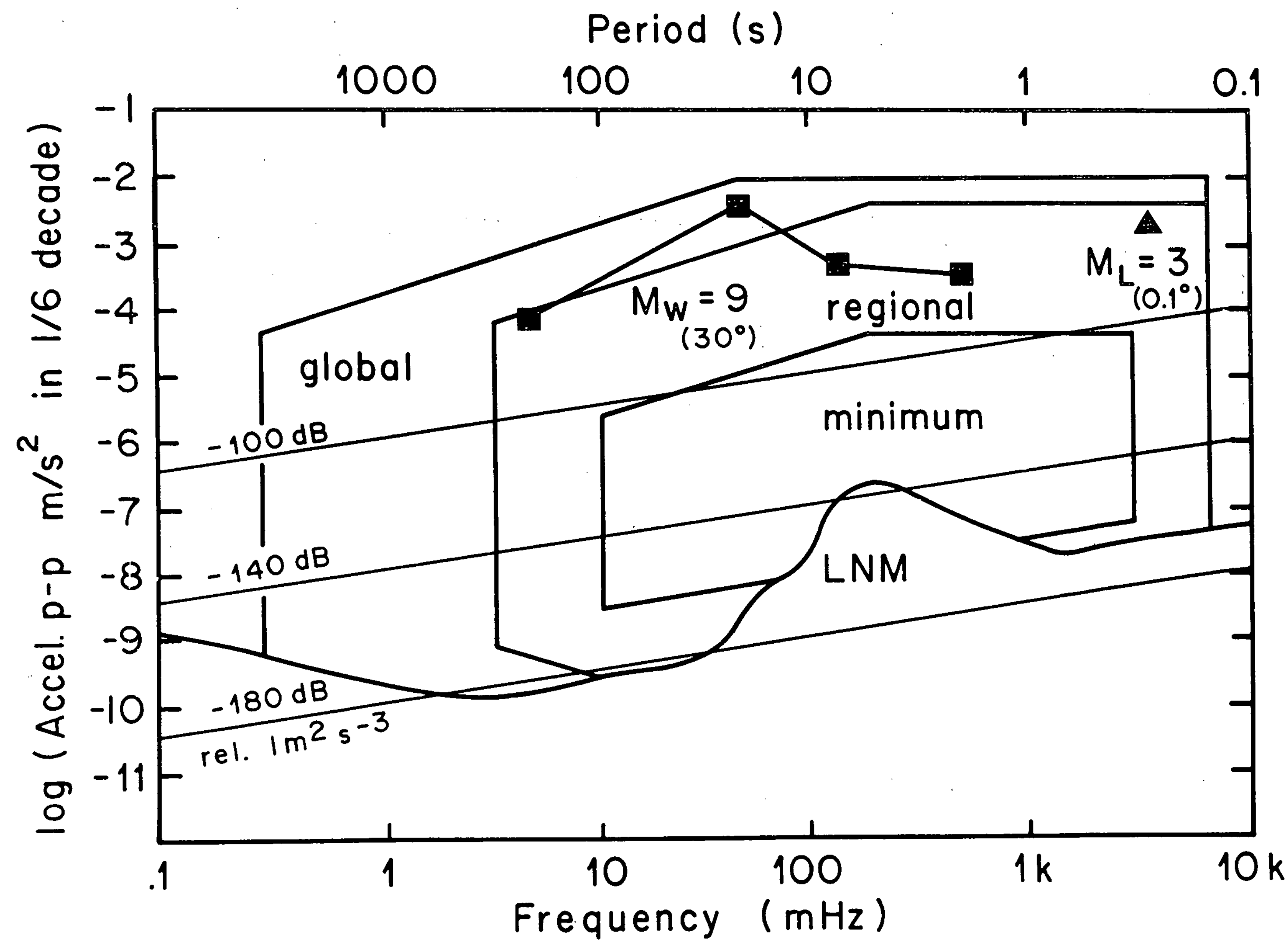

FIGURE 1: Federation Seismographic Response

DIGITAL BROADBAND SEISMOGRAPH STATIONS
APRIL 6, 1988

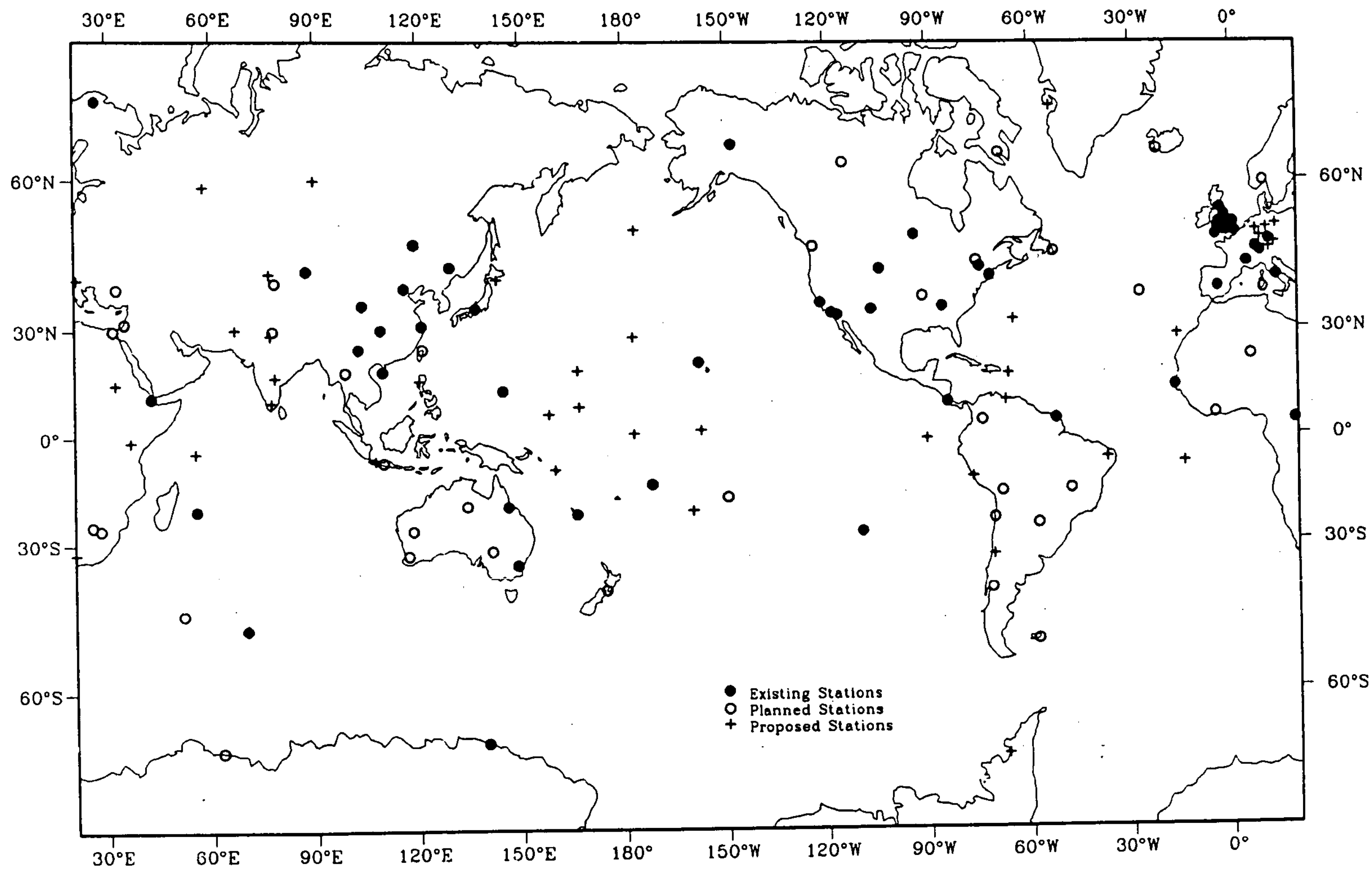

FIGURE 2

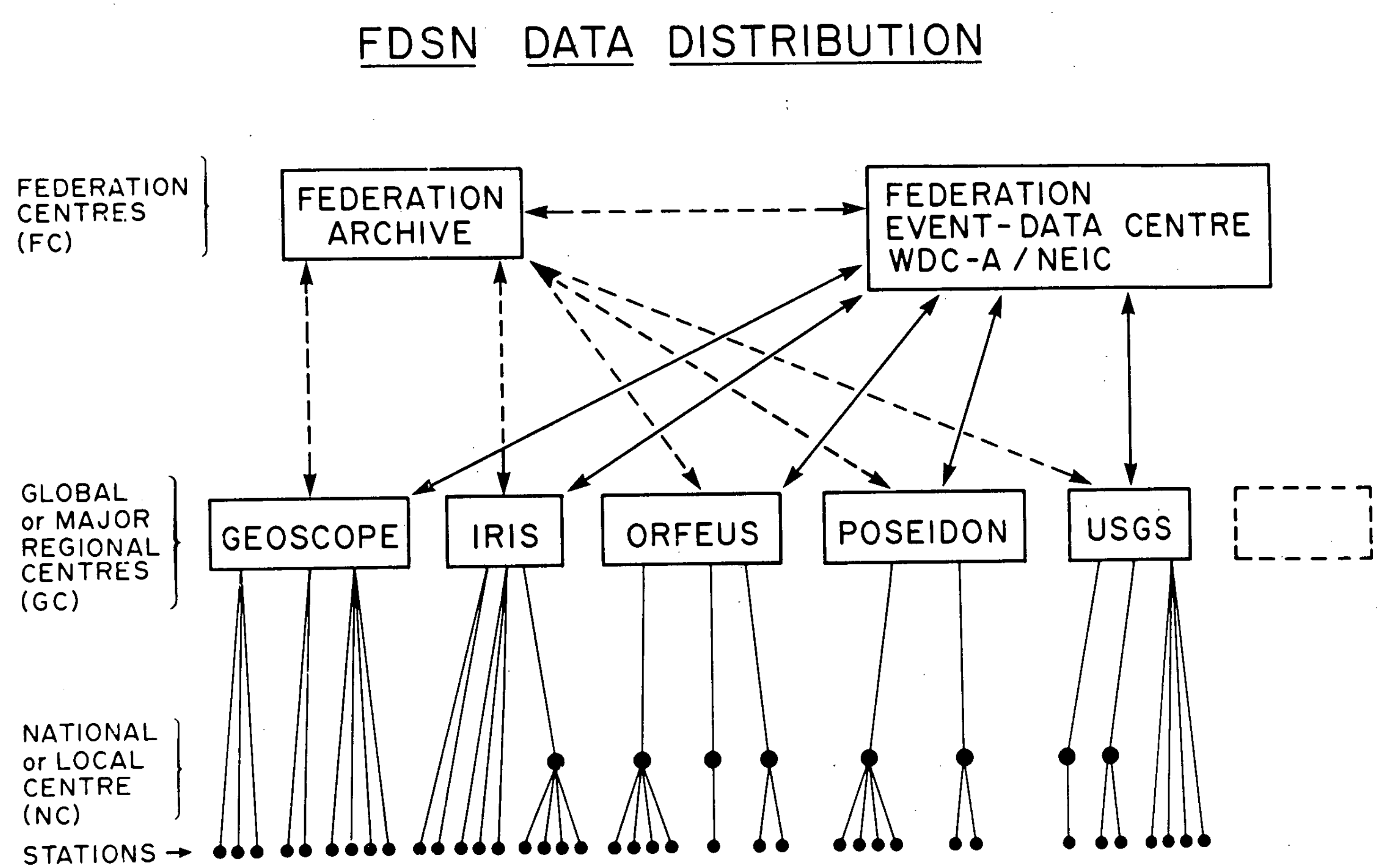
FDSN DATA DISTRIBUTION
FEDERATION CENTRES (FC)
FEDERATION ARCHIVE
FEDERATION EVENT-DATA CENTRE WDC-A / NEIC
GLOBAL or MAJOR REGIONAL CENTRES (GC)
GEOSCOPE
IRIS
ORFEUS
POSEIDON
USGS
NATIONAL or LOCAL CENTRE (NC)
STATIONS →

FIGURE 3

SITING PLANS OF THE FEDERATION OF DIGITAL BROADBAND SEISMOGRAPHIC NETWORKS

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The Federation of Digital Broadband Seismographic Networks was formed to facilitate the development of a global network of digital seismographic stations of high dynamic range and broadband frequency response. To coordinate the siting of stations in locations that will provide optimum global coverage, the Federation established a Working Group on Siting Plans. The Working Group has compiled an inventory of existing, planned, and proposed stations that meet certain minimum criteria for the seismic systems. The current global station distribution is presented in a series of maps that emphasize the coverage provided over large oceanic areas.

The station inventory was compiled only for stations that meet the following minimum criteria: a single data stream per component; a range of signals accessible with standard (WWSSN) short-period and long-period seismographs; a passband of 2 or 3 Hz to 100 s; and the recording of teleseismic signals up to M_S 6.5 without clipping. Stations were further classified as existing, planned (installation by 1990 with a written financial plan), or proposed (installation after 1990). The inventory also includes information on the station name, geographic location, station code, latitude, longitude, the supporting program, and the station characteristics. Station characteristics include a classification of the available responses, operating dynamic range, and other information such as satellite transmission and dial-up capability. Presently there are 58 existing, 41 planned, and 44 proposed digital broadband stations that are not colocated.

The maps reveal present characteristics of the station distribution and have as a specific purpose the identification of overlaps and obvious holes in coverage. The station density for some regions is much greater than for other regions (e.g. the Far East, Europe, and North America). Although the goal is a more uniform global network, some higher density regions can have beneficial effects for regional studies and, in addition, would allow particular kinds of global studies. Only a few stations are operating in South America, Africa, and Australia, but many new stations are planned. Other areas of the globe, including large oceanic regions, are not well covered by either existing, planned, or proposed digital broadband stations.

A network of broadband downhole ocean floor monitoring stations can be used both to augment the global network and to provide more precise information about the regional structure of the oceanic crust and upper mantle. Together, the oceans represent a unique tectonic province that has never been directly sampled with seismic methods, except for the most shallow structures. At present, recordings of surface and body waves generated by earthquakes in the oceans must be made on continental platforms or on oceanic islands that are seismically noisy and different geophysically from typical oceanic regions.

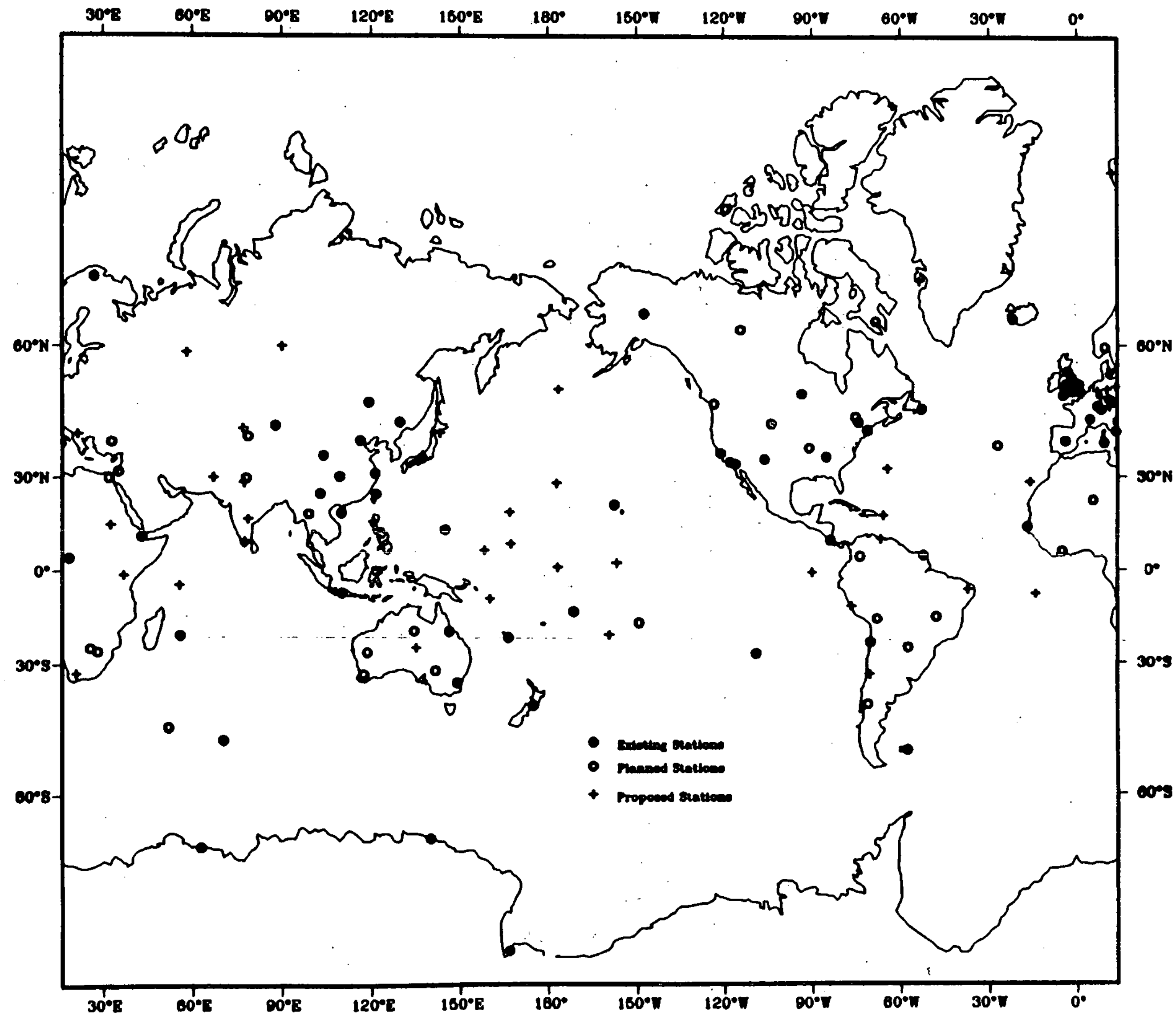
Major differences between continental and oceanic structure have been known for some time, and recent studies show that some of the more extended differences persist to great depths. A detailed understanding of upper mantle structure beneath the oceans is required if the composition and dynamics of the entire mantle are to be fully understood.

Although significant progress with primarily land-based coverage has been made, further progress is severely limited by the lack of permanent monitoring stations in the deep ocean.

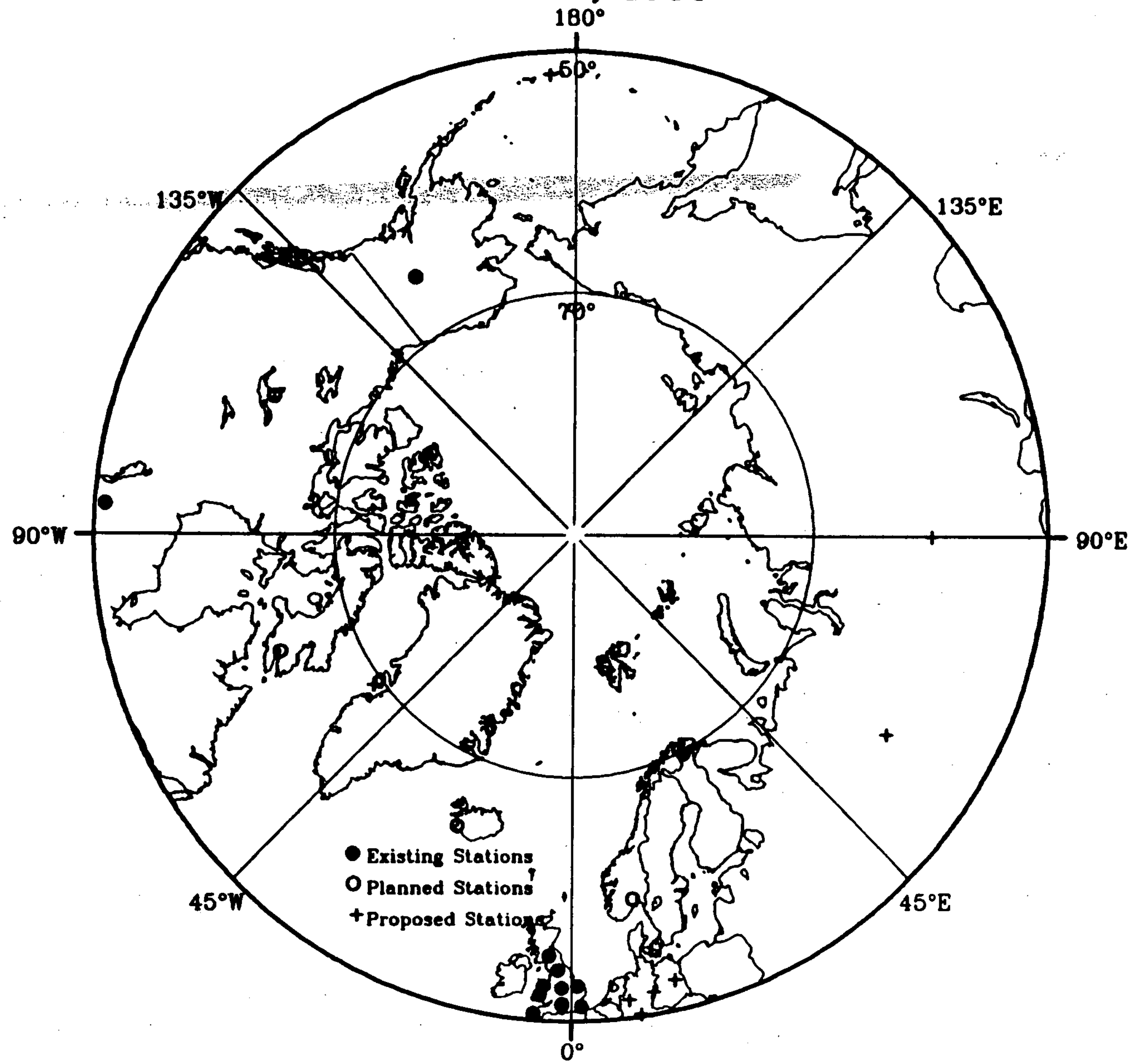
Maps of digital broadband station distribution over major oceanic areas show that the station deployment on oceanic islands and along oceanic margins is nearly physically maximized, yet we are far from a uniform global coverage. Where islands are abundant, a good density of stations can be achieved, and where they are not, there are obviously large gaps in coverage. Clearly, a uniform global coverage cannot be achieved without the installation of permanent ocean-bottom systems with broadband characteristics.

To date, the Federation has been successful in encouraging the international seismological community to deploy very broadband stations. If present plans come to fruition, there will be relatively dense global coverage of continents and of parts of some oceanic regions within the next five years. Any efforts to deploy broadband downhole seismometers in the deep ocean are certainly needed and would obviously complement a similar effort by the Federation to deploy land-based instrumentation.

DIGITAL BROADBAND SEISMOGRAPH STATIONS
APRIL 6, 1988



DIGITAL BROADBAND SEISMOGRAPH STATIONS
ARCTIC AREA
APRIL 6, 1988



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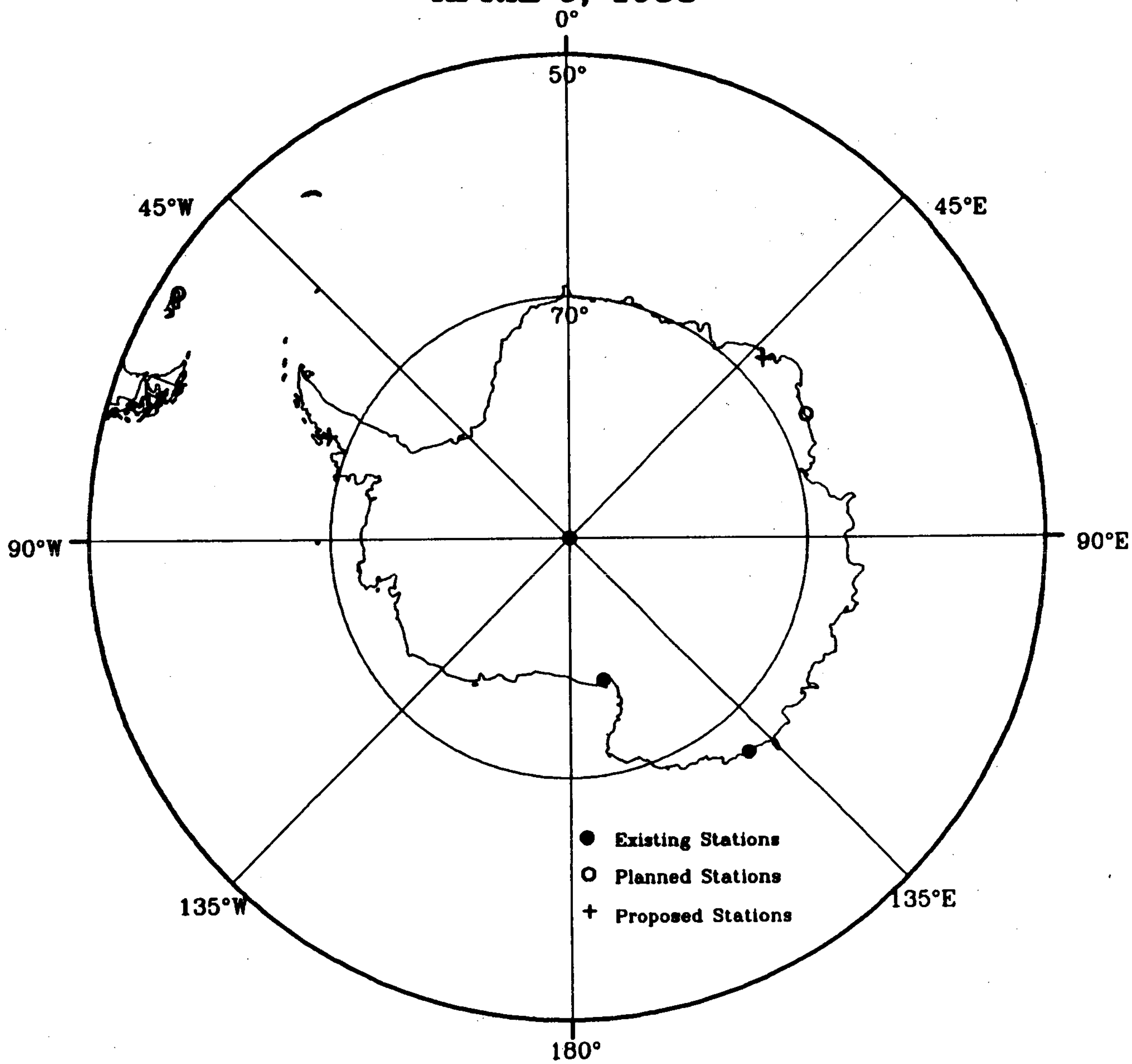
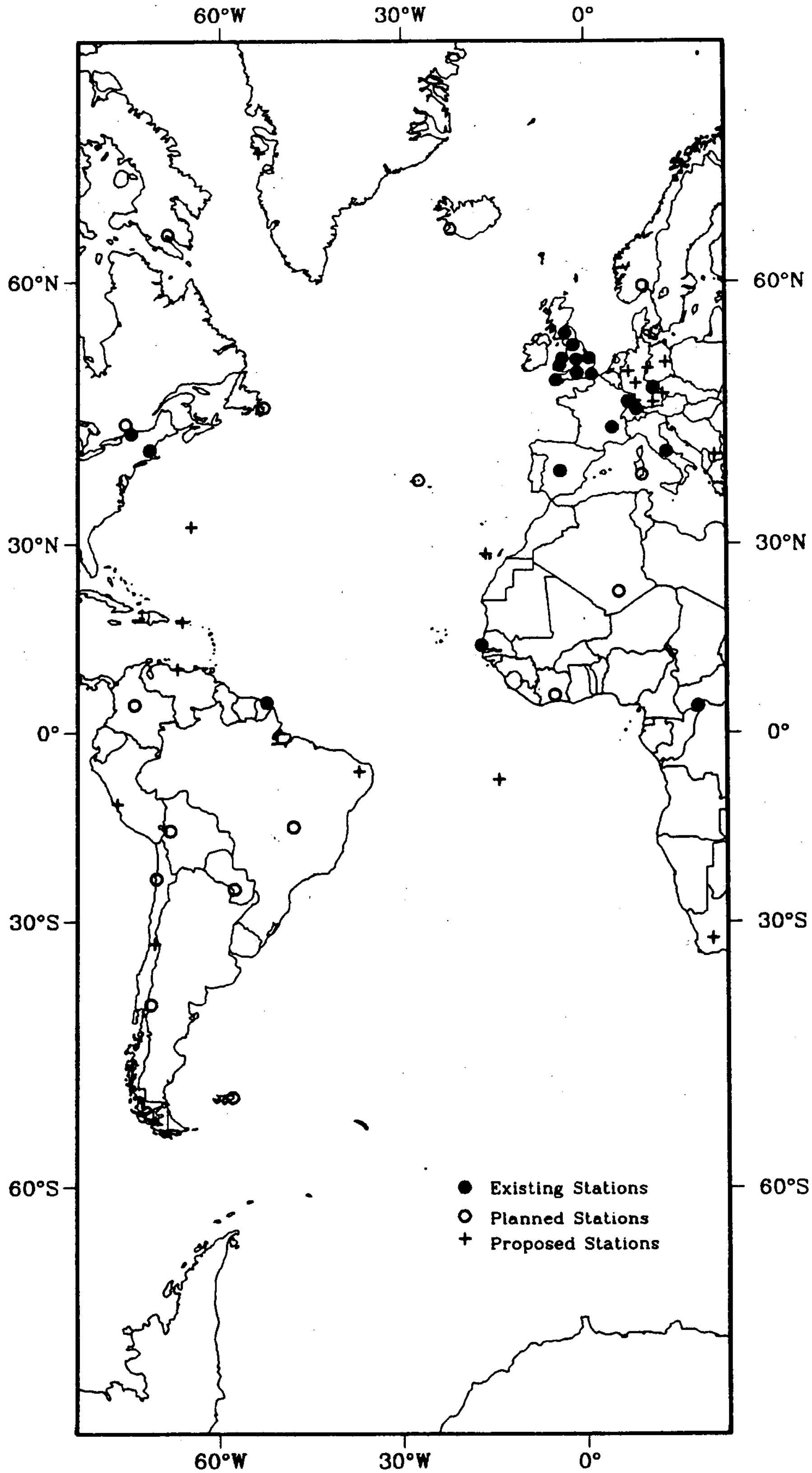
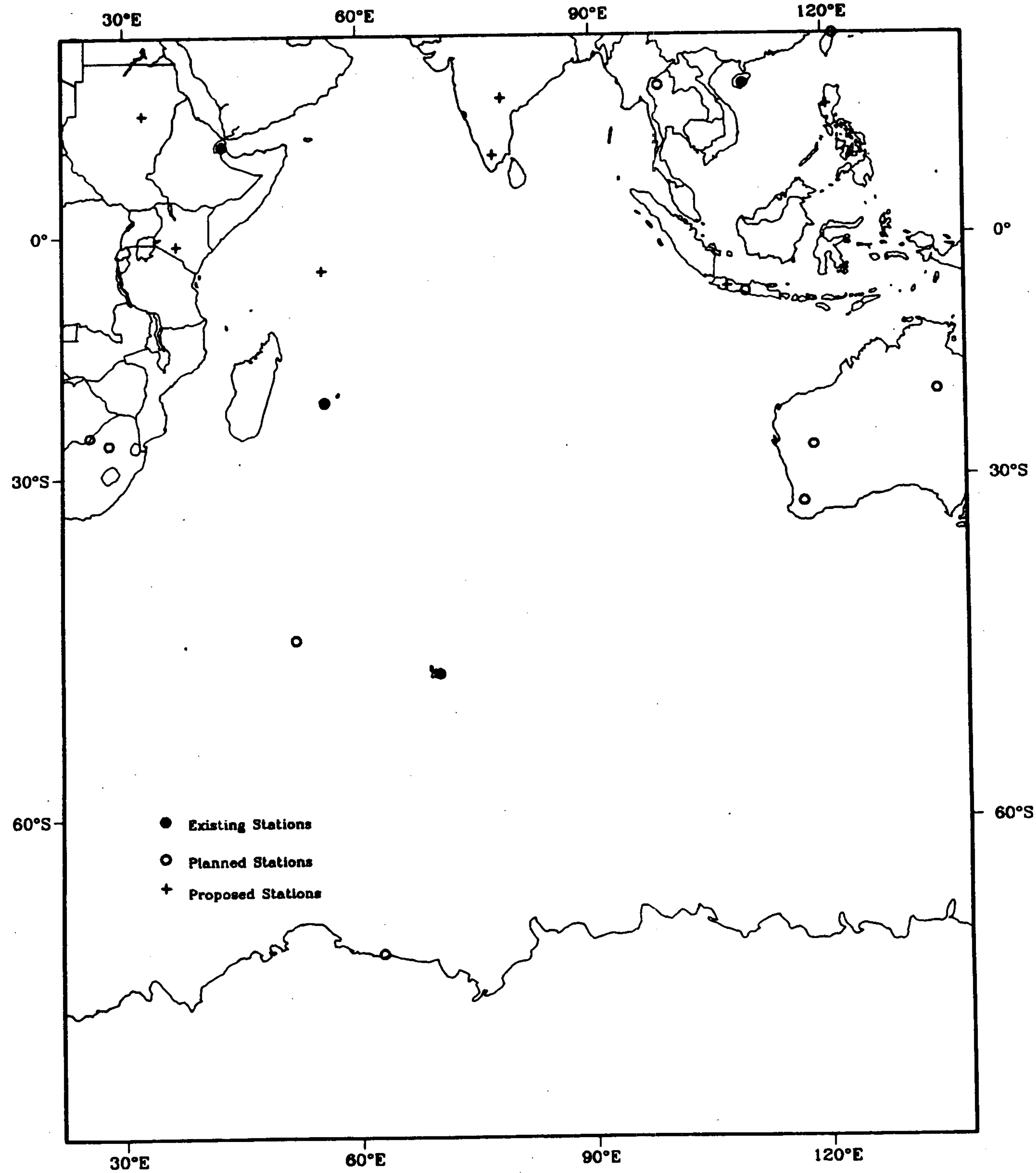
ANTARCTIC AREA
APRIL 6, 1988



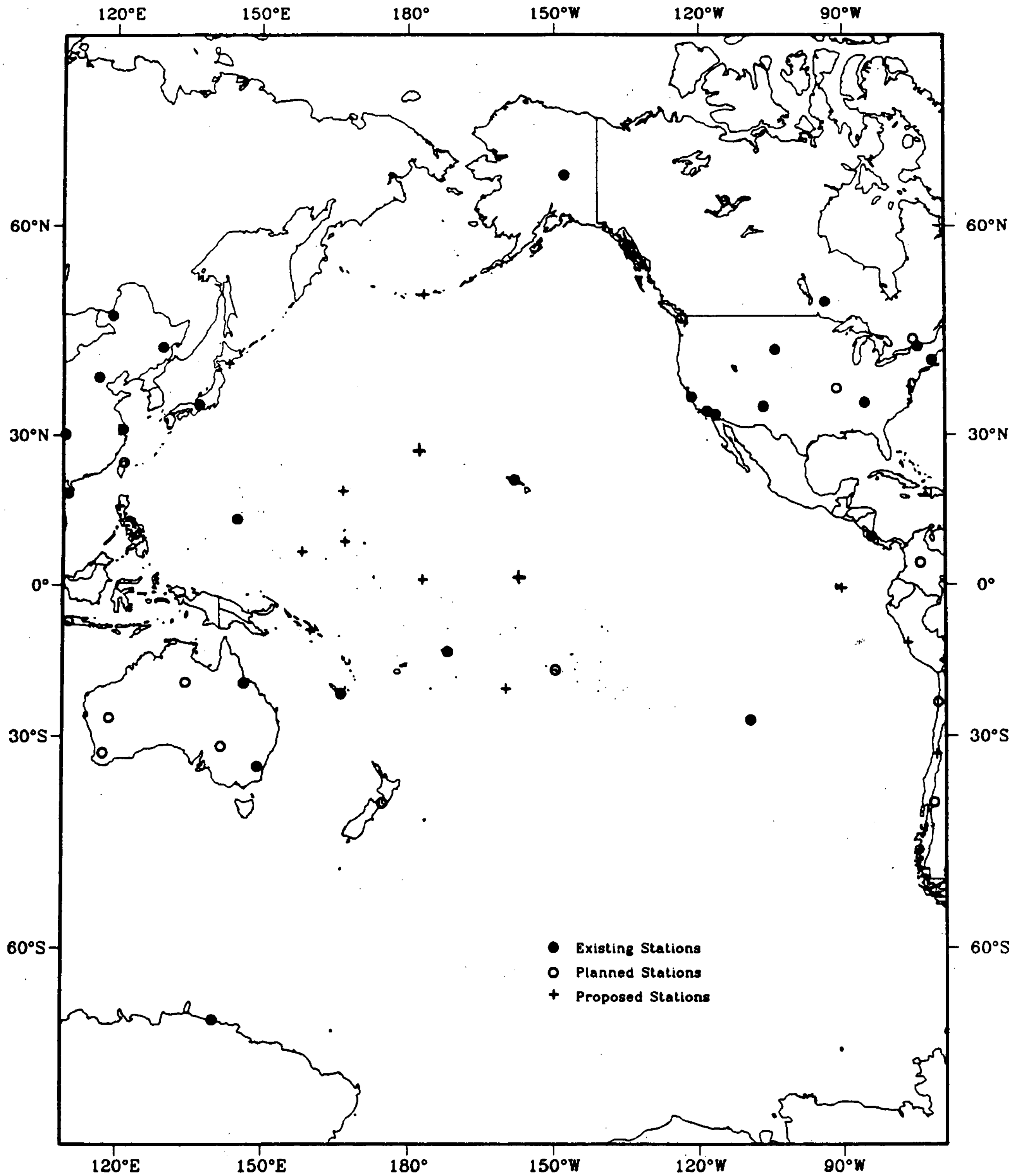
DIGITAL BROADBAND SEISMOGRAPH STATIONS
ATLANTIC OCEAN AREA
APRIL 6, 1988



DIGITAL BROADBAND SEISMOGRAPH STATIONS
INDIAN OCEAN AREA
APRIL 6, 1988



DIGITAL BROADBAND SEISMOGRAPH STATIONS
PACIFIC OCEAN AREA
APRIL 6, 1988



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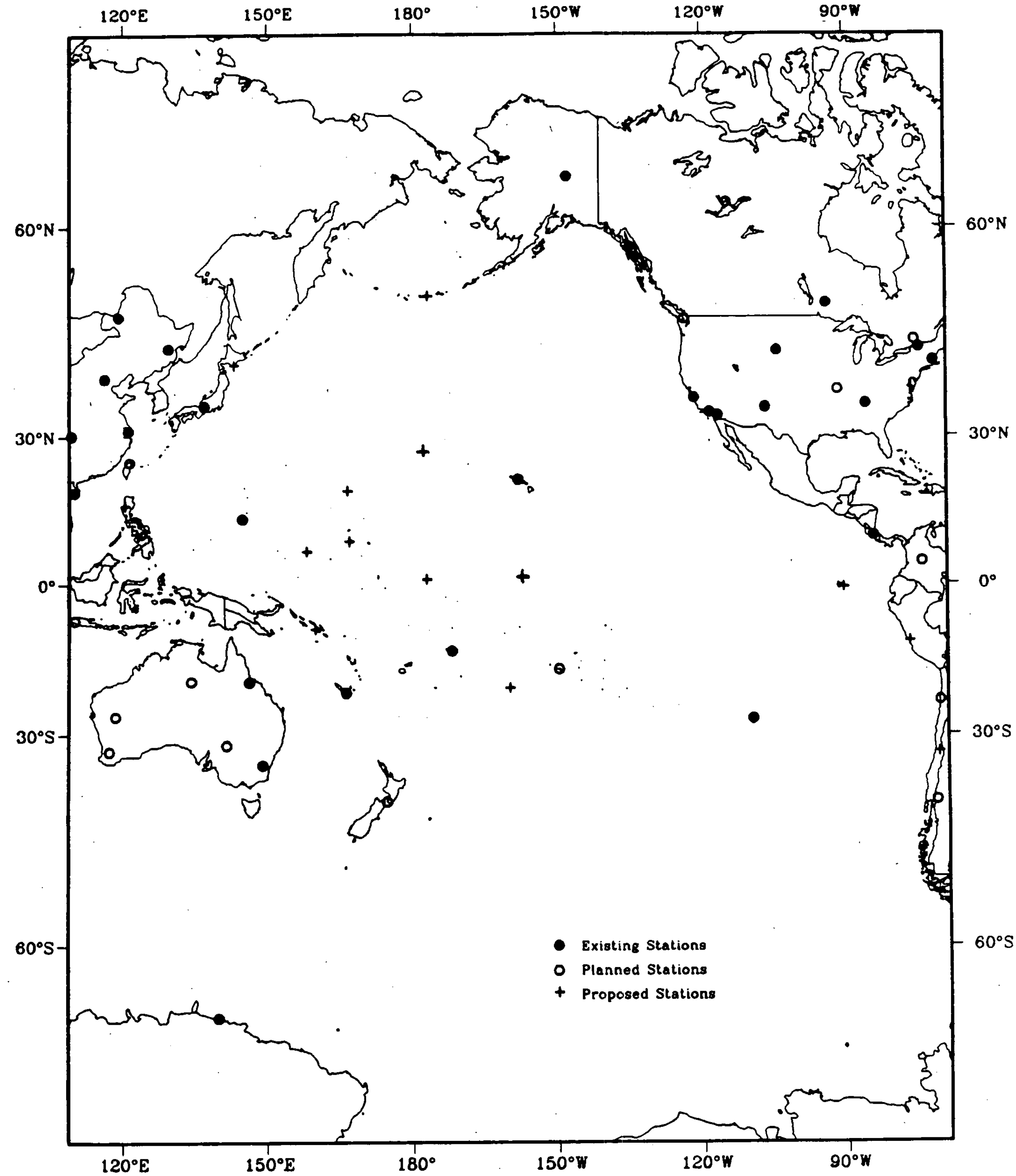
PACIFIC OCEAN AREA
APRIL 6, 1988



THE GEOSCOPE PROGRAM

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The GEOSCOPE global seismological network is a national program of France, sponsored since 1982 by I.N.S.U. of C.N.R.S., with contributions from several other French or foreign institutions*. It presently counts 19 stations distributed worldwide, with three more to be installed in 1988. The goal is 25 stations by the end of 1989, which will constitute the French contribution to the new global seismological network.

All stations are equipped with STS three component seismometers (Wielandt and Streckeisen, 1982) and a recording system designed by the Division Technique of I.N.S.U. All stations record three component very long period (VLP) data continuously at a sampling rate of 10 sec, and broad band (BRB) data on event detection at a sampling rate of 5 Hz. Stations in Japan, Australia, France and Hawaii, equipped with the newest version of the STS seismometers (Wielandt and Steim, 1986) also record the very broad band (VBB) channel at a sampling rate of 20 Hz, on event detection. All stations will be gradually upgraded to the VBB version of the seismometers and a new high density recording device is presently being tested to permit continuous VBB recording and possible future extensions to other channels.

A Data Center, located at the Laboratoire de Sismologie of the IPG in Paris serves to collect, verify, archive and distribute data to users worldwide. Through an agreement with the U.S.G.S., GEOSCOPE data are also available on CD ROMS, and will be part of the F.D.S.N. data collection. A particular feature of the present recording system is the possibility to teletransmit data back to Paris through dial up, via the ordinary international switched network and, in exceptional cases, via special packet transmission networks (e.g. the French system TRANSPAC). 12 stations are presently accessible in this manner. In the case of a major earthquake, 24 hours of VLP data and 15 mn of BRB data are brought back in quasi real time for preliminary determination of source parameters and rupture history. The effectiveness of this system has recently been demonstrated on the occasion of the occurrence of the Alaska earthquake of November 30, 1987, for which we were able to present results in avant-premiere at the December San Francisco AGU Meeting.

The teletransmission system also serves to daily monitor the stations. This is done automatically at night to reduce costs and provides an efficient way to optimize station operation time and quality of recorded data. Stations which are not easily accessible by phone (e.g. the planned Wushi station in Western China) are monitored, whenever possible, via the ARGOS international satellite system.

While the broadband data collection, initiated in 1985, is presently building up and starting to contribute to many innovative studies of earth structure as well as earthquake studies, the very long period collection is already sufficient for global large scale studies of free oscillations and surface waves. A global 3D upper mantle model based entirely on GEOSCOPE data has for example recently been obtained in Paris (Monfret, 1988). Its originality lies in the fact that it relies entirely on first arriving Rl and R2 Rayleigh wavetrains (not saturated for large events on the STS instruments), providing better resolution for a relatively small number of paths.

As any global network, GEOSCOPE strives to achieve an optimal distribution of stations around the world, and the problem of instrumenting the vast ocean areas is presently at the center of our concerns. Already, GEOSCOPE has helped to fill important gaps in the distribution of long period (and broad band) stations in the Indian Ocean (stations RER, PAF, CRZF and DRV), leading in particular to unexpected findings on the seismicity in the southern region of this ocean (many magnitude 4.5 to 5.5 events unreported to ISC). It has contributed several stations in the Pacific (PPT, NOC, KIP as well as INU). The need to further instrument the oceans is clear. The question then arises as to whether it is preferable to build stations on any available rock emerging from the sea (a relatively economical and technically straightforward solution), or to put them on the seafloor. A compilation of noise data on GEOSCOPE stations confirms that island sites are generally noisier than continental, even surface sites, although large variations exist between different locations, especially as far as horizontal long period noise is concerned. On the other hand, there are virtually no existing data on long period noise on the ocean floor. While many technical problems remain to be solved before a permanent ocean floor observatory can be installed, we are planning an experiment in the Spring of 1989, in the Indian Ocean, in order to measure long period noise out to several hundreds of seconds at a depth of about 5000m, using an OBS case recently designed in France. Depending on the results of this experiment, if successful, further steps will be taken towards the design of BRB/VLP stations on the ocean floor or in ocean boreholes. These developments, hopefully conducted in close connection with similar projects in other countries, naturally lead, in France, to active cooperation with the community of oceanographers, and steps towards this are presently being taken.

* ORSTOM has financed 2 stations and contributed to one, TAAF, one station and contributed to 2 more, UCSC financed a large part of station SCZ. A 19th station, BFO in Western Germany, is contributing very long period data to the GEOSCOPE data collection.

POSEIDON: JAPAN'S PROPOSED RESEARCH PROGRAM IN GLOBAL SEISMOLOGY

POSEIDON Working Group
(Presenter: Y. Hamano, Earthquake Research Institute, Tokyo University)

As Japan's contribution to international research in global seismology,we propose to deploy seismographs in the Western Pacific, and in Eastern and Southeastern Asia. Our proposed project has been named the POSEIDON (Pacific Orient SEIsmic Digital Observation Network) Project. The POSEIDON project will be coordinated with the FDSN (Federation of Digital Broadband Seismographic Networks) and its member countries, to ensure maximum productivity and eliminate duplication of effort. The official Japanese liaison to FDSN is through the Global Seismology Subcommittee of the National Committee for Seismology and Physics of the Earth's Interior of the Science Council of Japan.

The first ten year period of the POSEIDON Project will be divided into two stages, each of which will last about five years. In the first stage, seismic stations will be deployed on continents and islands, a data center will be established, and development work for a deep ocean seismic network will begin. In the second stage, which we hope will begin within five years of the onset of the project, we will deploy the deep ocean seismic network and implement efficient methods for transmitting large volumes of data to and from the data center. We expect that the POSEIDON network and data center will remain operational for a period of at least 20 years.

Highlights of the POSEIDON Plan

(1) Data from the POSEIDON network will greatly enhance the resolution with which we can study the tectonic processes occurring at the convergent margins in the western Pacific, and will also greatly enhance our ability to resolve the three dimensional structure of the earth's interior, on both a regional and global scale.

(2) We propose to conduct an extensive long-term program of broadband seismological observations in the deep ocean basins of the western Pacific. We hope to achieve an order of magnitude increase in the quality and quantity of ocean bottom seismic data, which greatly enhance our knowledge of the structure of the earth's interior.

(3) In addition to the permanently emplaced subareal stations, and the long-term ocean bottom network, we plan to develop a portable subareal array whose size or density can be flexibly varied to suit the requirements of particular projects. This mobile array will greatly enhance our ability to study particular regions in great detail.

(4) We hope to negotiate arrangements whereby the POSEIDON seismic array will be operated cooperatively by the western Pacific and East/ Southeast Asian countries. Data from the POSEIDON network will be exchanged freely in accordance with procedures to be set up by the FDSN.

The proposed POSEIDON seismic network will cover the western Pacific and east Asia. The exact area to be covered by the network will be decided after consultation with scientists from other countries in the region. At this time we are envisioning a network with about 40 to 50 stations that will include island arcs such as Japan, the Izu-Marianas, the Ryukyus, Philippines, and Indonesia and marginal seas such as the Japan Sea or the Banda Sea. It also will include the northwestern part of the circum-Pacific earthquake belt, in which a large number of shallow great earthquakes, and large deep focus earthquakes occur. We hope to negotiate arrangements for the cooperative deployment of stations in the USSR. Finally, we also propose to install a seismic observatory at the Showa Base in Antarctica.

Observations on the Ocean Bottom

Since a majority of the region to be covered by the POSEIDON network is covered by water, ocean bottom observations will play an important role in the POSEIDON project. We are currently exploring three possible methods for conducting such observations. (1) Pop-up OBS systems, but with a significantly broader bandwidth, wider dynamic range, and installation time than present systems. (2) Data transmission from ocean bottom seismometers to a buoy, and then via satellite to the POSEIDON Data Center. (3) Real-time telemetry using a dedicated cable (probably optical fiber).

The third possibility is the most attractive, but is also the most expensive. The first possibility represents an extension of existing technology, and also requires relatively small capital investment, although ship-time costs will be considerable. Finally, the second possibility appears to offer a good trade-off between cost and data quality, but requires the development of new technology. We are now intensively studying the various possibilities to determine the best approach, or combination of approaches.

Another possible approach to ocean bottom seismic observations is the use of existing telecommunications cables which are scheduled for retirement from active service. In particular, the present analog Trans-Pacific cable, TPC-1, appears to be a very promising possibility. For budgetary and administrative reasons, and because of considerations of timing, the possible use of TPC-1 for seismic observations is not formally included in the POSEIDON plan, but is being investigated independently by the Earthquake Research Institute of Tokyo University.

International Cooperation

International cooperation is essential both in installing digital broad-band seismographs, and in conducting seismic observation and data analysis. We also plan to develop research and training facilities for visiting students and scientists, possibly through cooperation with the International Institute of Seismology and Earthquake Engineering, Tsukuba, Japan.

Expected Scientific Results

Global seismology is a frontier research area that will greatly contribute to solving a wide range of problems in earth science. One of the major goals of global seismology is determining the earth's internal structure, and inferring the pattern of mantle convection on a global scale. Detailed investigations of mantle convection, the driving forces of plate tectonics, the mechanism of earthquake generation, the state of stress in the mantle and the chemical composition and physical properties of the earth's interior will be assisted by analyses of data from the POSEIDON network and other broadband digital networks. Thus the ongoing International Lithosphere Program, and future programs for geophysical observations of the earth's interior, will be provided with invaluable data.

UNITED STATES NATIONAL SEISMOGRAPH NETWORK

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Denver, CO 80225

Abstract

The USGS National Earthquake Information Center has planned and is developing a broadband digital seismograph network for the United States. The network will consist of approximately 150 seismograph stations distributed across the contiguous 48 states and across Alaska, Hawaii, Puerto Rico, and the Virgin Islands. Data transmission will be via two-way satellite telemetry from the network sites to a central recording facility at the NEIC in Golden, Colorado. The design goal for the network is the on-scale recording by at least five well-distributed stations of any seismic event of magnitude 2.5 or larger in all areas of the United States except possibly part of Alaska. All event data from the network will be distributed to the scientific community on compact disc (CD-ROM).

Introduction

The frequency of occurrence, geographical distribution and magnitude of earthquakes are important parameters for assessing the seismic hazard of a region and for establishing the design and construction criteria for critical facilities. These parameters are known collectively as the seismicity of a region and can only be determined through the operation of seismograph networks. For many years, scientists and government agencies have recognized the need for a high quality National Seismograph Network in the United States. Such a network is now becoming a reality through a cooperative effort between the U.S. Geological Survey (USGS) and the Nuclear Regulatory Commission (NRC). The network is being installed and will be operated by the USGS. The NRC is providing funds to the USGS for the completion of the network east of the Rockies.

The network will consist of approximately 150 seismograph stations distributed across the contiguous 48 states and across Alaska, Hawaii, Puerto Rico, and the Virgin Islands. This network should provide the capability to detect, locate, and quantify the energy release of earthquakes of magnitude 2.5 and larger in all states except possibly part of Alaska. This capability to characterize earthquakes will be greater than exists today in most parts of the United States. However, the U.S. network will not, even when complete, eliminate the need for additional very dense networks of seismograph stations in certain specific locations. Such dense local networks exist today in areas within the United States (for example, in parts of California, Utah, and around the city of St. Louis). The purpose of these dense local networks is to detect earthquakes down to very low magnitude levels (below the 2.5 threshold for the national network) and to achieve very high location accuracy. The dense local networks are thus targeted against a few specific identified seismic risk areas with the objective of acquiring data important for research in subjects such as earthquake prediction and ground motion estimation.

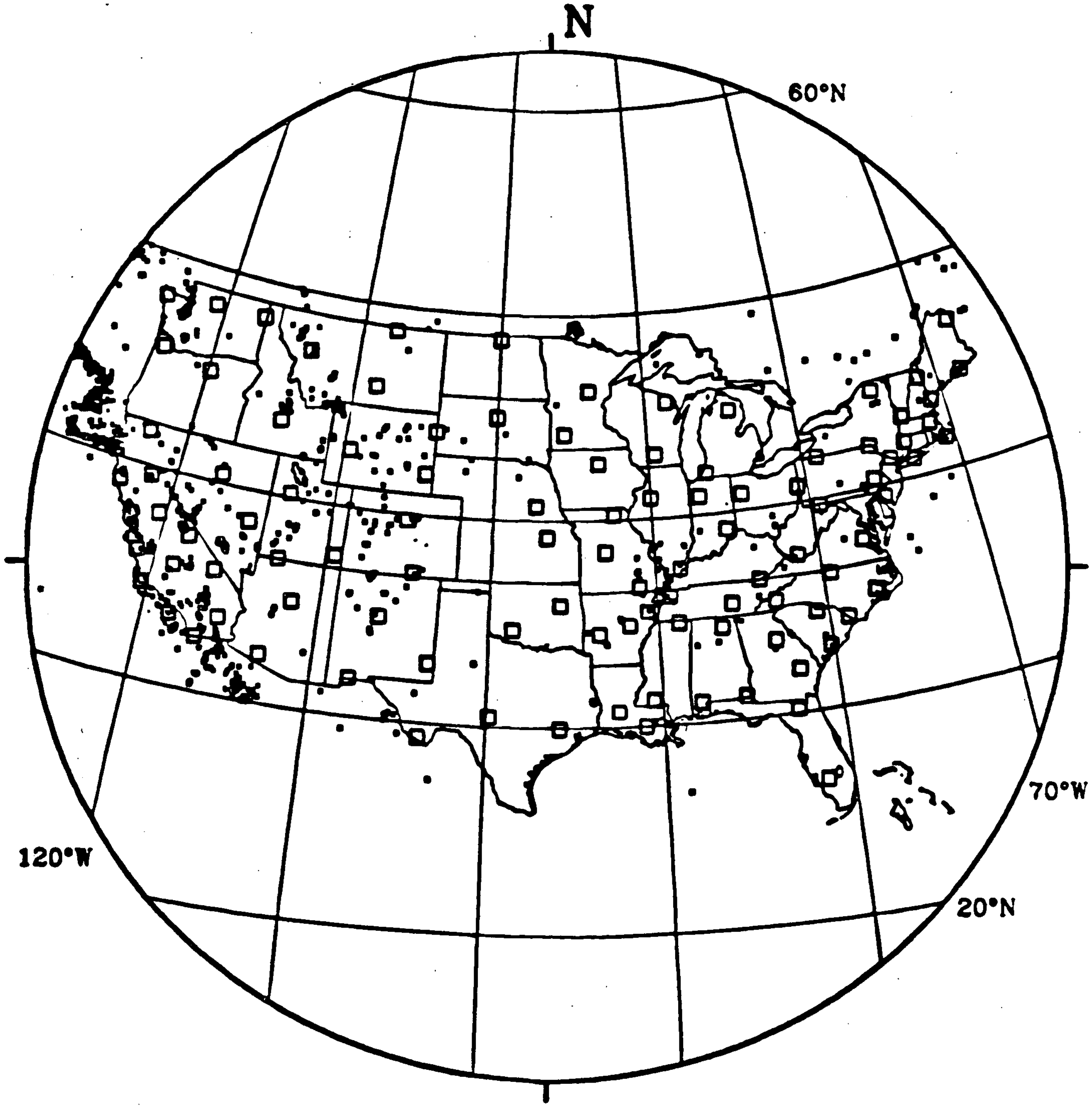
Data from the national network will be of very high quality and will provide, for the first time, near uniform coverage to fairly low magnitude levels for the entire country. All future seismic studies of the United States, including those studies using data from the very dense local networks, can be expected to rely very heavily on the high quality data base obtained from the National Network. The relationship between the National Seismograph Network and the dense local seismograph networks can therefore be considered complementary in nature. Proposed station locations for the national network stations are shown in Figures 1 through 4.

Data Distribution

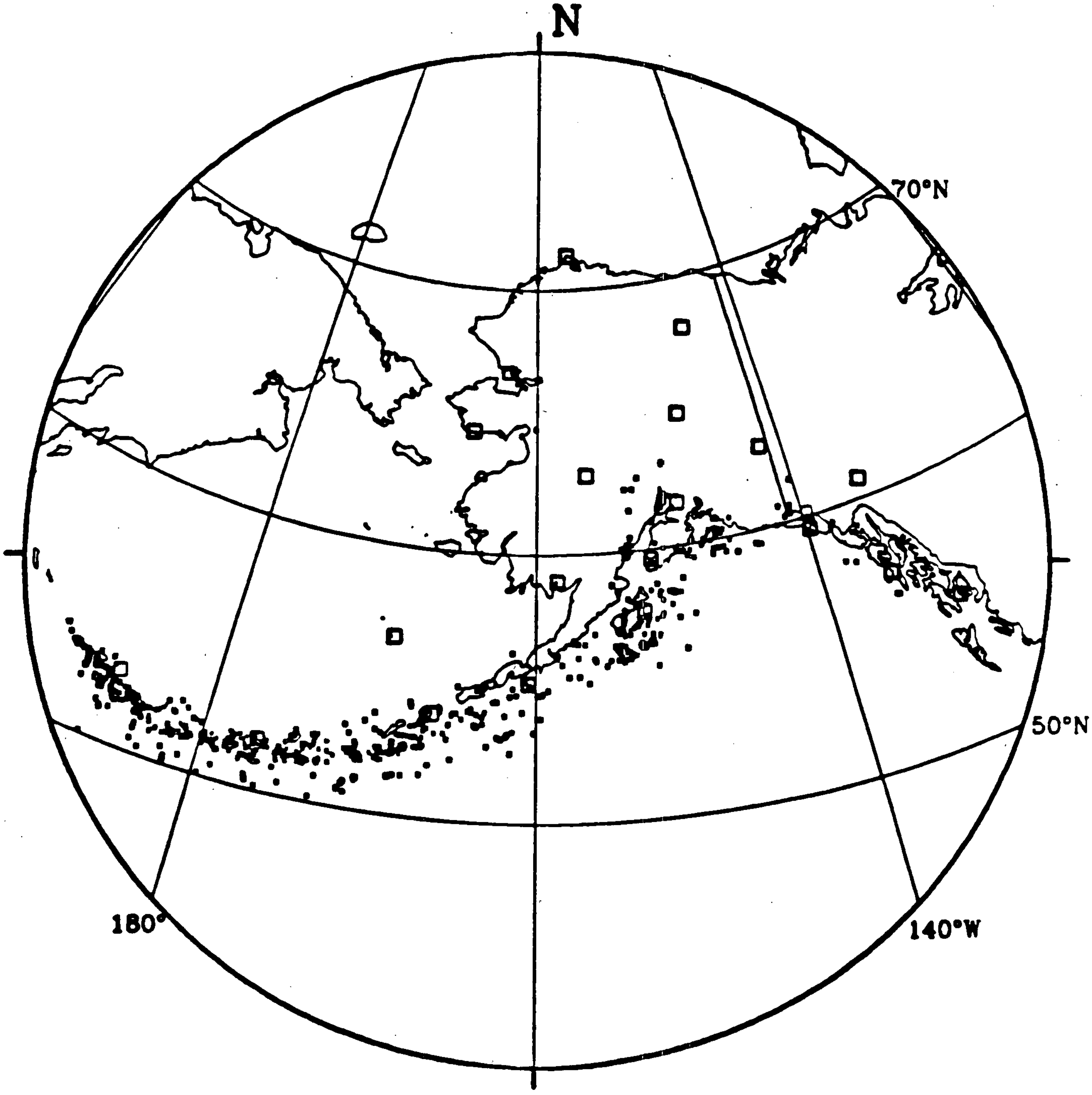
Rapid access to data from the National Seismograph Network will be provided to the scientific community by the USGS. Waveform data for earthquakes will be available in near real-time through a high-speed dial-up capability into the event waveform data base at NEIC. The USGS will provide a free (800) telephone number for this purpose. Waveform data for all earthquakes recorded by the National Network will also be provided by the NEIC on Compact Disc-Read Only Memory (CD-ROM). Satellite links to participating local networks will also be provided.

Processed results for the National Network in the form of epicenter bulletins will be available as part of the NEIC's Quick Epicenter Determination (QED) program. The QED bulletin is already available through a dial-up system to the NEIC computers with a free (800) telephone number. The QED information is also transmitted over the World Meteorological Organization (WMO) communications channels to countries all around the world. The NEIC also now makes the final monthly listing of epicenters available in each issue of the Seismological Research Letters of the Seismological Society of America.

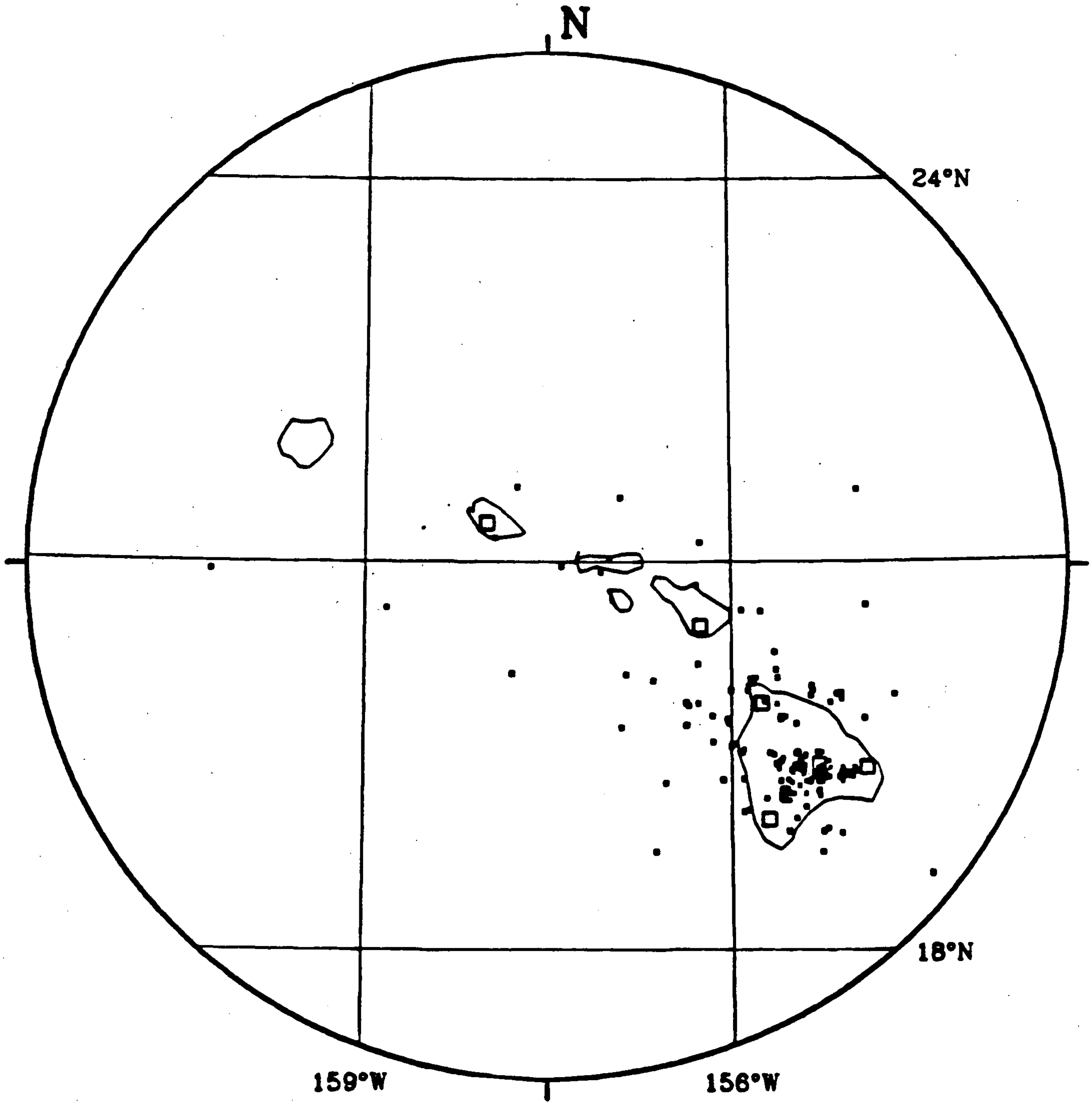
UNITED STATES DIGITAL SEISMOGRAPH SITES AND SEISMICITY



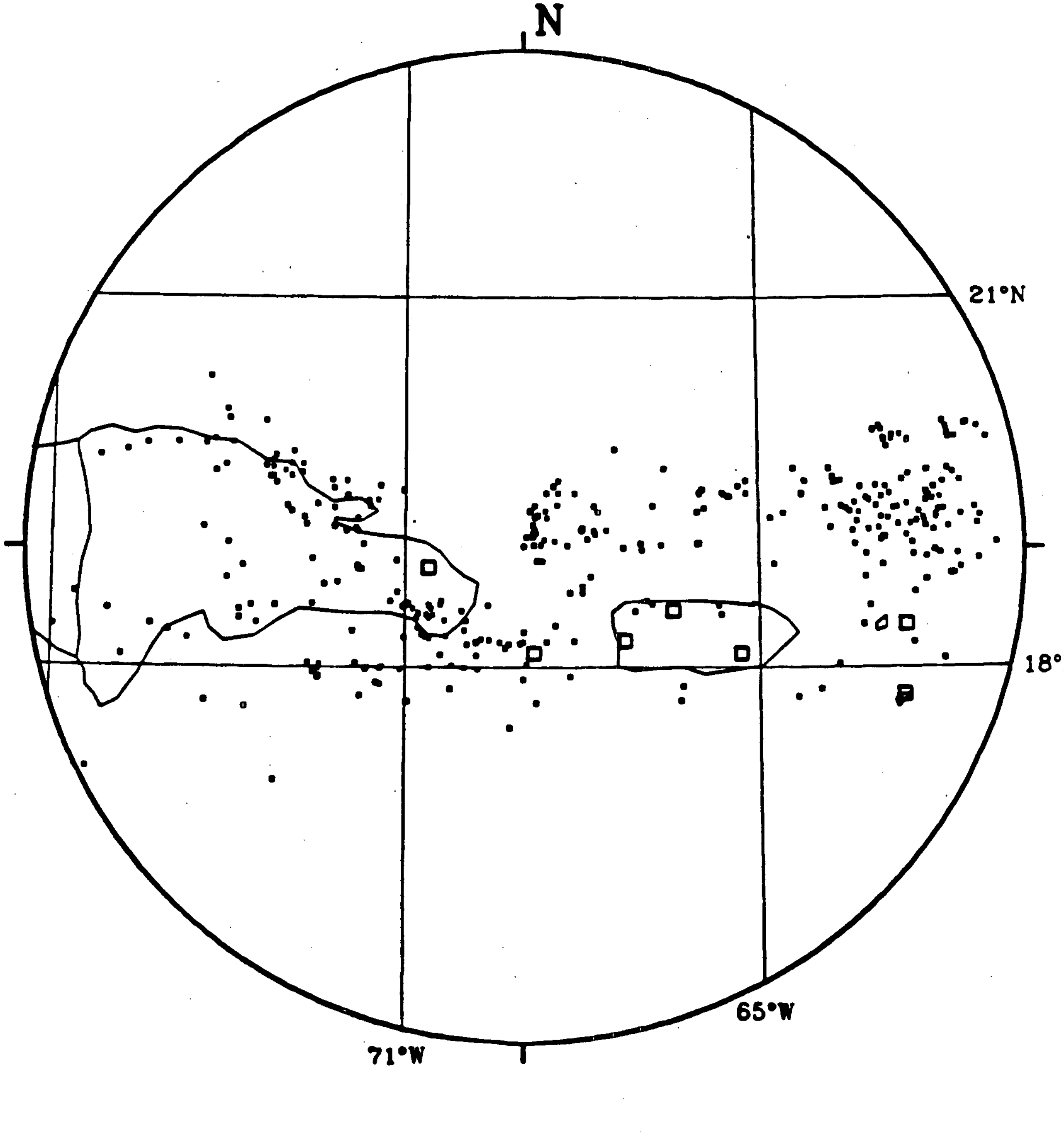
ALASKA
DIGITAL SEISMOGRAPH SITES
AND SEISMICITY



HAWAII
DIGITAL SEISMOGRAPH SITES
AND SEISMICITY



PUERTO RICO
DIGITAL SEISMOGRAPH SITES
AND SEISMICITY



THE ROLE OF OCEAN BOTTOM OBSERVATORIES IN THE GLOBAL SEISMIC NETWORK

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IRIS
Arlington, VA 22209

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.

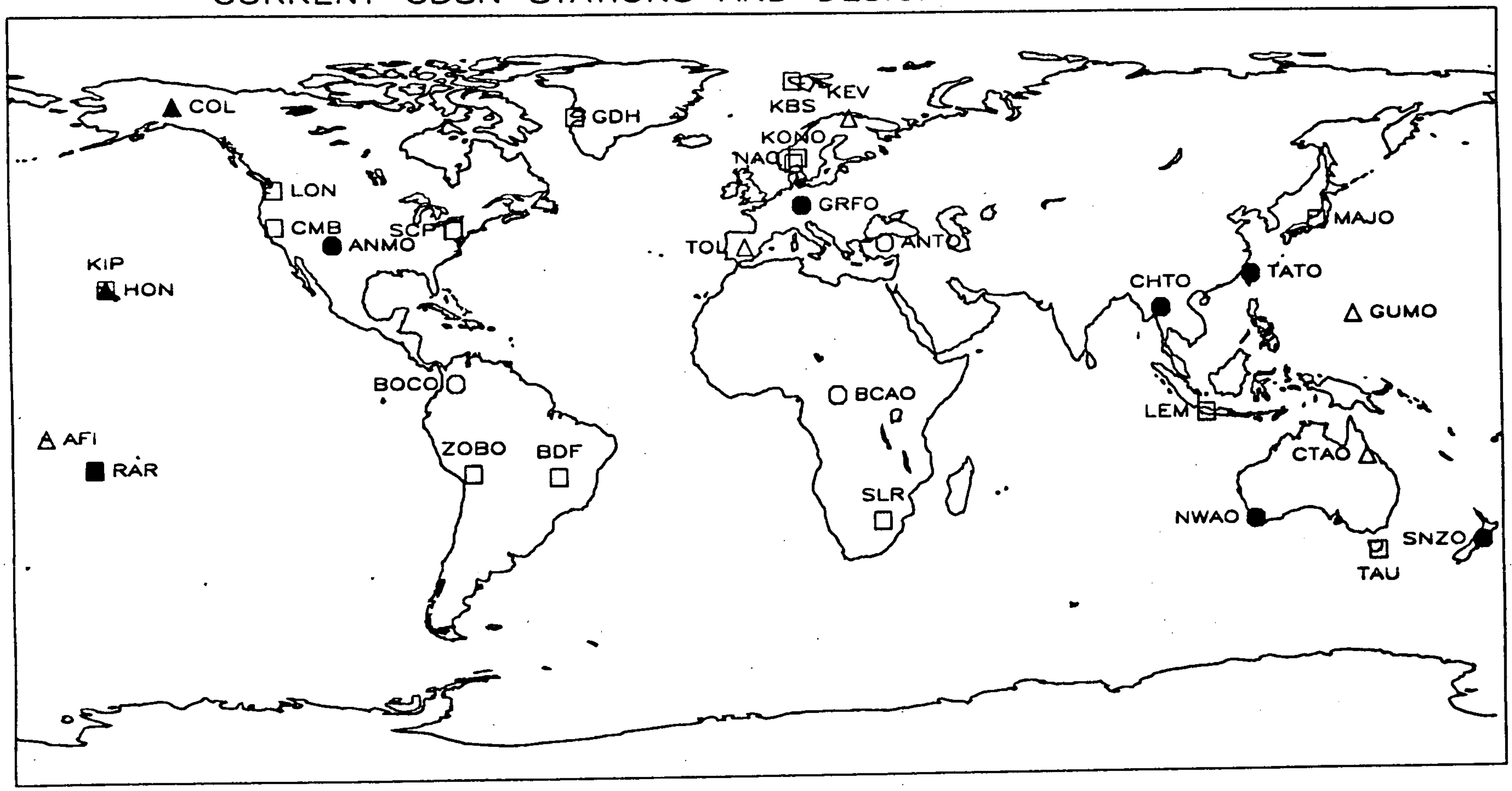
Beyond the fundamental question of the earth's structure, the non-uniform seismic coverage of the planet poses problems in the study of seismic sources, both natural and man-made. The present continental and island siting of seismic stations leads to large gaps in coverage, and the lack of adequate coverage 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.

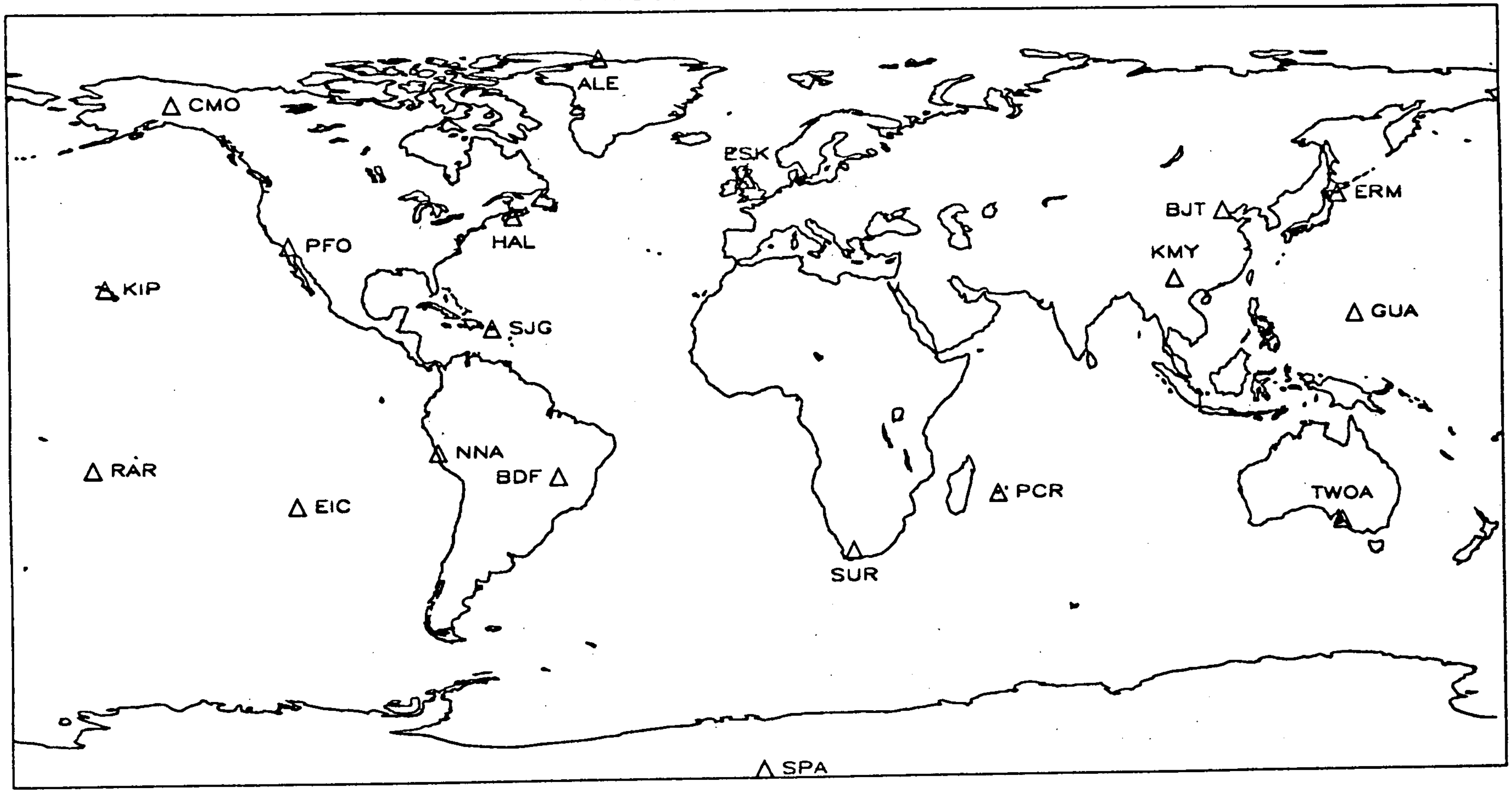
High-resolution seismic studies of the Earth would be advanced by ocean bottom or sub-bottom observatories sites throughout the oceans, and these advances would affect many fields in the geosciences. The seismic structure of the Earth's interior relates to problems of geodynamics and mantle convection. 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 with 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.

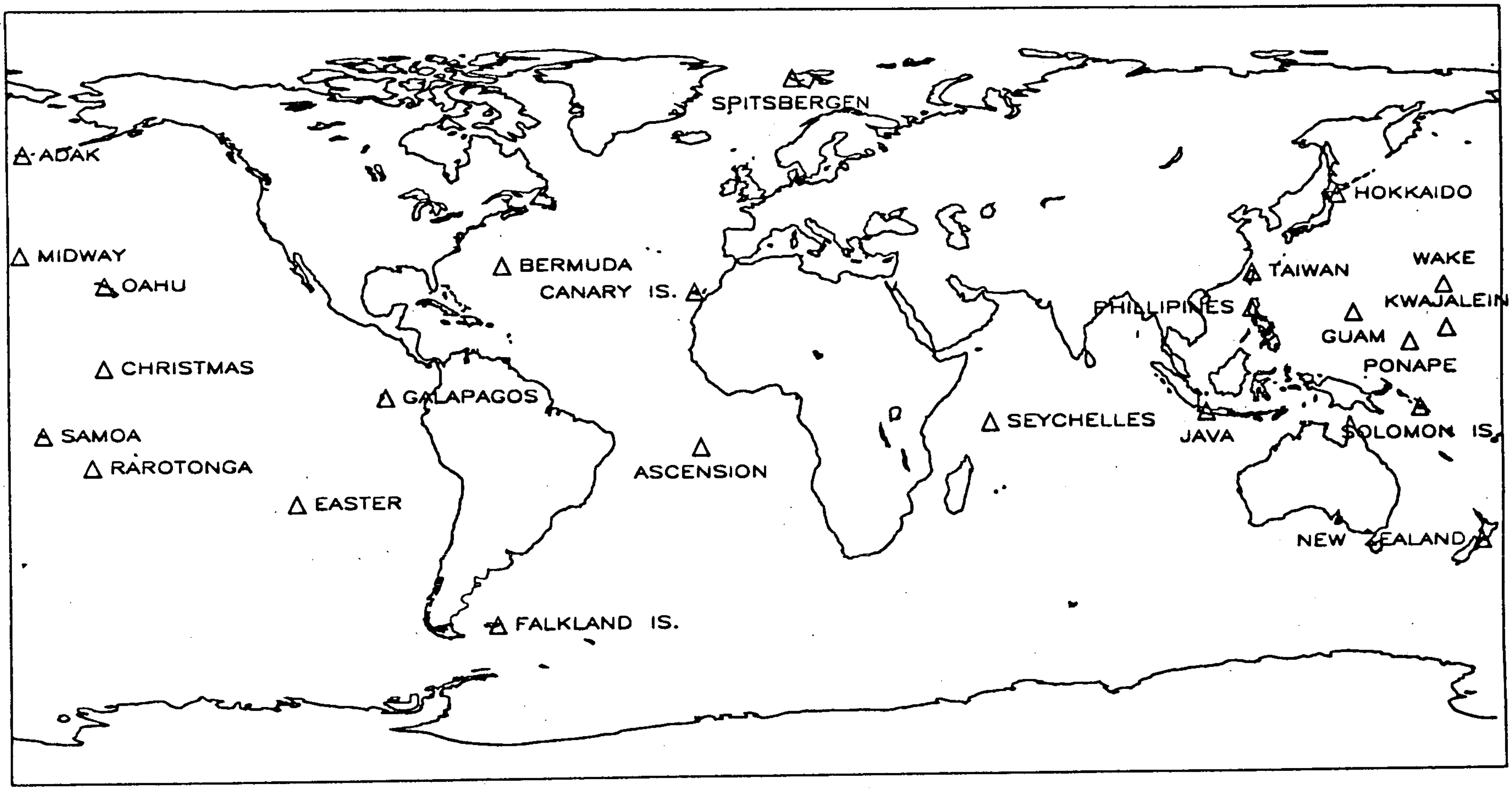
CURRENT GDSN STATIONS AND DESIGNATED IRIS-2 SITES



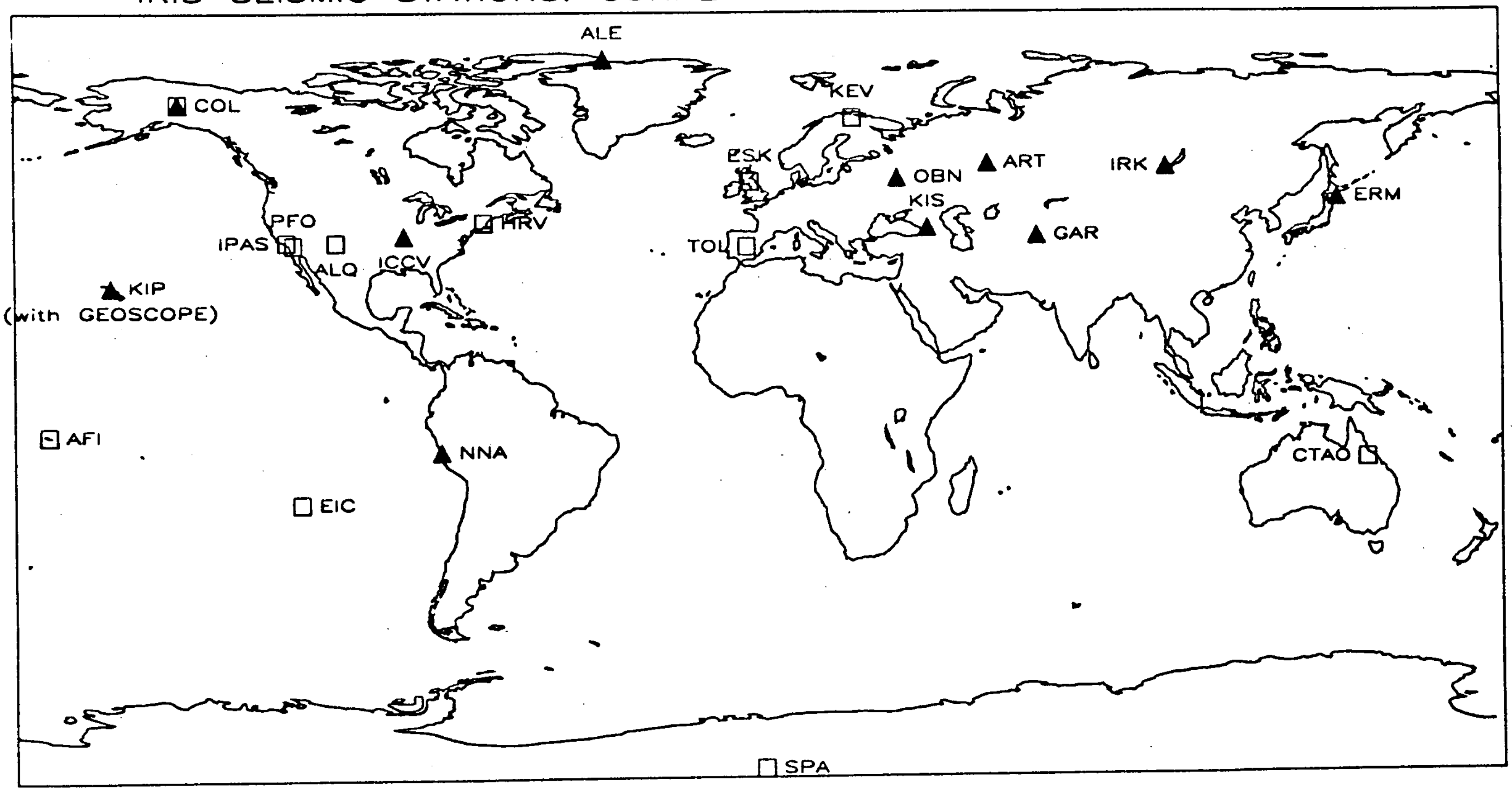
PROJECT IDA SITES



IRIS ISLAND SITES



IRIS SEISMIC STATIONS: CURRENT SITES AND 1988 DEPLOYMENTS



MEDNET - THE ITALIAN BROAD-BAND SEISMIC NETWORK FOR THE MEDITERRANEAN

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Istituto Nazionale di Geofisica, Roma

Broad-band seismology is assuming an ever more important role in modern seismology. Initiatives for deploying regional and global arrays of broad-band instruments are undertaken or proposed by many countries, and in the near future the Earth will be monitored by a global network of about 200 of these seismographic stations.

Advancements in seismometry and electronics have made this step feasible. One would ideally record with a seismograph the Earth's displacement in the whole frequency spectrum, which, for an event at teleseismic distances, range between a few hertz and a fraction of a millihertz. The amplitude of ground motions ranges over many orders of magnitude in the different frequency bands. The sensor-acquisition system has to cope at the same time with the long period free oscillations of the Earth - with characteristic amplitudes of the order of one micron - and with high-frequency, high-amplitude signal. This requires linear response over a very wide dynamic range.

This level of performance was once unattainable, and seismologists were forced to record seismic signal on band-pass instruments. Since 1977 the introduction of digital instrumentation has made it possible to reconstruct the broad-band signal from short and long period channels recorded separately. This is however a time consuming, imprecise and delicate task, and it has been performed only in limited cases. Nonetheless, the scientific benefits of having access to the whole spectrum of the Earth's displacement have become immediately obvious, and a major effort has been undertaken to make broad-band seismology possible.

A new class of seismometers characterized by a very wide response spectrum has become available recently. At the same time, the first commercial analog-to-digital converter with 24-bit resolution has been introduced last year. This, together with improvements in mass storage of data, increased capabilities of microprocessors, higher technology and lower costs in telephone and satellite telemetry, has made possible to design a seismographic station with the required broad-band specifications. Increased interest and funding in geophysics have fostered this effort.

International Framework

In 1986 the Federation of Digital Broadband Seismograph Networks (FDSN) was founded and officially sanctioned by the International Union of Geodesy and Geophysics (IUGG). Its duties are to coordinate on a worldwide basis the national efforts for the definition of the required technical standards (dynamic range, bandwidth, signal-to-noise ratio), in the deployment of stations, the collection and distribution of data.

Among the members of FDSN are networks already in the deployment phase: the Global Digital Seismic Network (GDSN) of the US Geological Survey, the GEOSCOPE global French project, the Chinese Digital Seismic Network (CDSN). With different geographical coverages and scientific goals, they are going to upgrade their instrumentation to broad-band standards in the next few years. Other members of FDSN are countries and institutions still in the planning stage of broad-band deployment. Among these is IRIS (Incorporated Research Institutions for Seismology), a consortium of 52 American universities with a goal of 100 broad-band stations globally distributed, POSEIDON

(Pacific Orient Seismic Digital Observation Network), a network of 50 stations plus a number of ocean bottom seismometers proposed by Japan to cover Southeast Asia and the Western Pacific, the Canadian project CANDIS and the national network planned by the German Research Council.

To coordinate the efforts of European countries, and to provide a centralized data center, in 1986 was founded ORFEUS (Observatories and Research Facilities for European Seismology), which is entering now in the funding stage for the data center and counts among its members countries involved in international projects (France, Holland, Italy) and national ones (Sweden, Germany, England, Belgium).

The MEDNET Project

The Istituto Nazionale di Geofisica has played an important role in the development of the new generation of broad-band seismic instruments, firstly assembled at the Department of Geological Sciences of Harvard University. Of the three prototypes built, two are now located at the Harvard Observatory and one is in the Seismic Observatory in L'Aquila, and is operated by ING.

Broad-band seismology is now ready to enter in the phase of worldwide deployment, and ING has undertaken the project of a network of 10 broad-band stations covering the Mediterranean area. Italy holds a central location in the Mediterranean, a region characterized by strong seismicity and active, complex tectonics, which, at present, are only in part monitored and understood for lack of sufficient data. The existing station coverage along the Mediterranean coasts is very uneven, and digital seismometry is almost absent.

The ING initiative, termed MEDNET, is motivated both by research interest and by considerations on seismic hazard monitoring. On a regional scale, it will allow to define the structure of the Mediterranean region to a higher detail, and to study properties of the seismic source for intermediate and large events in the area. At the same time, it will provide instrumental coverage of a large area as a contribution for a global monitoring of the Earth.

To reach these goals, MEDNET has been designed following the highest technical standards available today: STS sensors, 24 bits ADC, 140 dB dynamic range, real-time telemetry. The geometry is designed to provide a homogeneous coverage of the Mediterranean area, with an ideal spacing between stations of about 1000 km.

The proposed schedule calls for the deployment of 3 stations by the summer of 1988 (in Italy and North Africa) and 3 more by the summer of 1989. At the present stage (April 1988), only the Seismic Observatory in L'Aquila is operational, while two more sites are being finalized. Formal contacts have been opened with scientists from North African countries, for the selection of suitable sites. The location of the stations shown on the map is to be considered only indicative, and is presently being discussed with the countries involved in the project.

The cooperation among Mediterranean countries will be instrumental to fulfill the high scientific potential of this project, in choosing the installation sites, operating the stations and in the following analysis of the data. Each institution will have direct access to the data collected on its territory and will receive the waveforms of the entire network. Cooperative plans on regional and national research projects will be encouraged.

Broad-band stations already exist in the Mediterranean area and other are planned by different institutions (Toledo, Spain; Ankara, Turkey; Zurich, Switzerland; St. Sauveur, France; Bar Giyyora, Israel). We will avoid as much as possible the deployment of instruments in these areas, maintaining at the same time the planned characteristics of technical and geographical homogeneity. Rapid exchange of data and direct access to stations in areas of common interest will eliminate in the near future conflicts of interest among different institutions and will complement the data collected by MEDNET.

To this purpose, ING is participating in international cooperation programs, both on a regional and global basis. It is now a member of ORFEUS and of FDSN, and within these institutions it cooperates with other national members on a number of pressing issues: the selection of station sites, the definition of common format codes for data exchange, the organization of data centers for the collection and distribution of data, the codes for direct access to seismic stations.

ING supports the formation of a worldwide pool of 10-12 VBB stations, well distributed around the globe and accessible via telemetry, to serve as a first-order tool for real-time monitoring of seismicity. Long-period data from this limited number of installations would guarantee control of the geometry of the seismic source process, eliminating the need for fast distribution of all worldwide data and the deployment of stations motivated only by national interest on seismic hazard.

Scientific Goals: Structure of the Mediterranean

Twenty years after the "plate tectonics" revolution, the Mediterranean area still represents a challenge to the understanding of tectonophysicists. Plate tectonics provided a new perspective and consistent view of the surface expression of the dynamic processes active inside the Earth, but its application to the Mediterranean is still controversial. The region is commonly seen as very complicated, or even considered a place where plate tectonics cannot be applied.

This fact alone justifies any effort made to unravel the physical structure and behavior of the Mediterranean lithosphere. A concerted attack is to be moved by Earth scientists to gather new data to understand the system. Space techniques in geodesy now provide a way to measure and monitor displacement on a continental scale. New data based on Mediterranean measurements are appearing now as a result of international projects. In addition to measuring contemporary deformations, we need to map the state of stress in greater detail, as well as we need a better knowledge of the structure. Currently, the only way to do this passes through seismology.

Seismology has proven to be an essential tool in the study of the Earth. In the past decade seismic tomography has grown of importance, and is now applied at many different scales, from local to global. The resolution that one achieves is limited by the amount and distribution of the available data. For regional and global studies, earthquakes provide the energy source. Since their distribution is highly concentrated in narrow seismogenic areas, we need to increase the station density.

One of the main goals of the project MEDNET is to provide data to improve our knowledge of the structure underlying the Mediterranean. The resolution achieved by the first studies (appeared in the last few years) should be highly improved. Different portions of the seismic signal can be employed for the purpose. The travel time of body-waves is a very sensitive parameters, which yields precise answers when the data coverage is dense enough. Travel time, however, is just one parameter of the seismogram, which contains much more information. Modelling the whole waveform has been widely used in structural inversions. So far, however, long-period data have been mainly used. Their resolution is limited. To sharpen the image, fitting of broad-band seismograms is necessary: triplications in the travel-time curve, for instance, are otherwise almost undetectable.

Surface waves represent the most common choice in upper mantle studies. Efficient nonlinear waveform inversion techniques for heterogeneous media are now available, and also surface wave coda has begun to be used in studies which model the effect of scattering. The station distribution of MEDNET was planned to achieve the optimal path coverage of the area under study, considering also other existing broad-band stations.

Scientific Goals: The Seismic Source

A better understanding of the structure underlying the Mediterranean is instrumental to study the seismic source processes of events in the area. An obvious and extremely important motivation for the study of source mechanics is the comprehension of earthquake generation. This has a particular urgency for many countries in the Mediterranean area (like Italy, Yugoslavia, Greece, Turkey and Algeria), where high population density in seismogenic areas is responsible for numerous casualties and damage.

The study of the seismic source is conveniently done using a global network. In that case, however, the magnitude threshold is too high for all but few Mediterranean events. A way to decrease that threshold is to deploy seismographs at shorter distances. From this point of view, MEDNET represents a ring of sensors to monitor the area from nearby, concentrating on events of moderate size. The long-period portion of the spectrum can be used to retrieve the source mechanism. Finer details, like the reconstruction of the source time function, the details of the fault area and of the rupture propagation can only be retrieved using the broad-band spectrum. As for structural studies, broad-band data is therefore essential to improve the current state of knowledge.

An important factor to consider in the event of a seismic crisis is rapidity in gaining information. The real-time access to the data is one of the most important requisites of MEDNET, and will allow prompt evaluation of the details of the seismic source, providing valuable information for the deployment of local arrays, monitoring of the seismic crisis and assessment of seismic hazard.

Source studies of medium-size events are of extreme interest also from a purely academic point of view. So far, broad-band studies of the seismic source have concentrated on very large events, which generally involve a complex rupture process with the breaking of more asperities. The rupture of a single asperity is the basic process in every earthquake, and yet, for lack of sufficient data, it is a very mysterious process. MEDNET will allow us to study in detail the rupture process of events in the Mediterranean, gathering precious information in our quest to understand earthquake genesis.

The control of how the crust deforms is also of primary importance. Earthquakes are the expression of the stresses accumulated in the deforming lithosphere. Evaluation of the source mechanisms allow us to estimate the amount and geometry of the deformation taking place through earthquakes. This knowledge is of fundamental importance to understand contemporary tectonics, and is therefore still intimately connected to the origin of earthquakes.

Technical Specifications

MEDNET will be formed by a network of state of the art very-broad-band seismographic stations, based on the STS sensor and the QUANTAGRATOR 24-bit AD converter. This configuration will allow flat velocity response in the frequency band from 5 to 0.003 Hz, with a full dynamic range of more than 140 dB (Figure 2). It is capable of recording on scale ground acceleration over 9 orders of magnitude, from a few nanogals at 300 sec period to nearly 0.1 g at 10 Hz. The station (Figure 3) is controlled by a 68000-based microcomputer, the station processor, which controls the sensors and handles the digital data stream and the remote communications. The station will allow local storage of data and remote access to a circular random-access buffer, via a modem or a packet-switching network.

The sensor derives from the well-known STS-1 force-balanced leaf-spring sensor. The mass position is always kept close to its center by an electrically generated restoring force. In this way, linearity and stability are requirements of the electronic circuitry rather than of the mechanical assembly. The feedback circuit is modified to expand the response at the low-frequency end (STS-1/VBB). Free-mode signals are resolved without affecting the higher-frequency response.

The analog to digital conversion is performed by a 24-bit QUANTAGRATOR for each component, at a rate of 20 samples per second. The high dynamic range could also be attained using a lower-resolution converter and gain ranging, but in such a way signal distortion and noise are introduced. Also, linearity is maintained, permitting to restore low-amplitude data superimposed on high-amplitude signals without affecting the sensitivity. This for instance allows to extract on-scale normal-mode data even when the high-frequency, high-amplitude signal from a local small earthquake is present. The study of regional events is also possible. As an example, this system has recorded magnitude 3 events at a few kilometers epicentral distance at less than 1 percent of full scale without disturbing very-long period recording. The system should be able to record a magnitude 6 event on scale at about 50 km.

For each sensor, a data stream from the digitizer consisting of 20 samples per second contains the full VBB signal (Figure 4). The station processor applies digital FIR low-pass filtering in order to extract from incoming data long period (LP, 1 sample per second) and very-long period (VLP, 0.1 sps) streams. Data are then compressed with a first-difference algorithm and time-marked. The system records continuously all seismic data, thus eliminating the possibility, inherent in any trigger algorithm, to lose important data. Permanent storage is on 120 Mb streamer tape for up to a few weeks of operation. The adoption of large-capacity mass storage optical disks will allow operator-free functioning for several months. This will be particularly convenient for the most remote sites. Temporary storage is made on a circular random-access buffer on a reliable Winchester disk. This enables the retrieval, from a remote location, of event data.

Continuous digital telemetry will enable monitoring from the data collection center in Roma. Italian stations will be connected via a dedicated telephone line, capable of transmitting all the VBB data. Farther locations will be telemetered either through telephone links or by using the European satellite data transmission facility (DCP). This will employ several 100-baud transmission channels of the meteorological satellite METEOSAT, reserved for scientific usage. Due to the limited baud rate of this link, only long-period (LP, 1 sample per second) streams will be transmitted.

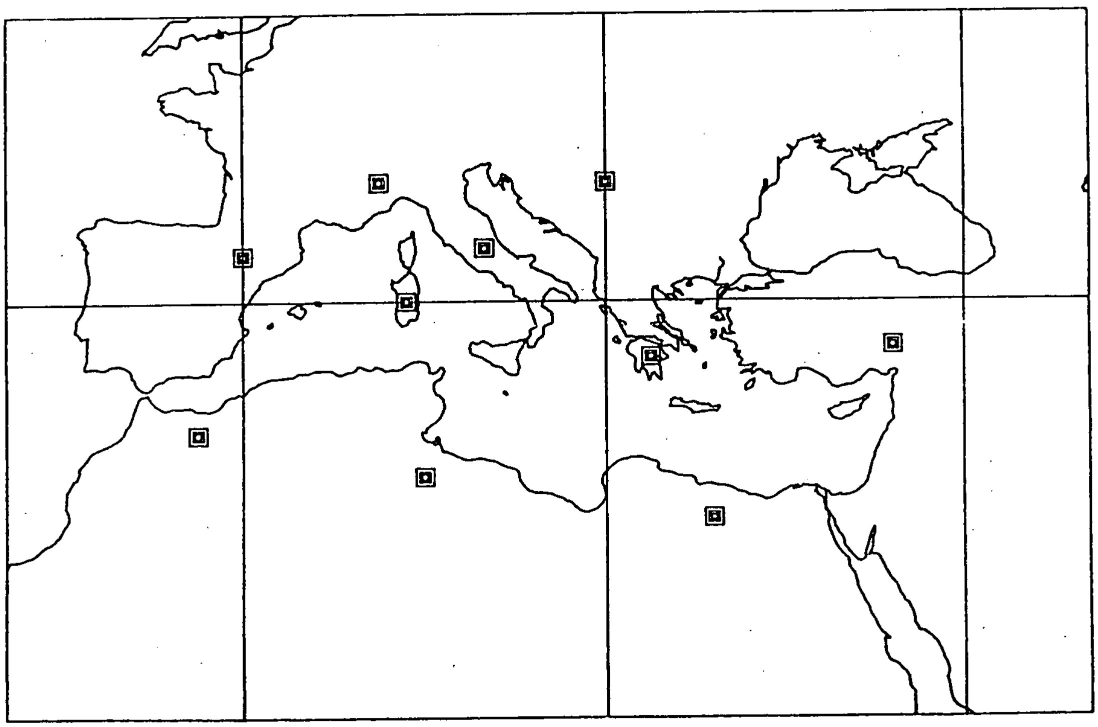
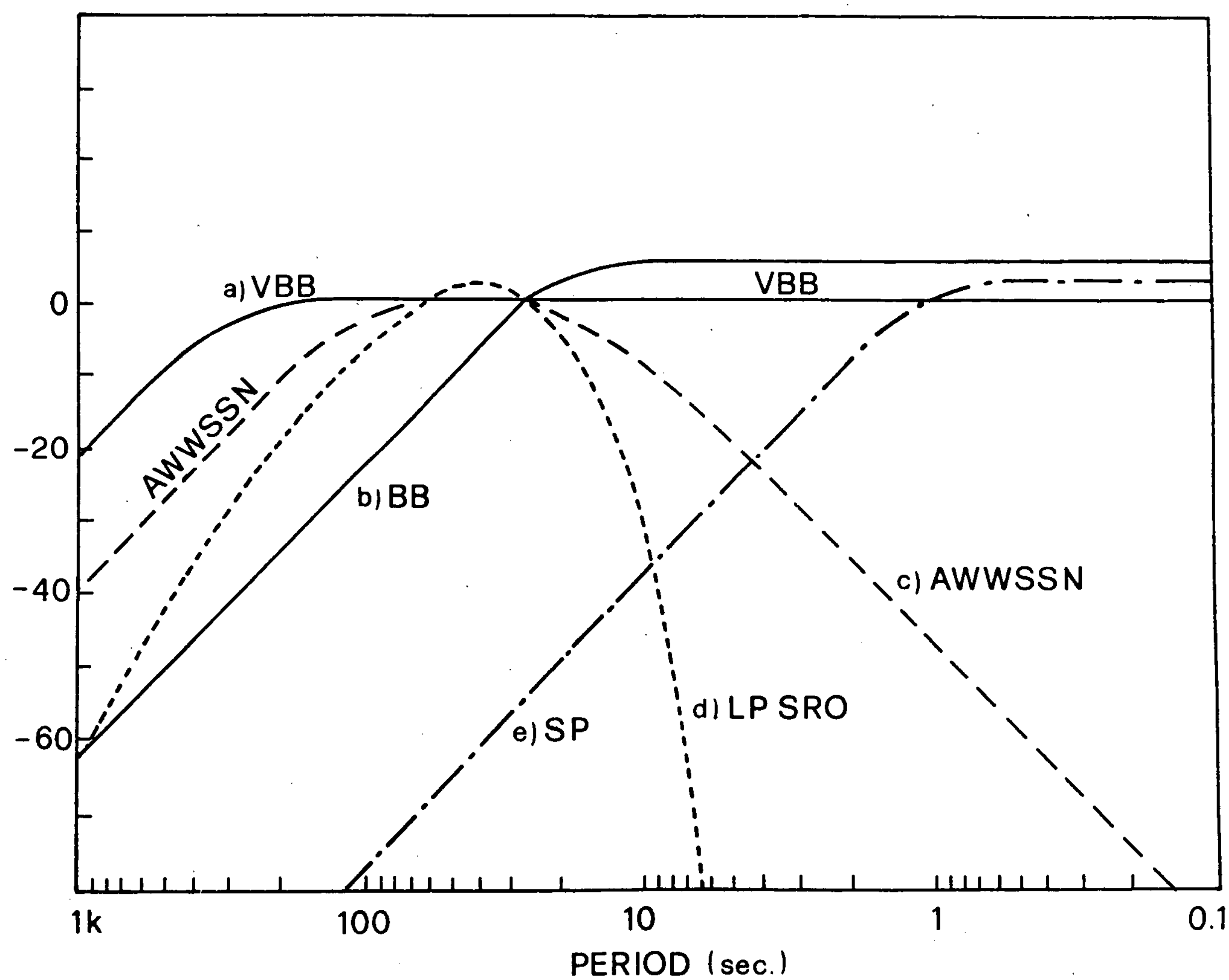


Figure 1



a) VBB
VBB
AWWSSN
b) BB
c) AWWSSN
d) LP SRO
e) SP
0
-20
-40
-60
1k
100
10
1
0.1
PERIOD (sec.)

Figure 2

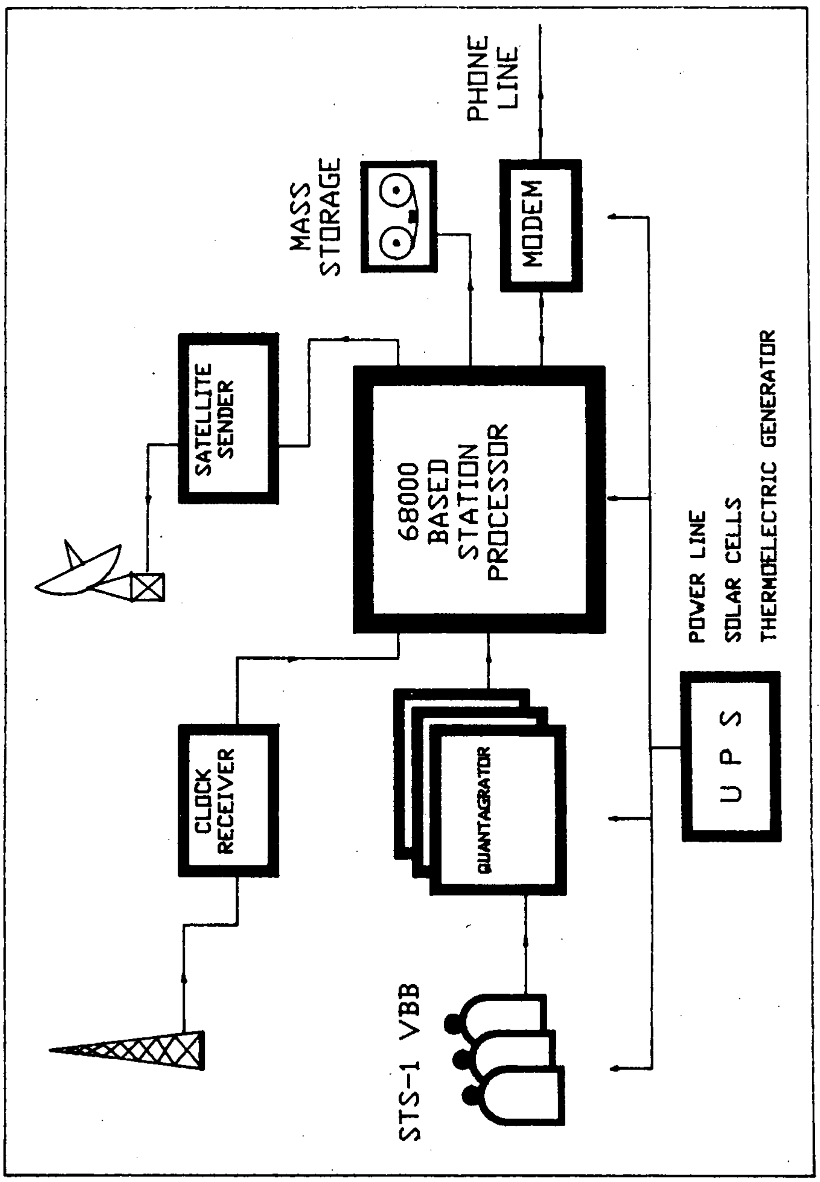

PHONE LINE
MASS STORAGE
MODEM
SATELLITE SENDER
68000 BASED STATION PROCESSOR
CLOCK RECEIVER
QUANTAGRATOR
U P S
POWER LINE
SOLAR CELLS
THERMOELECTRIC GENERATOR
STS-1 VBB

Figure 3

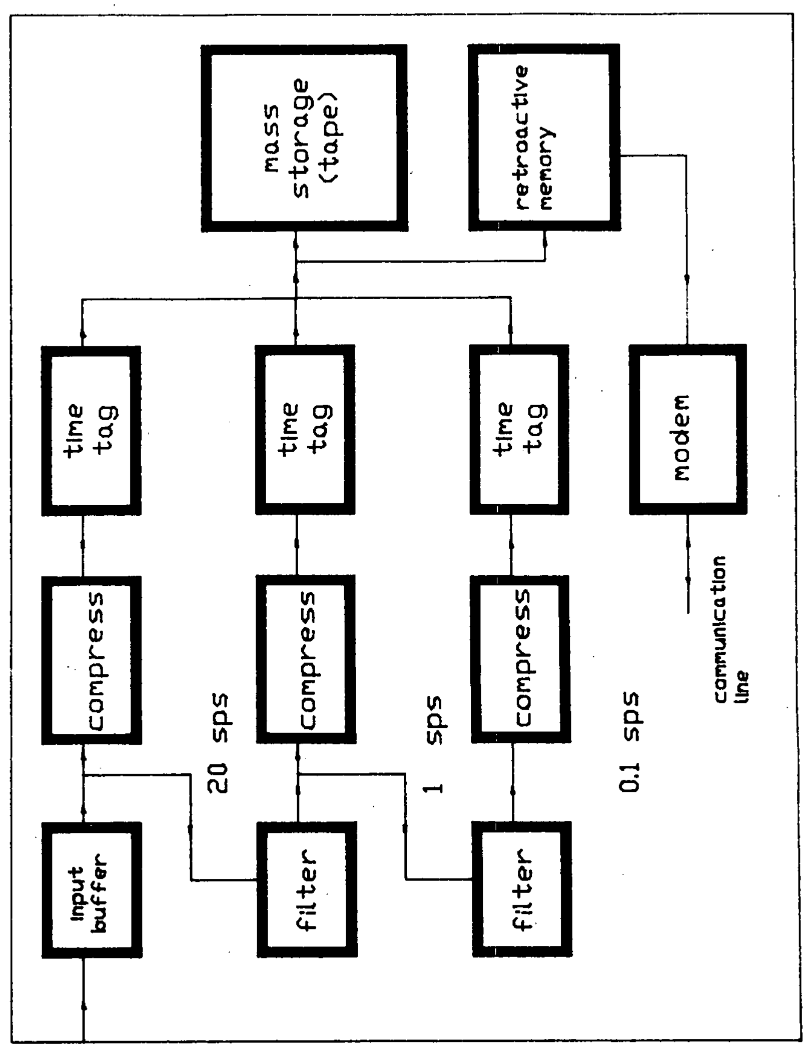

input buffer
filter
filter
compress
compress
compress
time tag
time tag
time tag
mass storage (tape)
retroactive memory
modem
communication line
20 sps
1 sps
0.1 sps

Figure 4

Section B3: Noise

1. Long-Period Seismic Measurements on the Ocean Floor: George H. Sutton, Noel Barstow and Jerry A. Carter

2. On the Relevance of Seafloor Array Experiments to Borehole Seismic Systems: L.M. Dorman, A.E. Schreiner, A.W. Sauter, L.D. Bibee, J.A. Hildebrand and F.N. Spiess

3. Effects of OBS Burial on Ground Coupling and S/N Ratio Enhancement: T. Yamamoto, A. Turgut, M. Trevorrow, D. Goodman and M. Badie

4. Continental Downhole Installation: Charles R. Hutt

5. Seismic Broadband Signal and Noise Levels on and Within the Seafloor and on Islands: M.A.H. Hedlin, J-F Fels, J Berger, J.A. Orcutt, D. Lahav

6. A Comparative Study of Island, Seafloor and SubSeafloor Ambient Noise Levels: M.A.H. Hedlin and J.A. Orcutt

LONG-PERIOD SEISMIC MEASUREMENTS ON THE OCEAN FLOOR

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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 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. Currently we are analyzing OBSS data in the frequency range .001 to 40 Hz to study the characteristics of ULF/VLF ocean bottom noise.

Some results of analyses to date are:

Coherent spectral peaks are observed on vertical seismometers and hydrophones near .01, .06, .14, and .3 Hz;

The peak at .06 Hz also shows strong coherency with horizontal motion perpendicular to the coast;

The peaks at .06, .14, and .3 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 they are predominantly fundamental mod Rayleigh waves propagating toward shore;

The peak at .01 Hz has 5-10 times greater pressure/vertical velocity ratio;

The amplitude of the .01 Hz pressure peak correlates with ocean wave heights observed at nearby coastal stations; and

Tidal bottom currents with amplitudes less than 8 cm/sec produce large long-period disturbances on the horizontal seismometers.

Continuing studies are aimed at identification of noise sources and their time and location dependence; determination of the influence of sub-bottom velocity structure on pressure/velocity relationships for free-running and forced bottom motions; and the influence of the instrument package, itself, on observed signals.

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 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. Power and commands from shore and multiplexed data from OBSS were transmitted via single conductor deep-sea cable. The data were recorded on two 7-channel instrumentation-quality FM tape recorders and on seismograph-type drum recorders and strip chart recorders. All of these data are presently stored at Lamont-Doherty Geological Observatory.

The location of OBSS, on the Delgada depositional fan, is shown in Figure 1. Many local earthquakes from the Mendocino Escarpment to the north were recorded. The OBSS, shown in Figure 2, was not of optimum hydrodynamic design and was especially susceptible to long-period noise on the horizontal instruments produced by long-shore tidal currents. The H1 component, which was oriented perpendicular to the lowering bail (containing the current meter) and parallel to the isobaths, was most seriously disturbed.

Short- and long-period response curves are shown in Figure 3; response is relative to ground placement and pressure for the seismometers and hydrophones, respectively.

The effect of bottom current on the horizontal components is demonstrated in Figures 4 and 5. Note that current amplitudes shown are less than 8 cm/sec; the Qi and Ni identify relatively quiet and noisy time intervals that were digitized for analysis. The spectra for the quietest and noisiest times (Q2 and N2) demonstrate the strong influence of current for all frequencies below the principal microseism peak, i.e., below about 0.1 Hz, with a maximum change of about 20 dB near .04 Hz.

Spectral peaks observed from all local earthquake phases near 5, 6, and 7 seconds for the $H\|$, $H\perp$, and V, respectively, indicate that these are resonant bottom-coupling modes of OBSS and can be used along with moments of inertia and calculated current-derived forces to estimate how much of the observed disturbance is caused by direct influence of currents on OBSS and how much might be from actual deformation of the bottom.

In Figure 6 the power spectrum of vertical velocity during quiet time Q2 is compared with seismic noise on continents (from Aki and Richards). Curves for continents represent noisy and quiet conditions on hard rock. The ocean bottom noise is comparable to continents between about .03 and.10 Hz and above 1 Hz; near maximum between 0.10 and 1.0 Hz and well above maximum below .03 Hz. The OBSS spectrum shown is limited by system/reproduction noise below .004 Hz and we are attempting to improve this condition. The greatest "real" differential occurs near .01 Hz where the OBSS noise is over 30 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 is an important question for this workshop to consider.

Figures 7 through 11 compare long-period pressure and vertical motion. Pressure and vertical velocity spectra, corrected for instrument response, are shown in Figure 7. (Below about 0.004 Hz both curves are limited by system noise.) The pressure spectrum is similar to that obtained more recently by Cox and Webb. The pressure and vertical velocity spectra are quite similar except that the P/V ratio increases significantly below about .02 Hz. Uncorrected spectra and P-V coherency between .01 and .5 Hz are shown in Figure 8. Coherency maxima near .06, .14, and .30 Hz coincide with maxima in the spectra. Corrected phase relations and amplitude ratios for these spectral and coherency peaks appear to be appropriate for fundamental mode Rayleigh waves.

In Figure 9 a P-V 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 as shown in Figure 10. 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.

Pressure and vertical Rayleigh wave signals from an Easter Island earthquake are shown in Figure 11. As expected, the P/V ratio decreases with increasing period. We also note that signal to noise is greater for the vertical motion sensor and improves relative to the hydrophone with increasing period. Noise spectra and coherency among pressure and vertical and horizontal motion perpendicular to shore, between .01 and .5 Hz, are shown in Figure 12. All show relative maxima near .06 Hz ("single frequency" microseisms). Assuming retrograde particle motion the observed phase relationship between V and H⊥ indicate propagation toward shore at the OBSS location (about 160 km off-shore). Although V-H⊥ coherency is not as strong for the .14 and .3 Hz spectral peaks relatively stable phase coherency indicates predominant $\pi/2$ phase difference of the opposite sign from that of the .06 peak. Theoretical results for appropriate velocity structures indicate that at .3 Hz fundamental mode Rayleigh waves have prograde motion and at .14 Hz are also possibly prograde but near cross-over from retrograde to prograde. The most likely estimate at this time is shoreward propagation for all three peaks.

From the preceding it is clear that significant improvements in instrumentation can be realized using current scientific and engineering capabilities. In Figures 13 and 14 we only consider possible improvements in signal-to-noise ratio and signal fidelity by relatively straightforward and inexpensive improvements in instrument-bottom coupling (and instrument-water decoupling!). The "Opihi" sensor package, conceived at Hawaii Institute of Geophysics, is shown in Figure 13. This design is an attempt to make the sensor package approximate a small irregularity on the bottom having roughly matched density and a small vertical cross-section in the water. OBS with high profiles in the water and a large vertical density gradient (between anchor and flotation) are subject to large tilts and current generated noise. A relatively simple procedure for burying the sensor package the depth of a coring pipe is sketched in Figure 14. Matched density and symmetrical design should minimize signal distortion; direct bottom-current pressures are eliminated.

A number of tests to evaluate improvements in signal quality to be obtained by such design and implanting procedures are currently being planned. I **strongly** endorse such tests. Although it appears self-evident that a sensor firmly coupled to hard-rock under the sediment would be optimum for detecting seismic (as opposed to water-born) signals, the additional cost and resultant reduction in observation points must be considered. A combined approach is likely to be optimum.

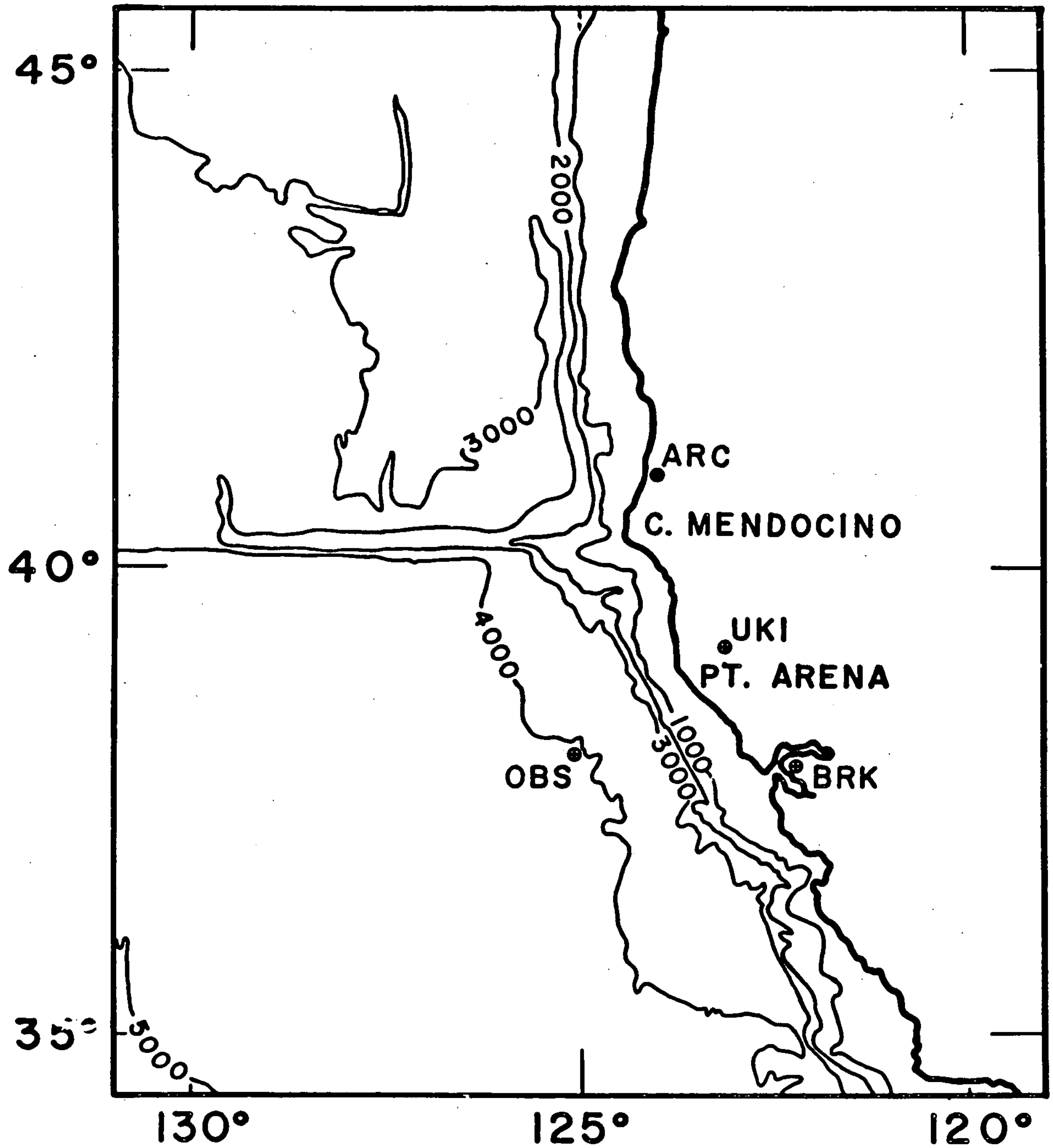
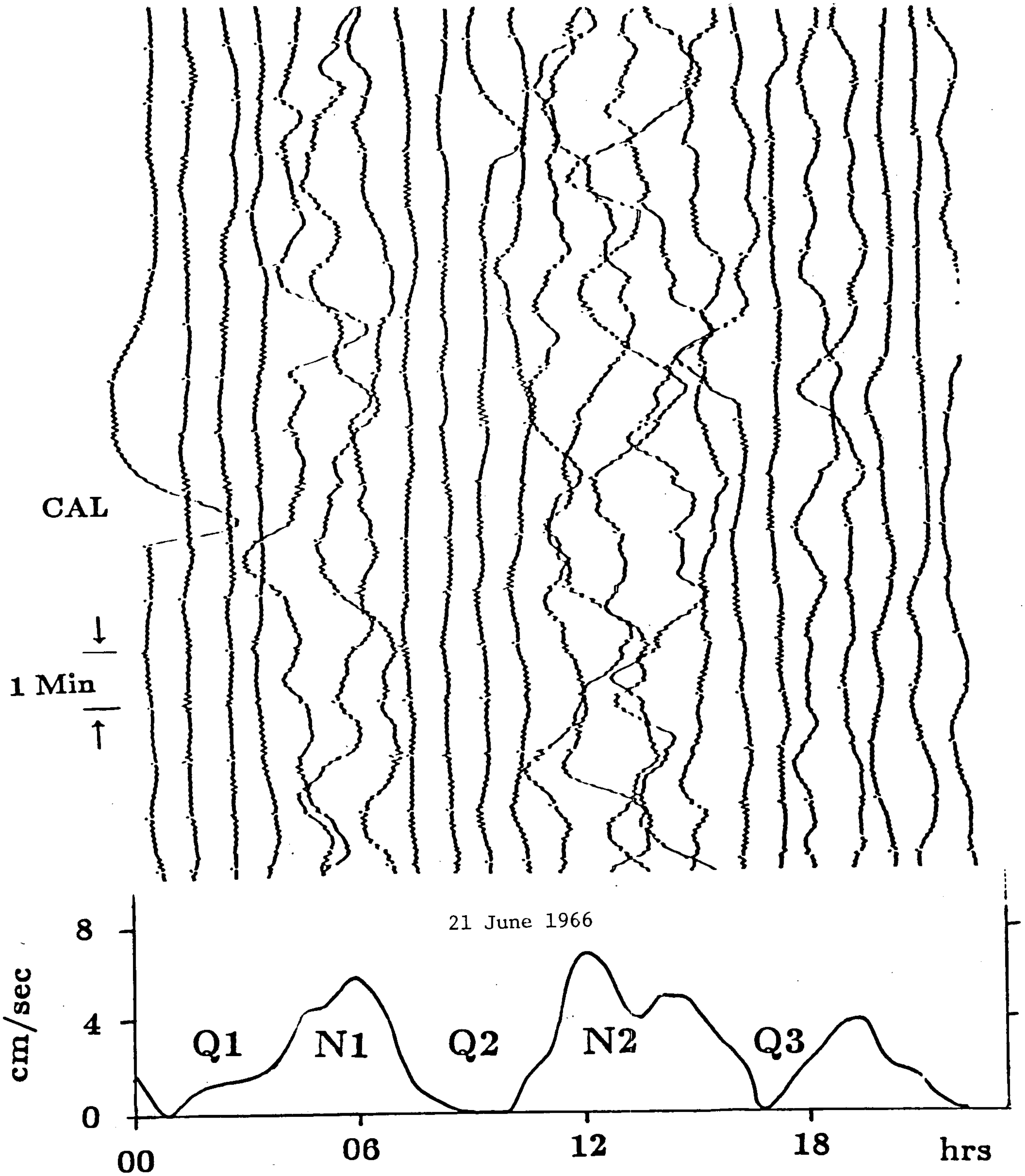

45°
40°
35°
130°
125°
120°
2000
3000
4000
5000
1000
3000
ARC
C. MENDOCINO
UKI
PT. ARENA
OBS
BRK

Figure 1



ALONG-ISOBATH HORIZONTAL MOTION
VS. BOTTOM CURRENT

Figure 4

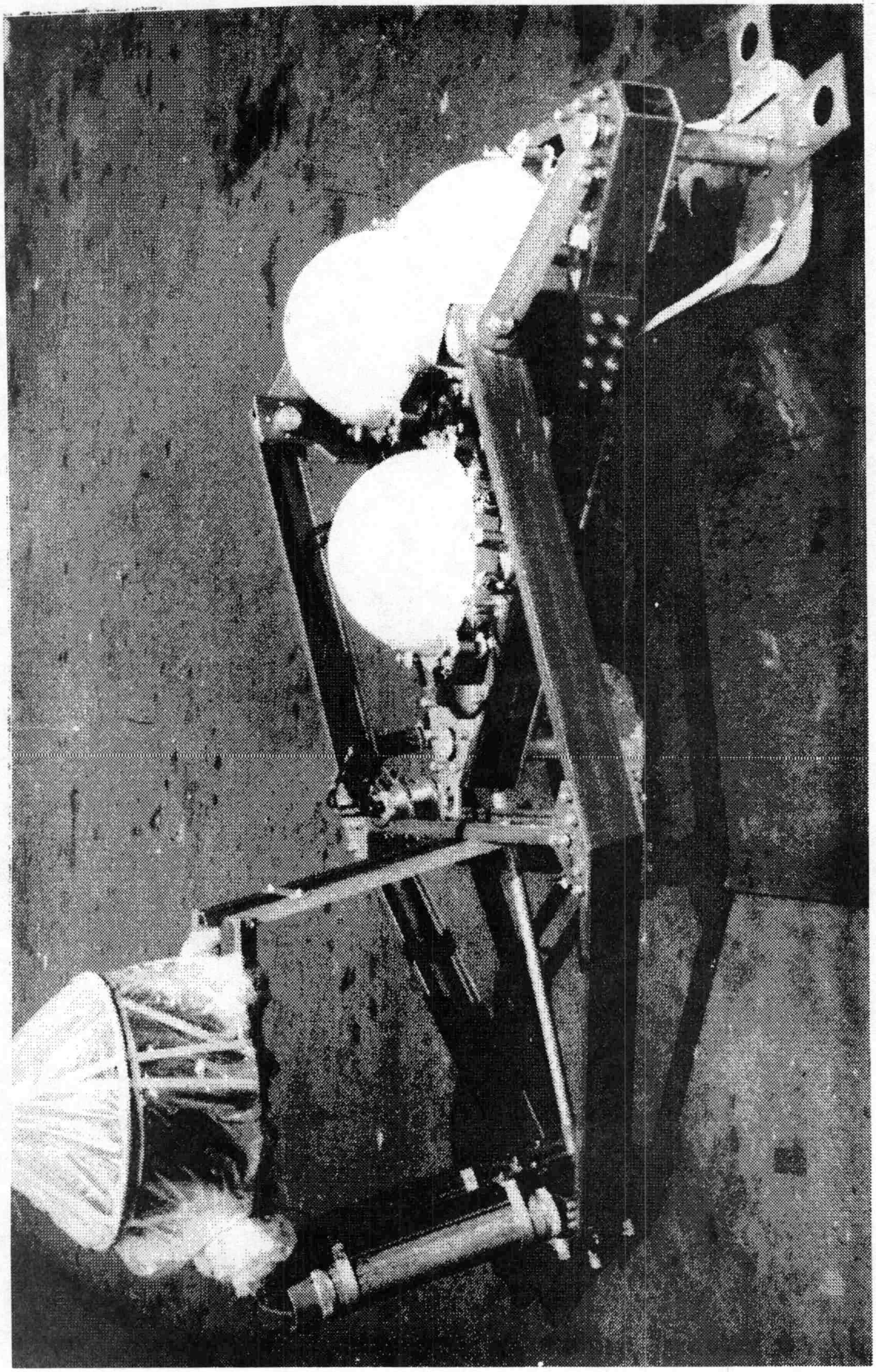
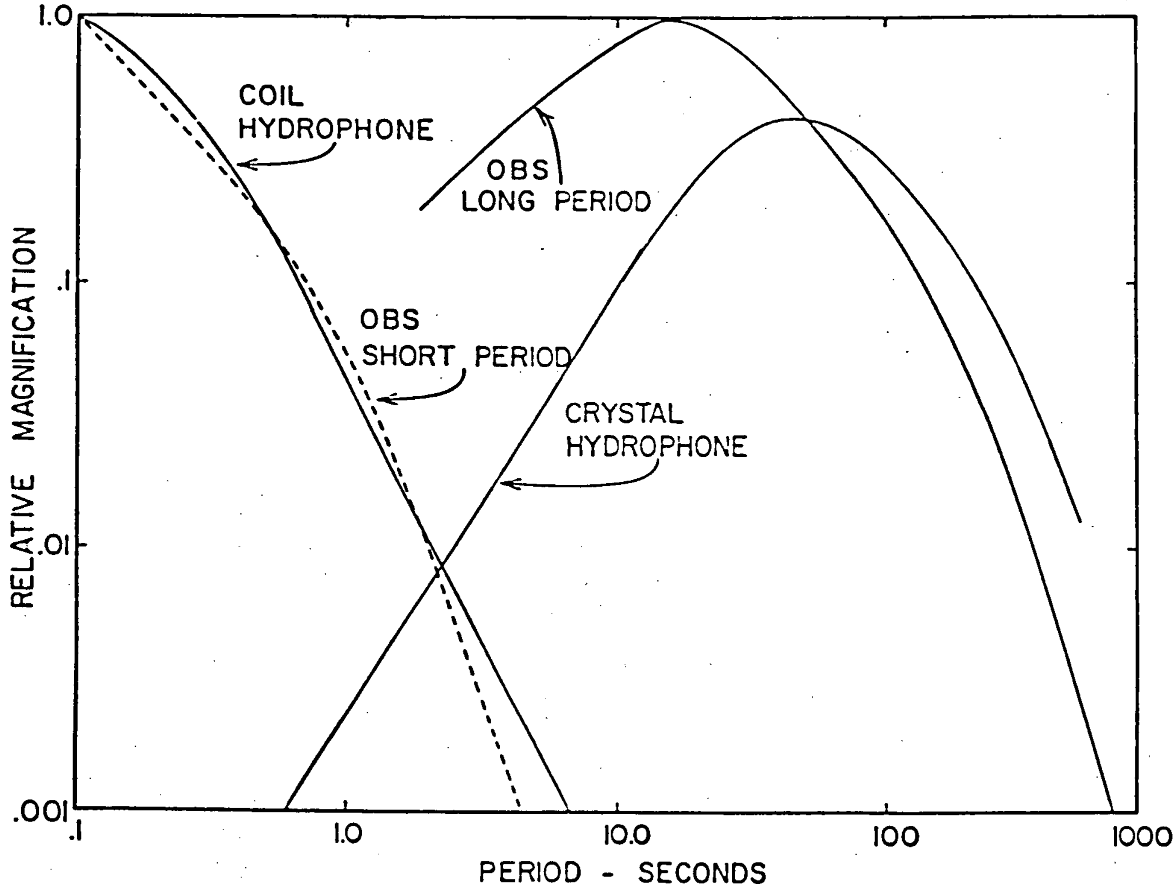


Figure 2



INSTRUMENT RESPONSE CURVES

Figure 3

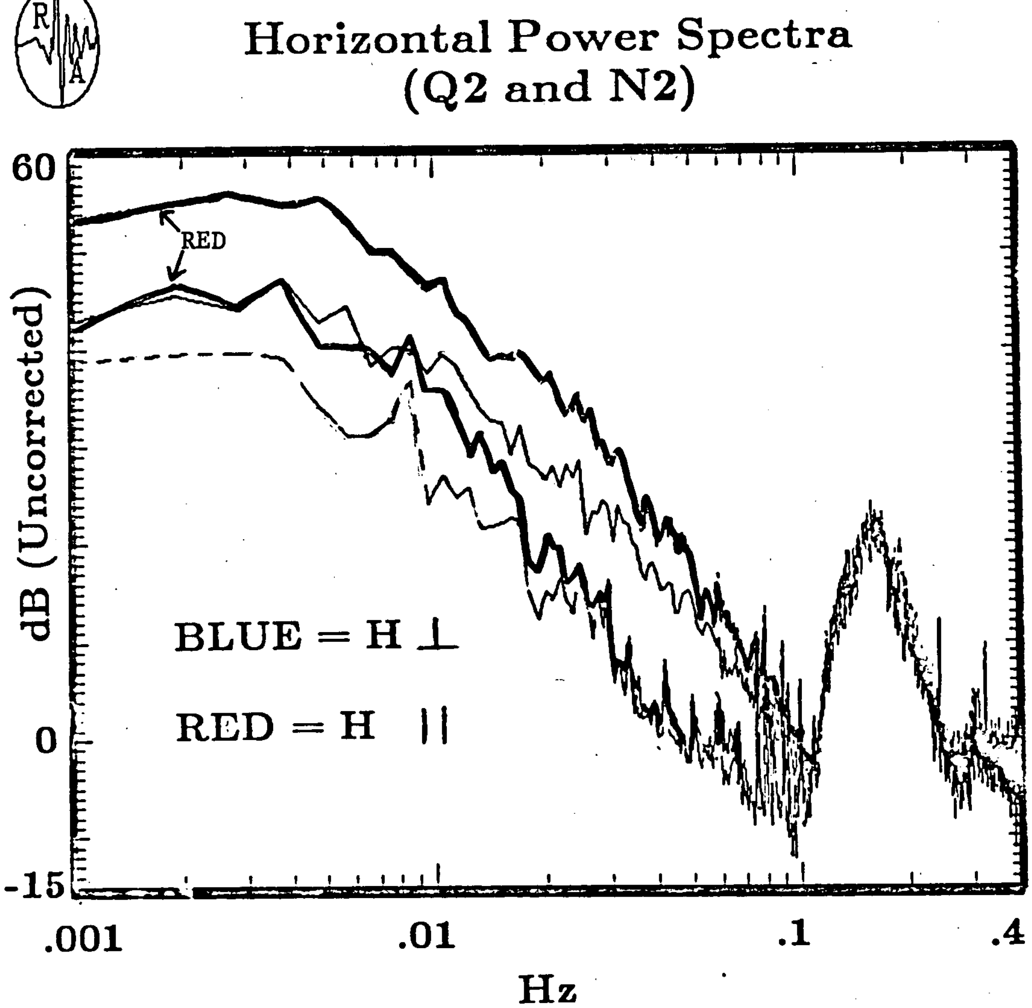
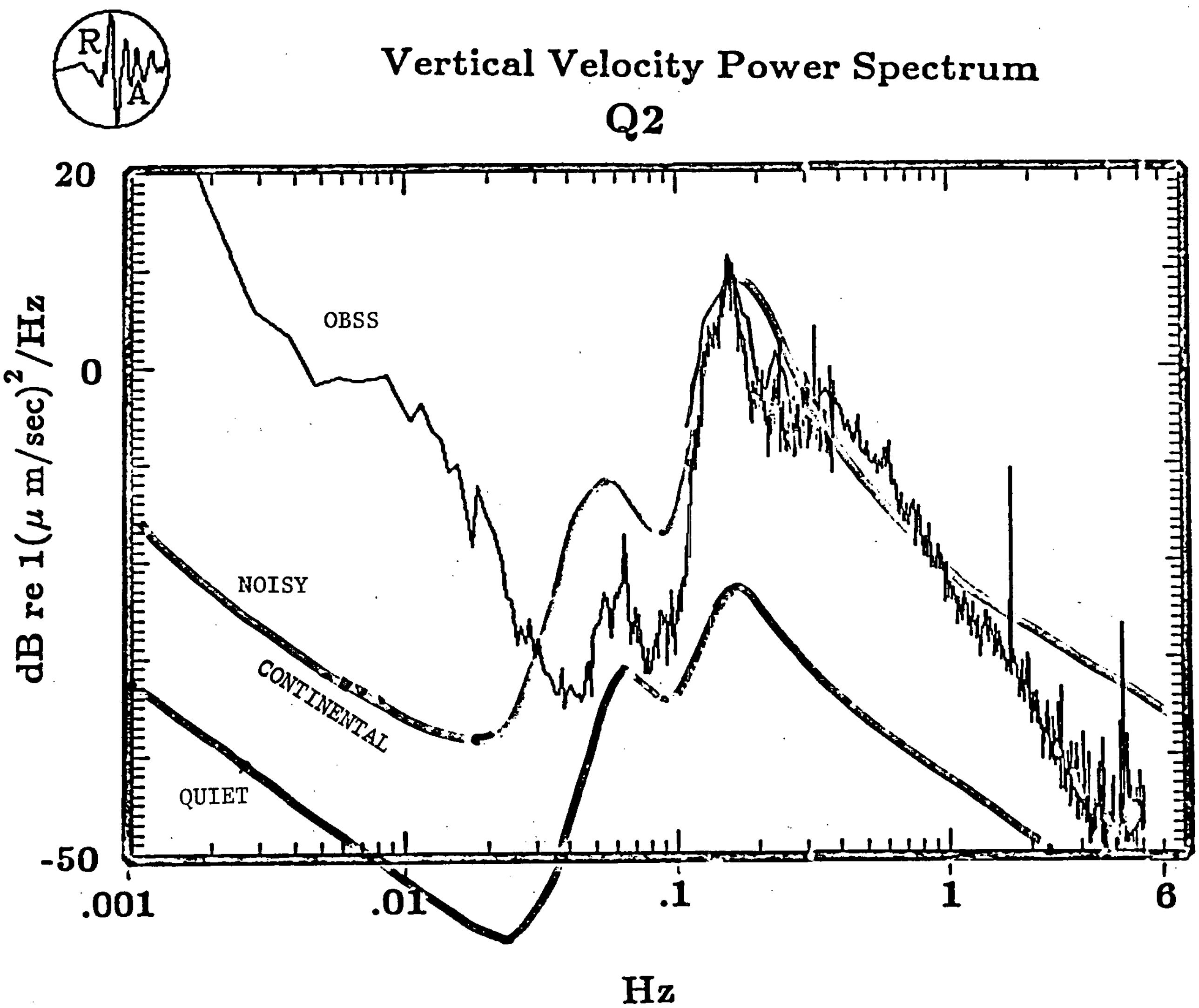

Horizontal Power Spectra
(Q2 and N2)
60
0
-15
dB (Uncorrected)
RED
BLUE = H ⊥
RED = H ||
.001
.01
.1
.4
Hz

Figure 5



R
A
Vertical Velocity Power Spectrum
Q2
20
0
-50
dB re 1(μ m/sec)²/Hz
OBSS
NOISY
CONTINENTAL
QUIET
.001
.01
.1
1
6
Hz

Figure 6

PRESSURE AND VERTICAL VELOCITY POWER SPECTRA
Q2

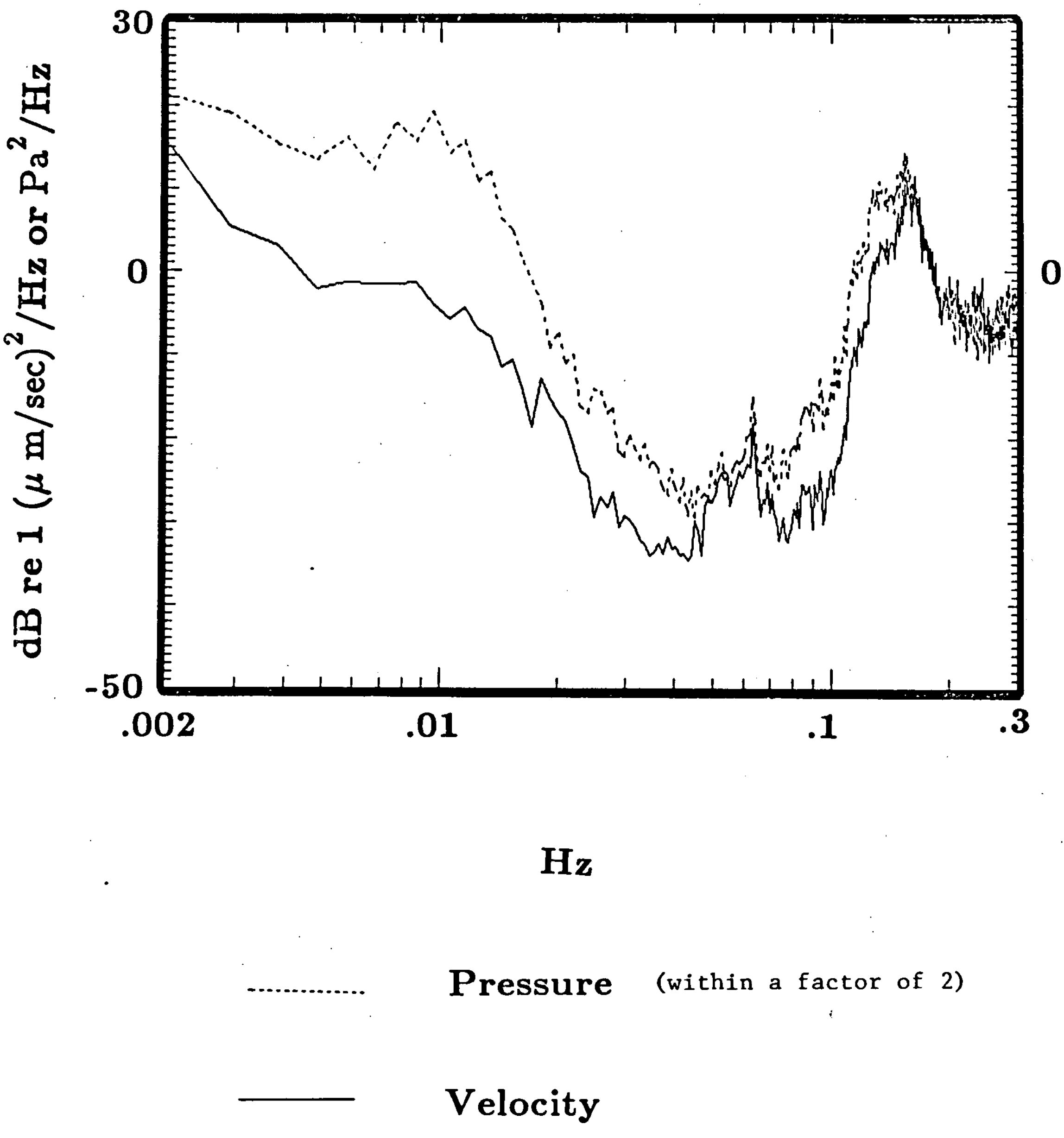


Figure 7

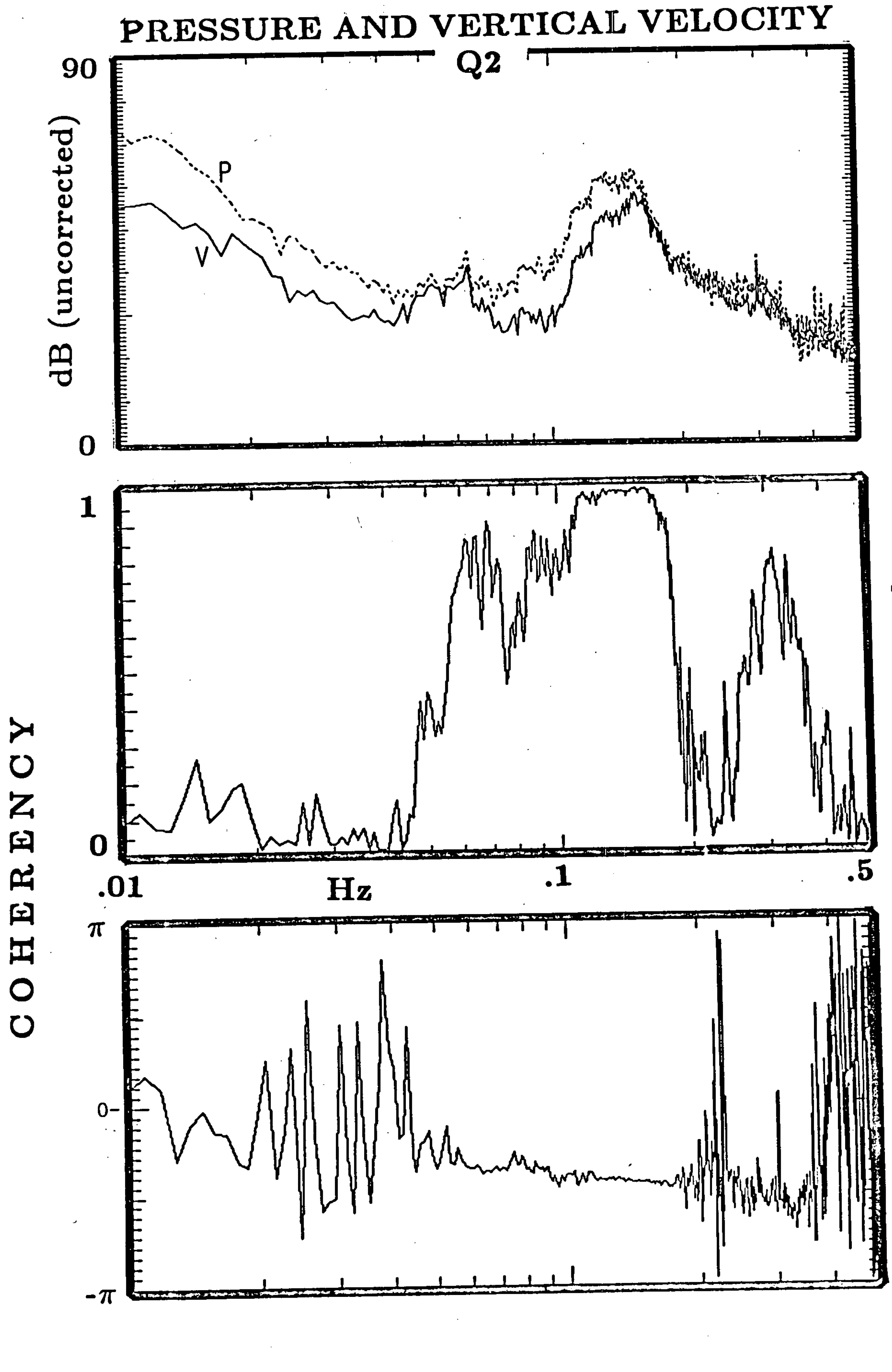

PRESSURE AND VERTICAL VELOCITY
Q2
90
0
dB (uncorrected)
P
V
1
0
Amplitude
COHERENCY
.01
Hz
.1
.5
π
0-
-π
Phase

Figure 8

PRESSURE AND VERTICAL MOTION
N1

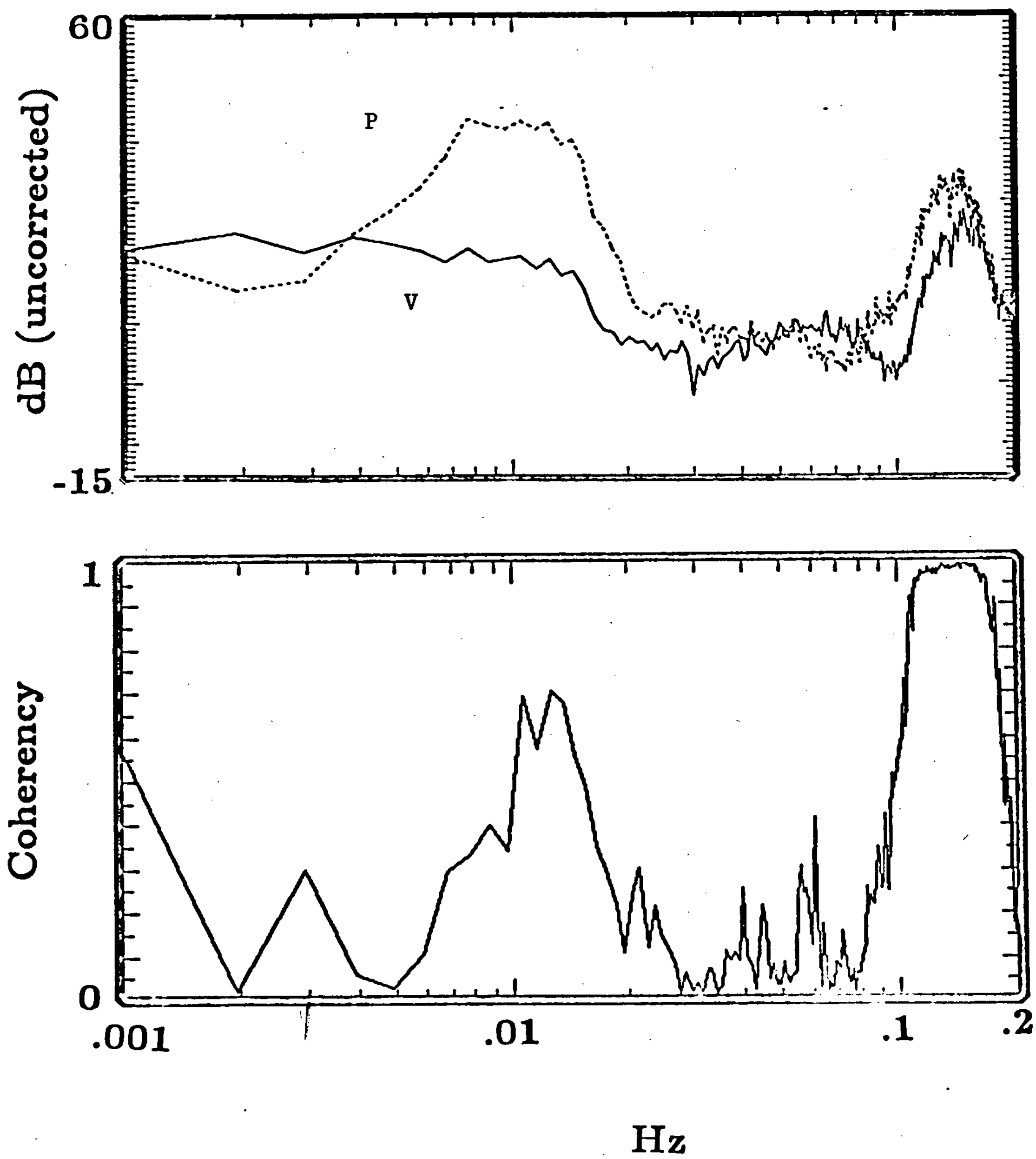
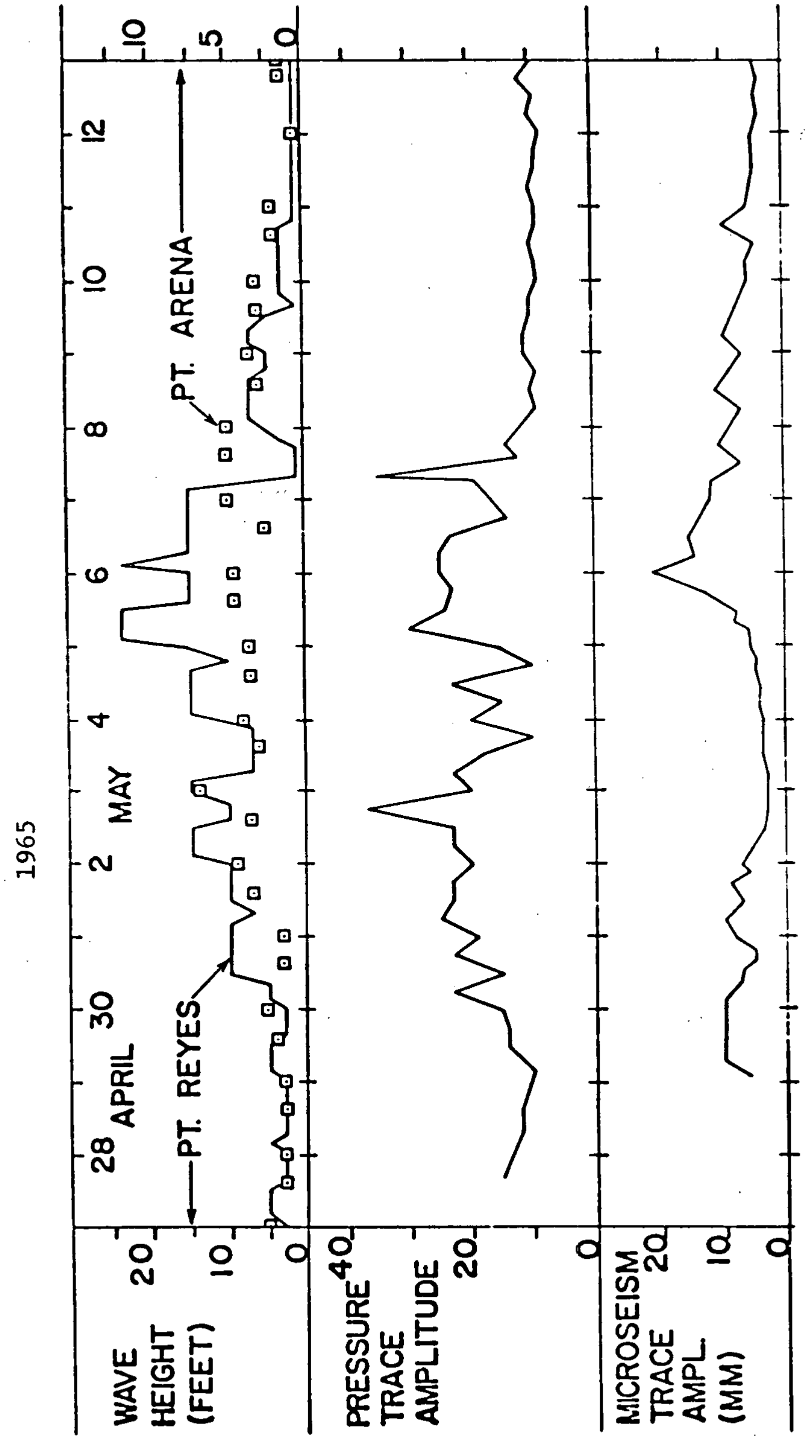


Figure 9



Comparison of local ocean wave height with amplitudes of 100 second pressure disturbance and with normal double-frequency microseisms.

Figure 10

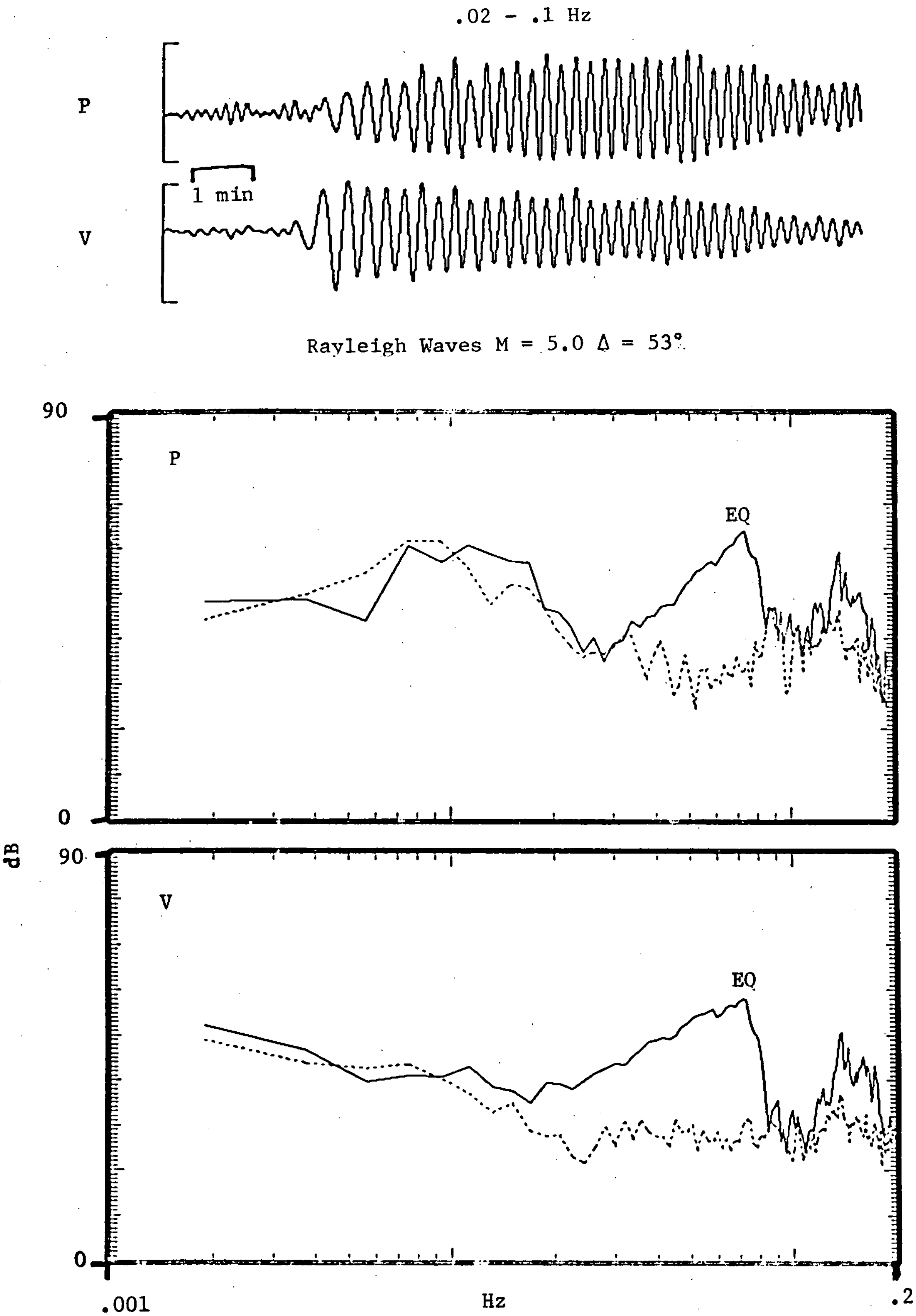

.02 - .1 Hz
P
1 min
V
Rayleigh Waves M = 5.0 Δ = 53°
90
P
EQ
0
dB
90
V
EQ
0
.001
Hz
.2

Figure 11

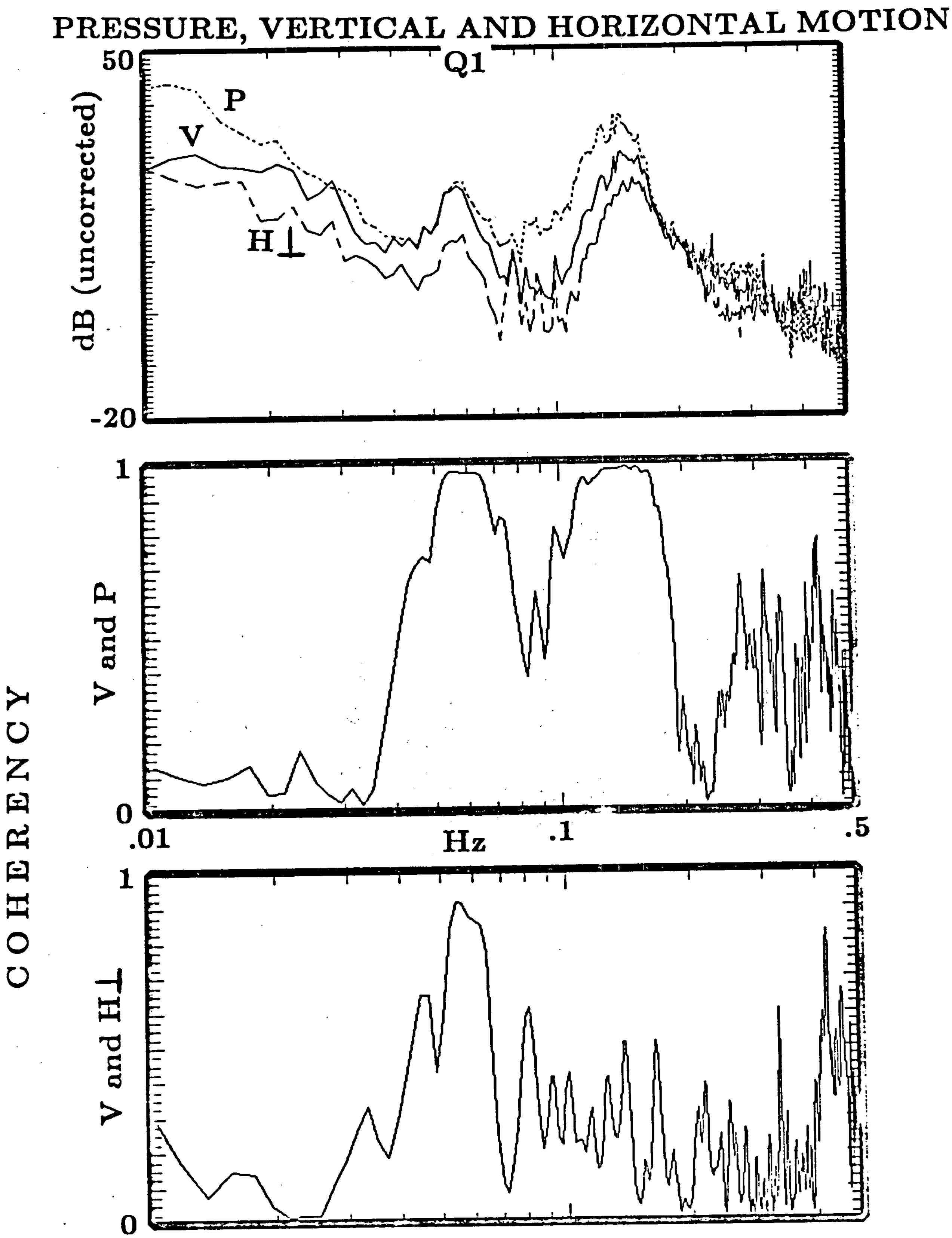
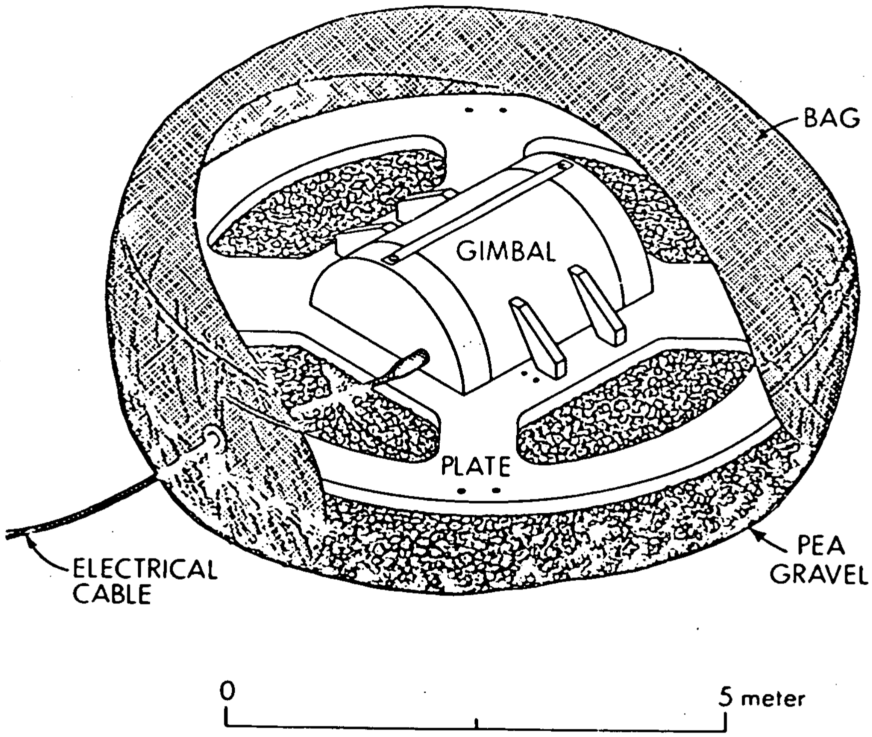

PRESSURE, VERTICAL AND HORIZONTAL MOTION
Q1
50
-20
dB (uncorrected)
P
V
H⊥
COHERENCY
V and P
1
0
.01
.1
.5
Hz
V and H⊥
1
0

Figure 12



'Opihi isolated sensor package.

Figure 13

BURIED SENSOR PACKAGE

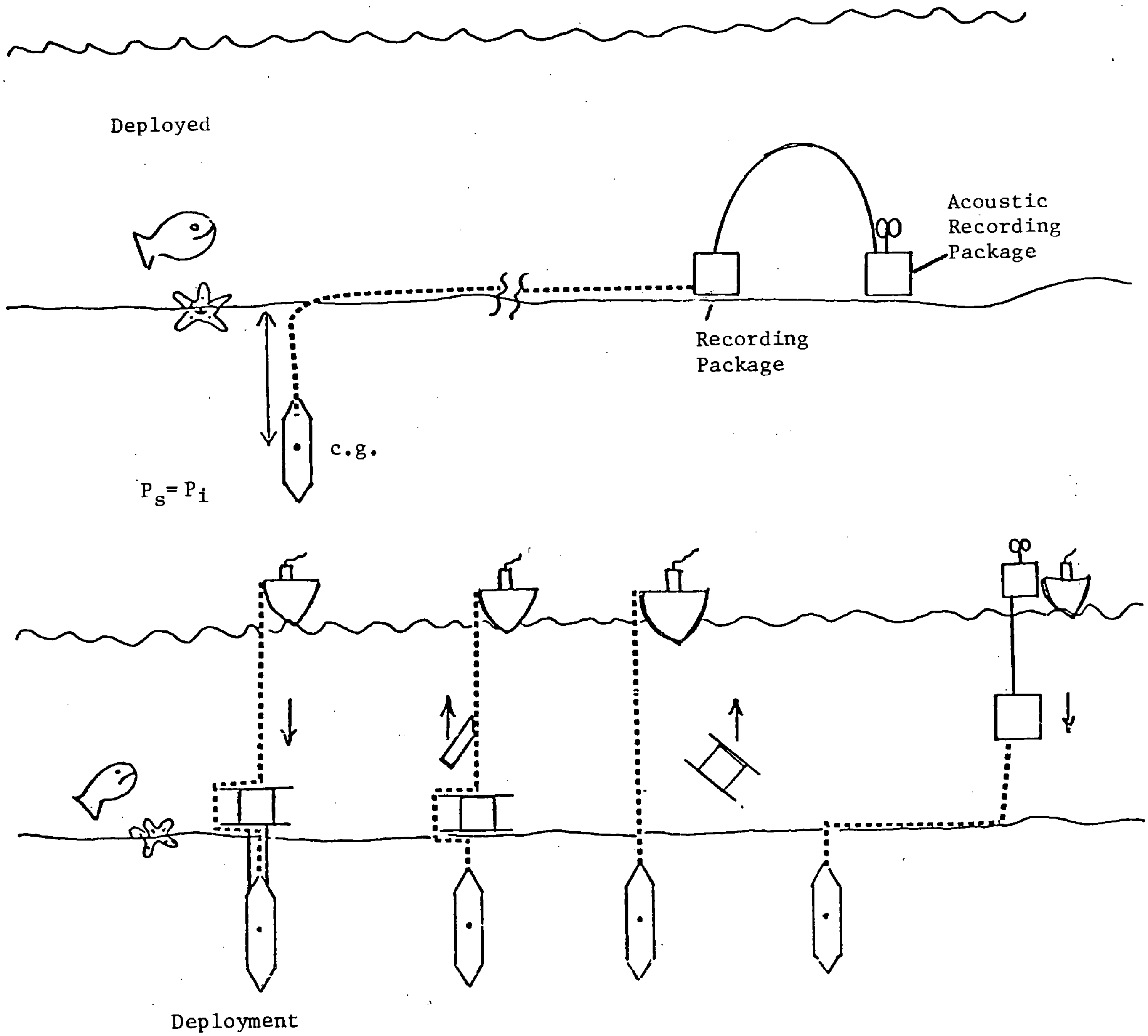


Figure 14

ON THE RELEVANCE OF SEAFLOOR ARRAY EXPERIMENTS TO BOREHOLE SEISMIC SYSTEMS

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J.A. Hildebrand, F.N. Spiess
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and (*) Naval Ocean Research and Development Activity

The motivation for emplacing seismic sensors beneath the seafloor instead of upon it is the hope that we may avoid much of the noise present at the seafloor and in the water column. Additionally, more competent materials at depth offer better coupling opportunities than do the soft sediments which cover most of the seafloor.

An important question, then, is "What is the cost/benefit ratio for emplacements of various depths?" We assert that the key to this question lies in studies of the noise field at the seafloor using arrays of sensors. Taking the mode representation of the noise wavefield, we see that the depth dependence of the noise will be determined by the vertical wavefunction of the noise. Calculating the depth dependence of the noise (or signal) be accomplished by computing the wavefunctions, establishing the excitation (or energy partition) among the various modes, and summing their contribution for various depths.

The depth dependence of any mode is closely linked to the horizontal wavefunction through the dispersion relation. In the language of body waves, the equivalent statement is that the vertical and horizontal components of the rayslowness vector must sum to the medium slowness. If we can measure the energy density as a function of horizontal wavenumber (or velocity or slowness), we can infer the vertical wavenumber of the associated wavefunction and calculate the depth dependence of the wavefield.

The challenge inherent in this procedure is the measurement of the two-dimensional wavenumber spectrum of seafloor noise. We have made significant progress toward this end. In April of 1987, we recovered data from nine instruments which had been emplaced in a small (maximum sensor spacing of 156 meters) array at the site of Hole 469 of the Deep Sea Drilling project. Figure 1 shows the location of the site and Figure 2 shows the instrument disposition.

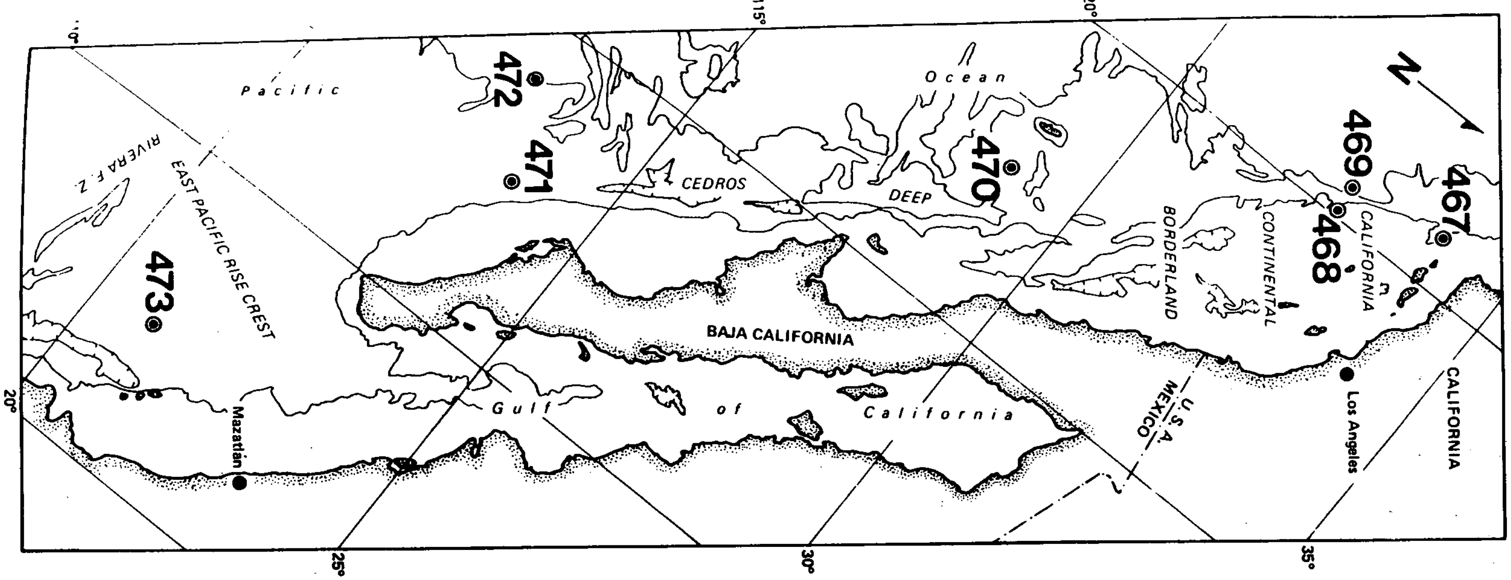
Figure 3 is a true-scale cartoon of the experimental geometry. Note well the relative dimensions of the ship, the wire, and the array itself. The time required for a wire of nearly four kilometers in length to respond to movement of the ship is about twenty minutes. We used a thruster to apply directional force to the end of the wire as a maneuvering aid. The thruster contained acoustic navigation equipment with accuracy of a meter or so. After some practice, it was possible to drop an OBS within two meters of the intended position within 45 minutes of reaching bottom with the instrument. About six hours is required to lower an instrument to this water depth.

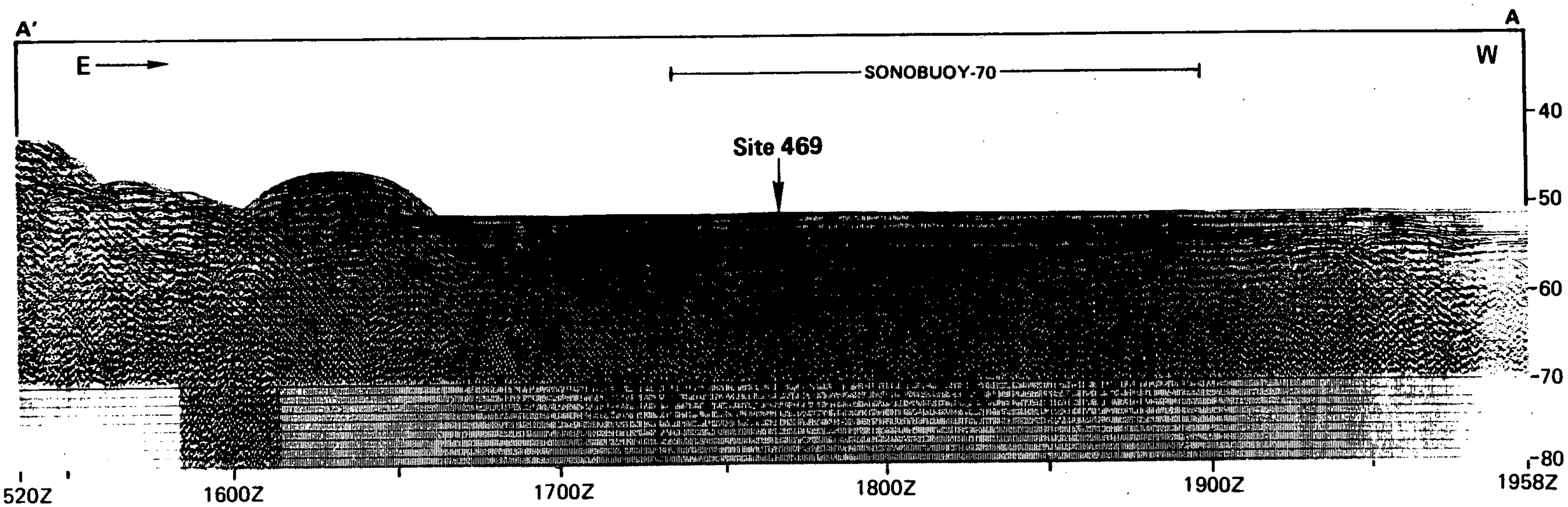
Because of the number of OBSs is limited, we adopted an array design with minimal redundance. The array produces a set of interelement spacings (the coarray) with reasonably uniform distribution. Figure 4a and 4b show the target and actual coarrays. Figure 5 shows the coarray beam pattern (the two-dimensional Fourier Transform of the coarray). The small size of the array was necessary to sample the short (down to 20 meters) wavelengths of the interface waves traveling at the seafloor. Figures 6, 7, and 8 show spectra of vertical, horizontal and pressure components. These are computed by the multiple-window method using a time-bandwidth product of four from records of 64 seconds length.

The coherence between vertical component signals as a function of frequency and separation is shown in figure 9. These coherences were calculated using ten blocks of 16 seconds length. Note that the behavior changes at about 0.4 Hz. Below this frequency there is little decay of coherence with increasing distance, while above 0.4 Hz a ridge of high coherence centered at about 3 Hz forms. We have previously observed that the 1-5 Hz band is the dominant frequency of propagation of the seafloor interface waves (Scholte or Stoneley modes). Figure 10 shows a dispersed wavetrain from a small seafloor explosion and the multiple filtering decomposition of that wavetrain. Figure 11 shows the velocity structure of the seafloor inferred from the analysis of the dispersion.

In Figure 12, we see some vertical wavefunctions for a simple model. We show two wave types: Rayleigh waves (bound to the earth's surface), and the Scholte waves (bound to the seafloor) mentioned earlier. The displacements are normalized to unity at the seafloor. Note that the Scholte wave amplitudes decay much more rapidly with distance away from the seafloor.

Our initial conclusion is that both high-velocity and low-velocity noise are present at the seafloor. The low-velocity noise will decay much more quickly with depth than will high-velocity noise.





Multichannel seismic-reflection profile AA′ showing proposed position of Site 469 (*S. P. Lee*). (The location of this profile is shown on Fig. 1.)

Figure 1

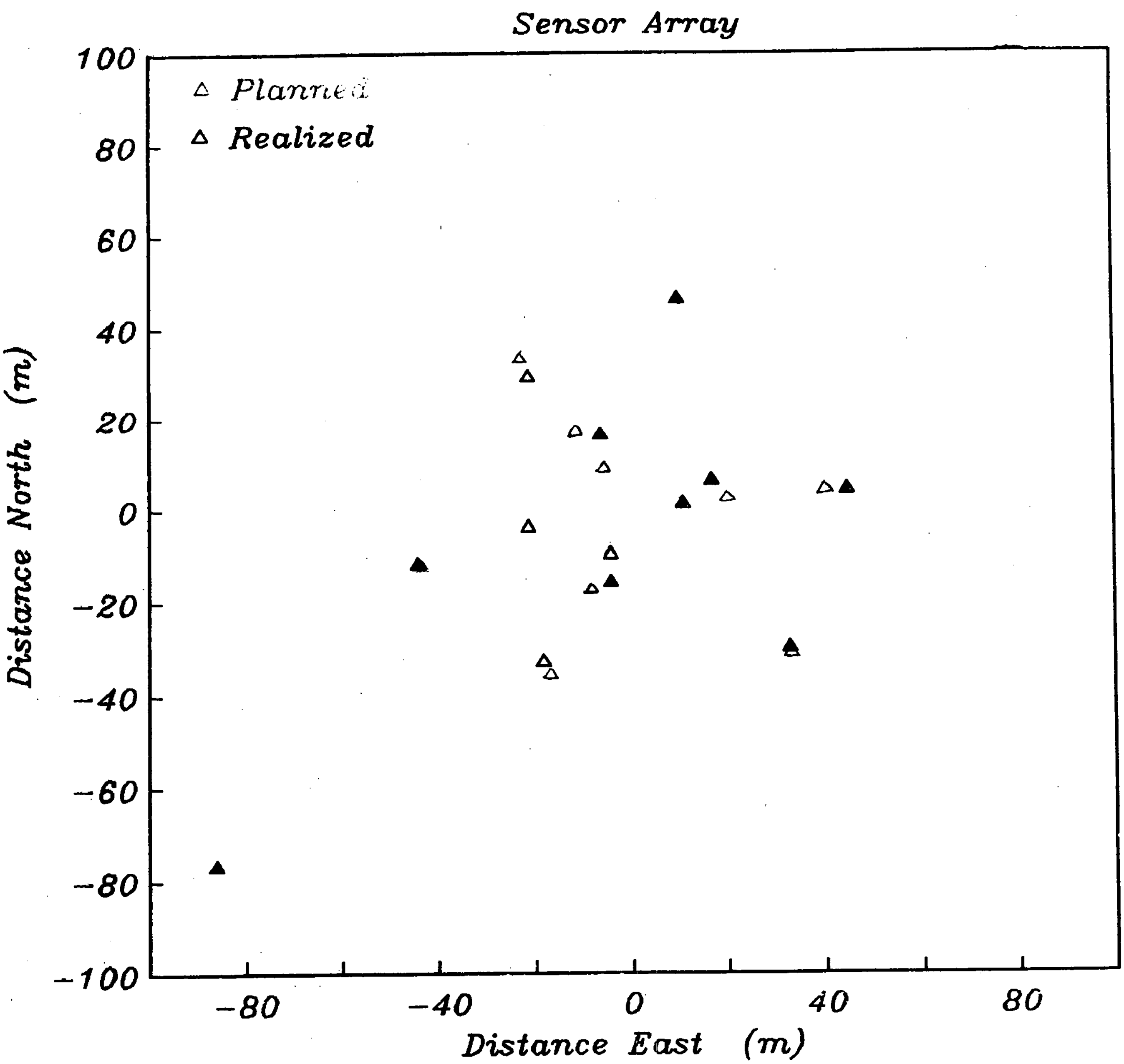


Figure 2. Geometry of the ocean bottom seismometer (OBS) array.
The water depth is 3800 meters.
The red positions are the planned sensor locations.
The black positions are the locations actually achieved.
The solid black triangles correspond to instruments that returned data.

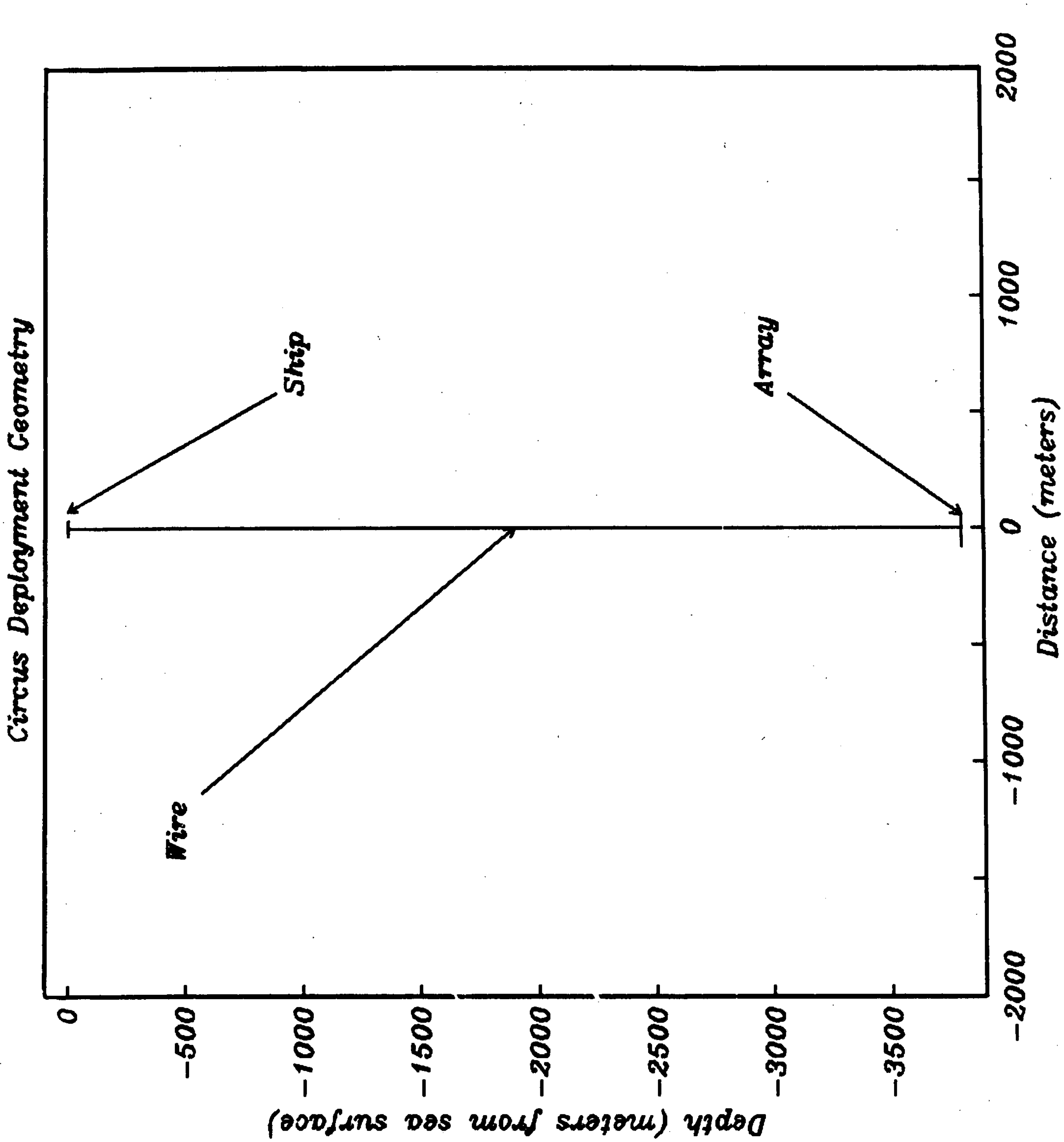


Figure 3. A true-scale cartoon of the experimental geometry.

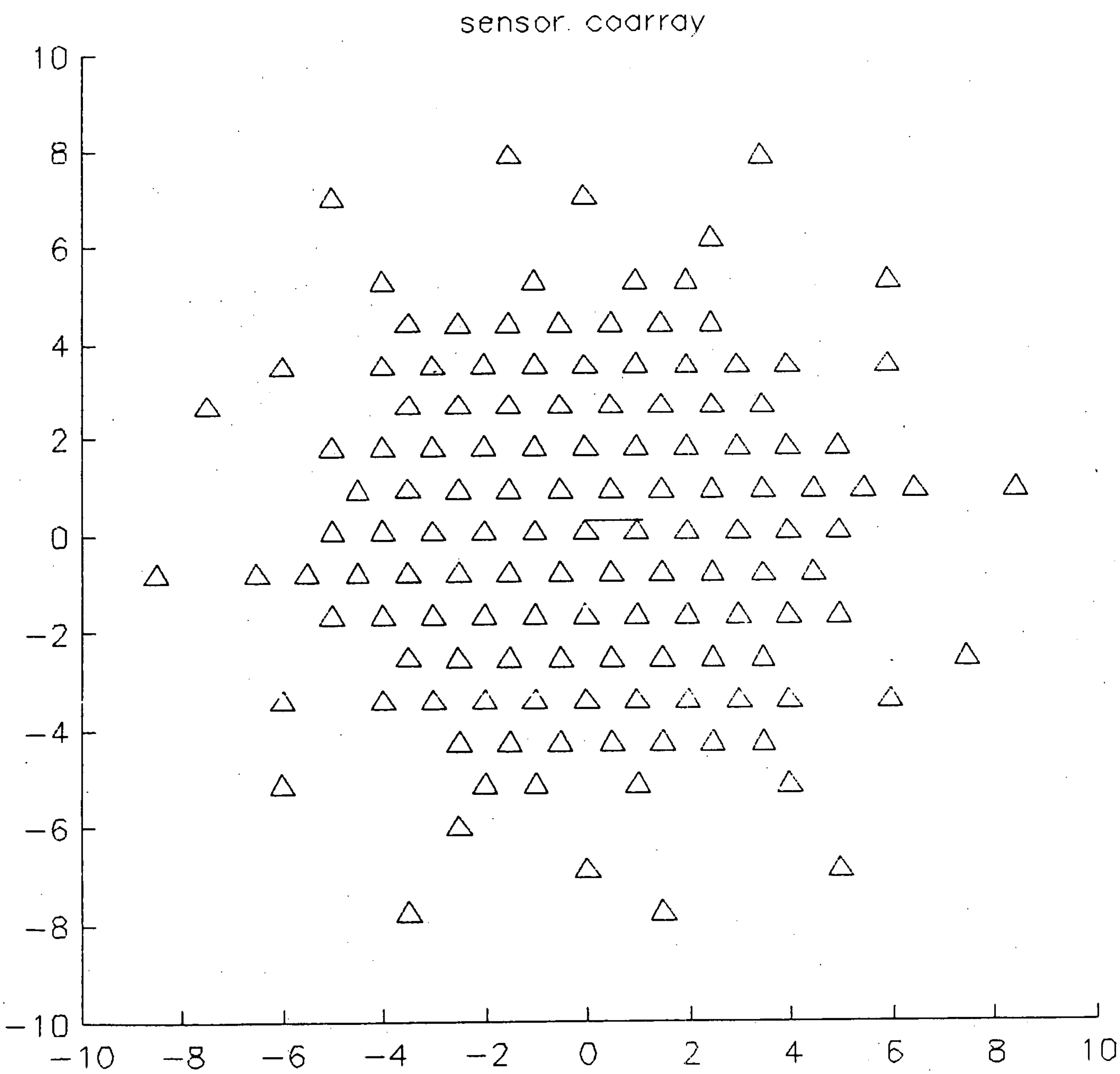


Figure 4a. The sensor coarray (the convolution of the array itself.
(a, proposed; b, realized)

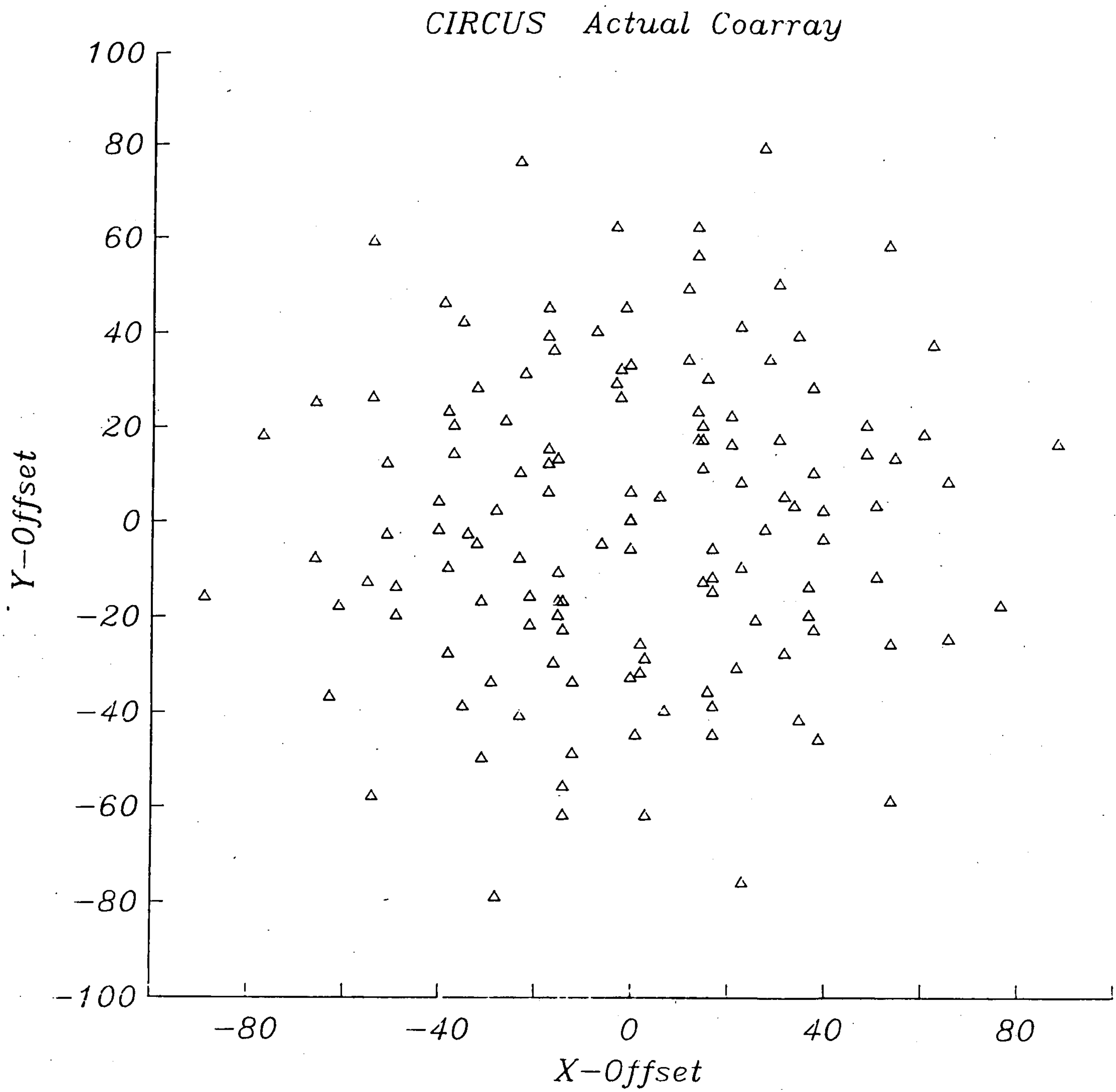

CIRCUS Actual Coarray
100
80
60
40
20
0
-20
-40
-60
-80
-100
Y-Offset
-80
-40
0
40
80
X-Offset

Figure 4b

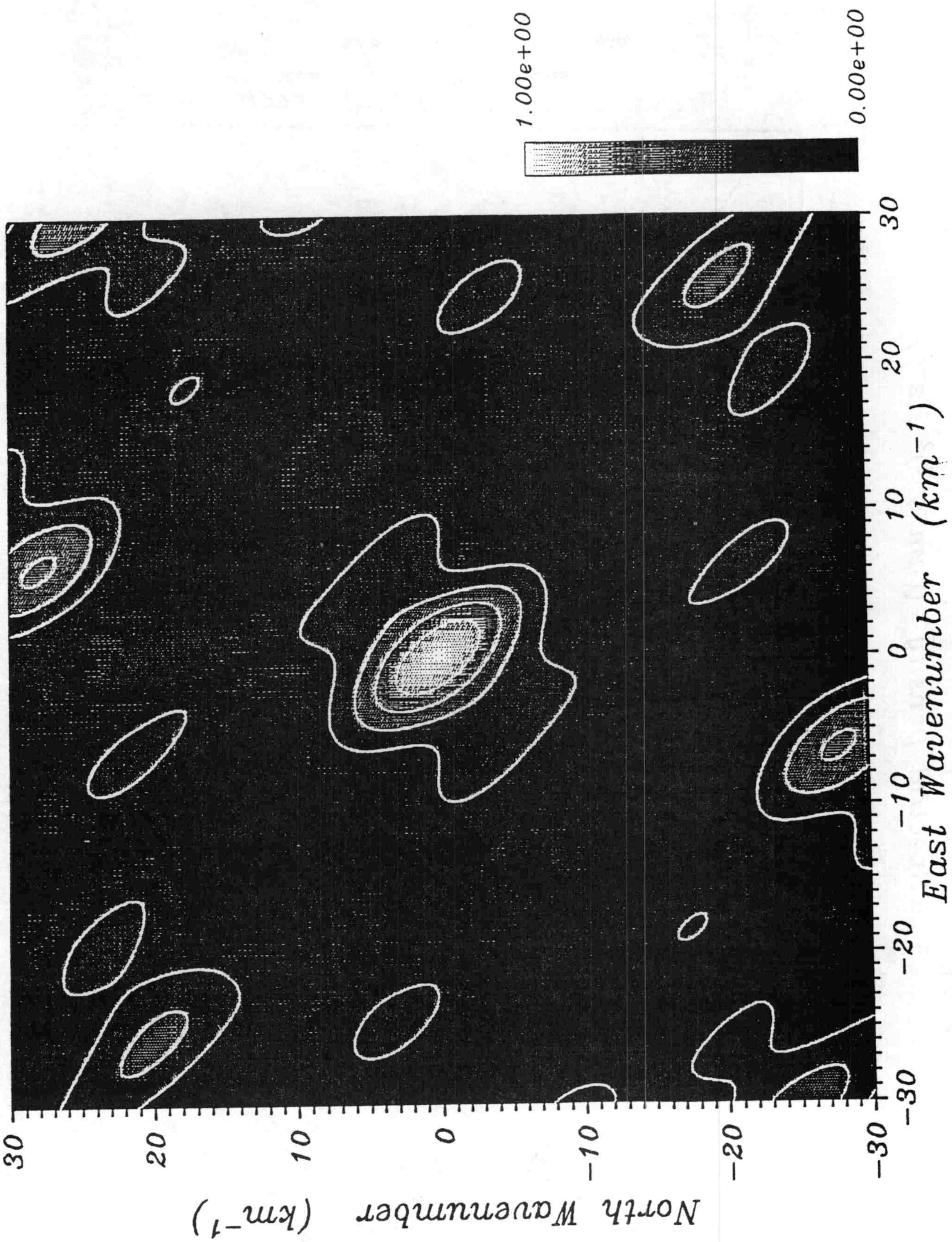


Figure 5. The beam pattern (the 2-dimensional fourier transform of the coarray).

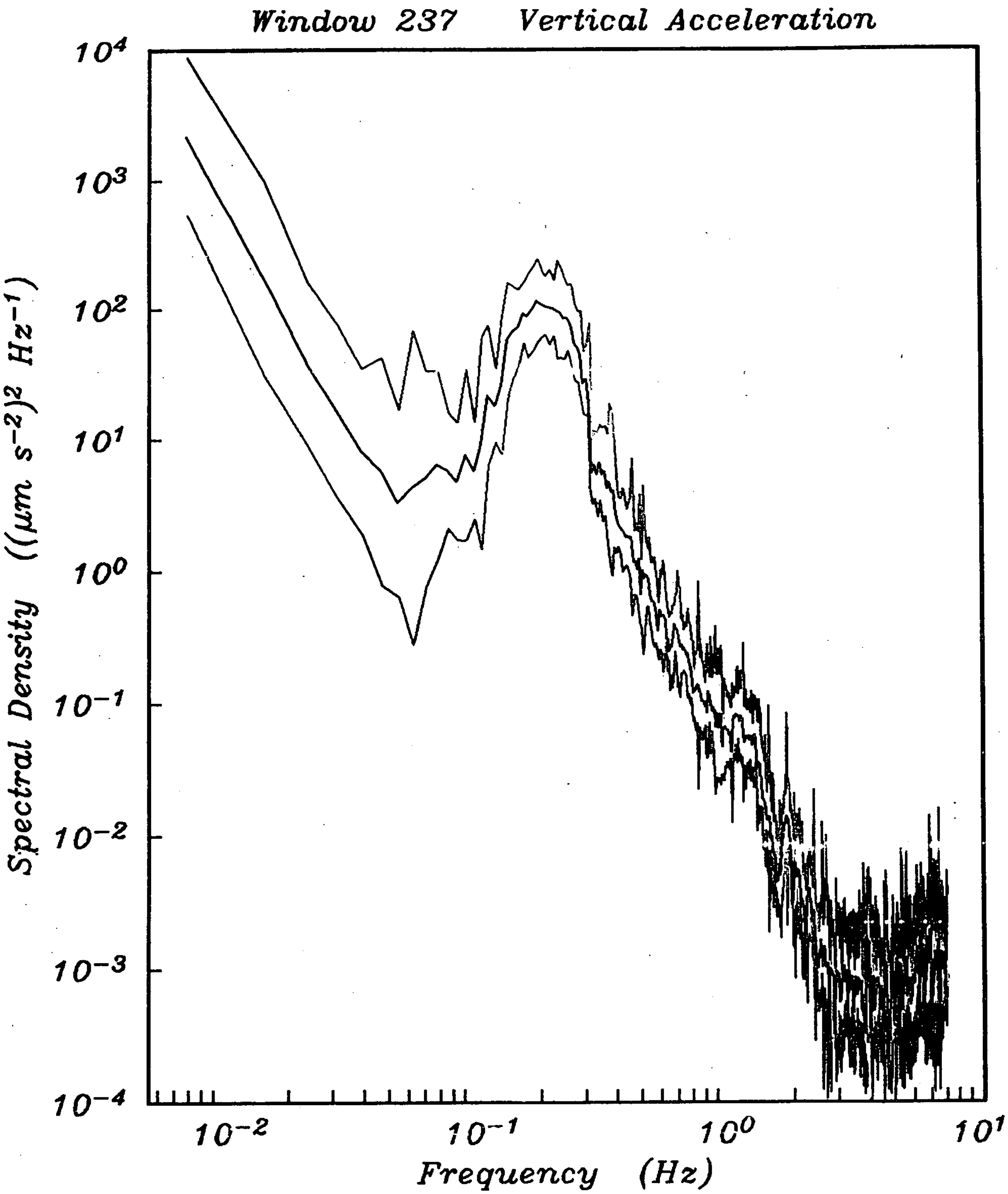


Figure 6. These power spectra were computed using the multiple prolate shperoidal window method. In each case 8192 points (64 seconds) were used. The time-bandwidth was 4 and the first 8 windows were averaged. Confidence intervals are 95% and were computed by the 'jackknife" method.

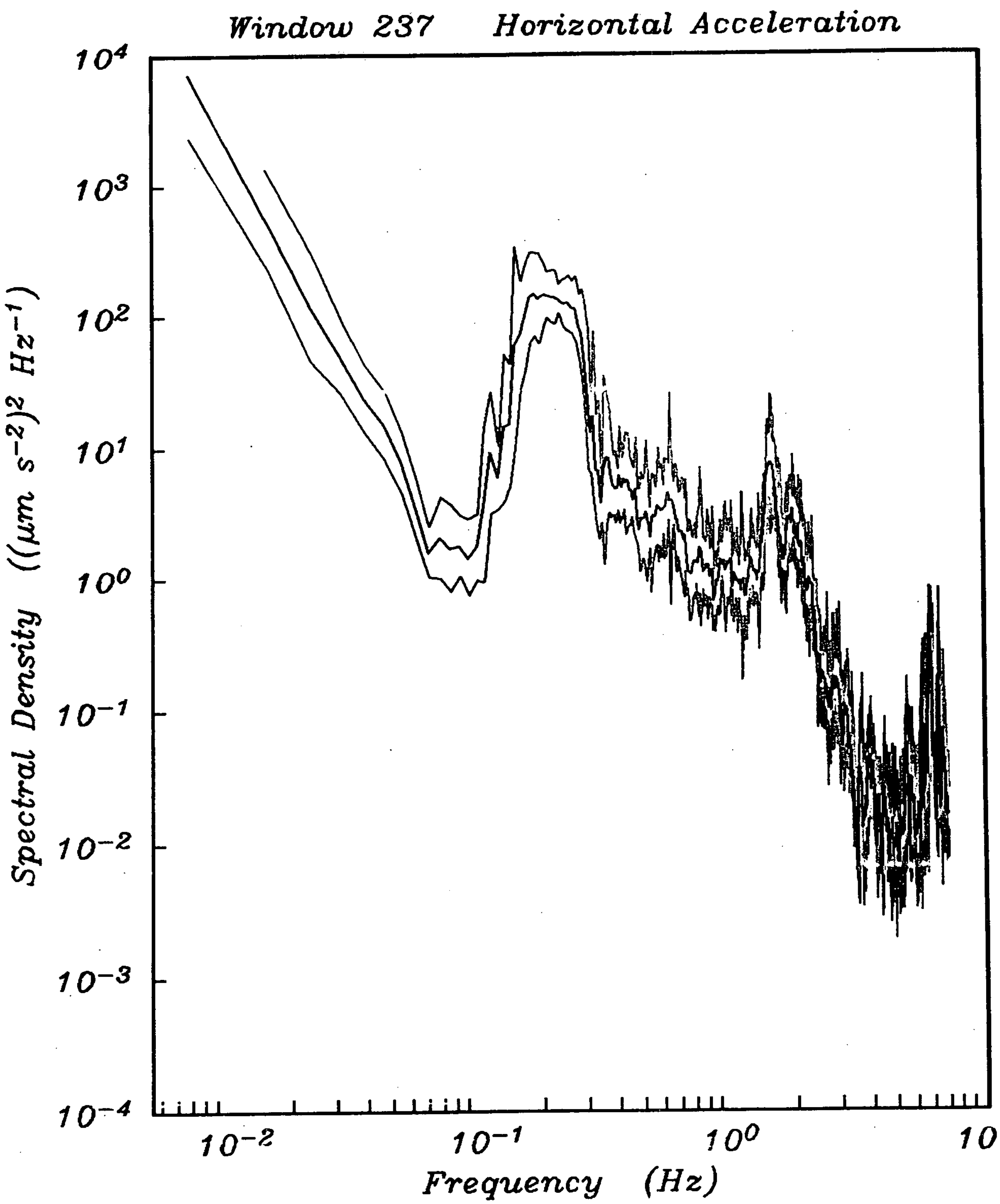


Figure 7. These power spectra were computed using the multiple prolate shperoidal window method. In each case 8192 points (64 seconds) were used. The time-bandwidth was 4 and the first 8 window functions were averaged. Confidence intervals are 95% and were computed by the "jackknife" method.

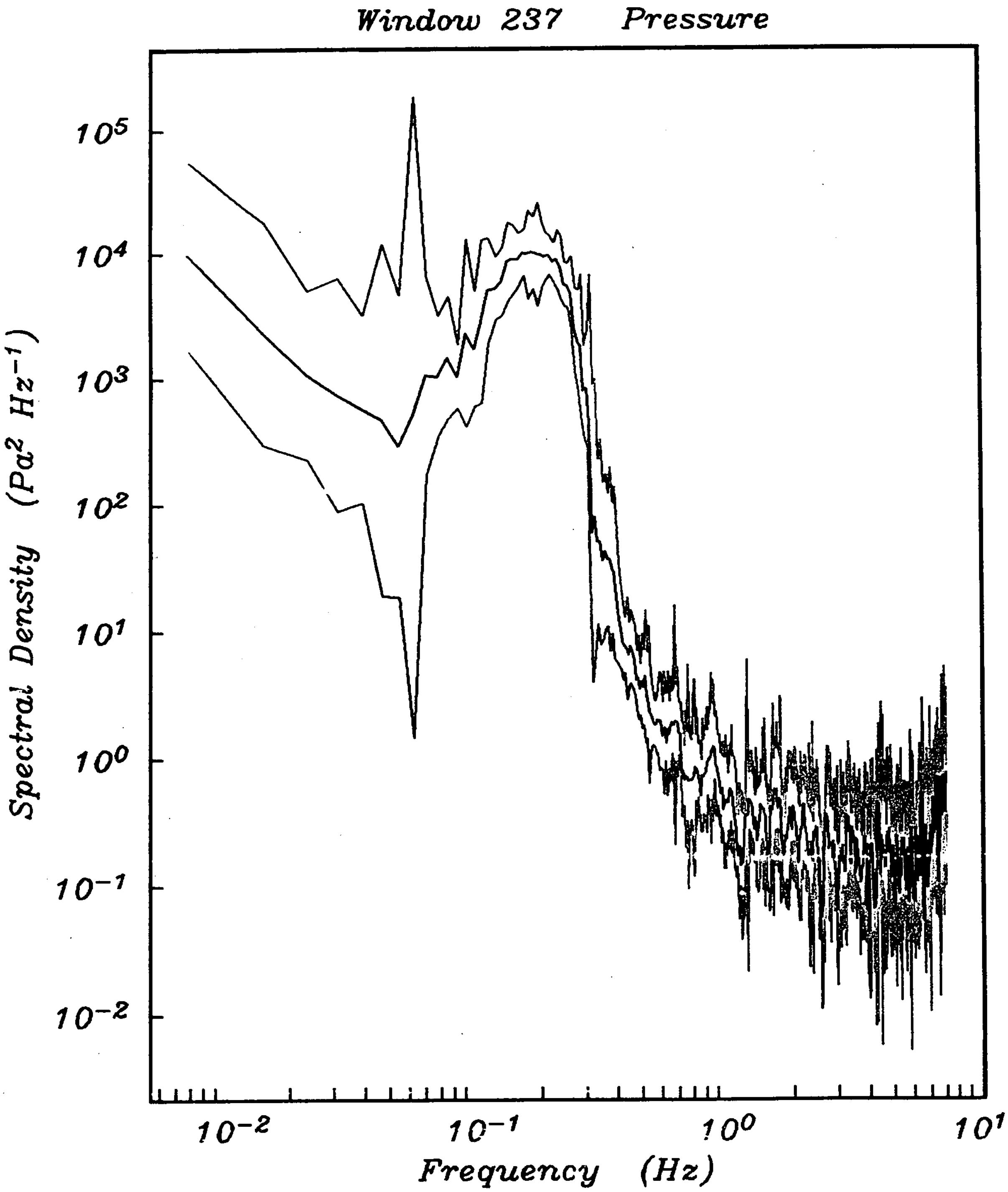


Figure 8. These power spectra were computed using the multiple prolate shperoidal window method. In each case 8192 points (64 seconds) were used. The time-bandwidth was 4 and the first 8 window functions were averaged. Confidence intervals are 95% and were computed by the "jackknife" method.

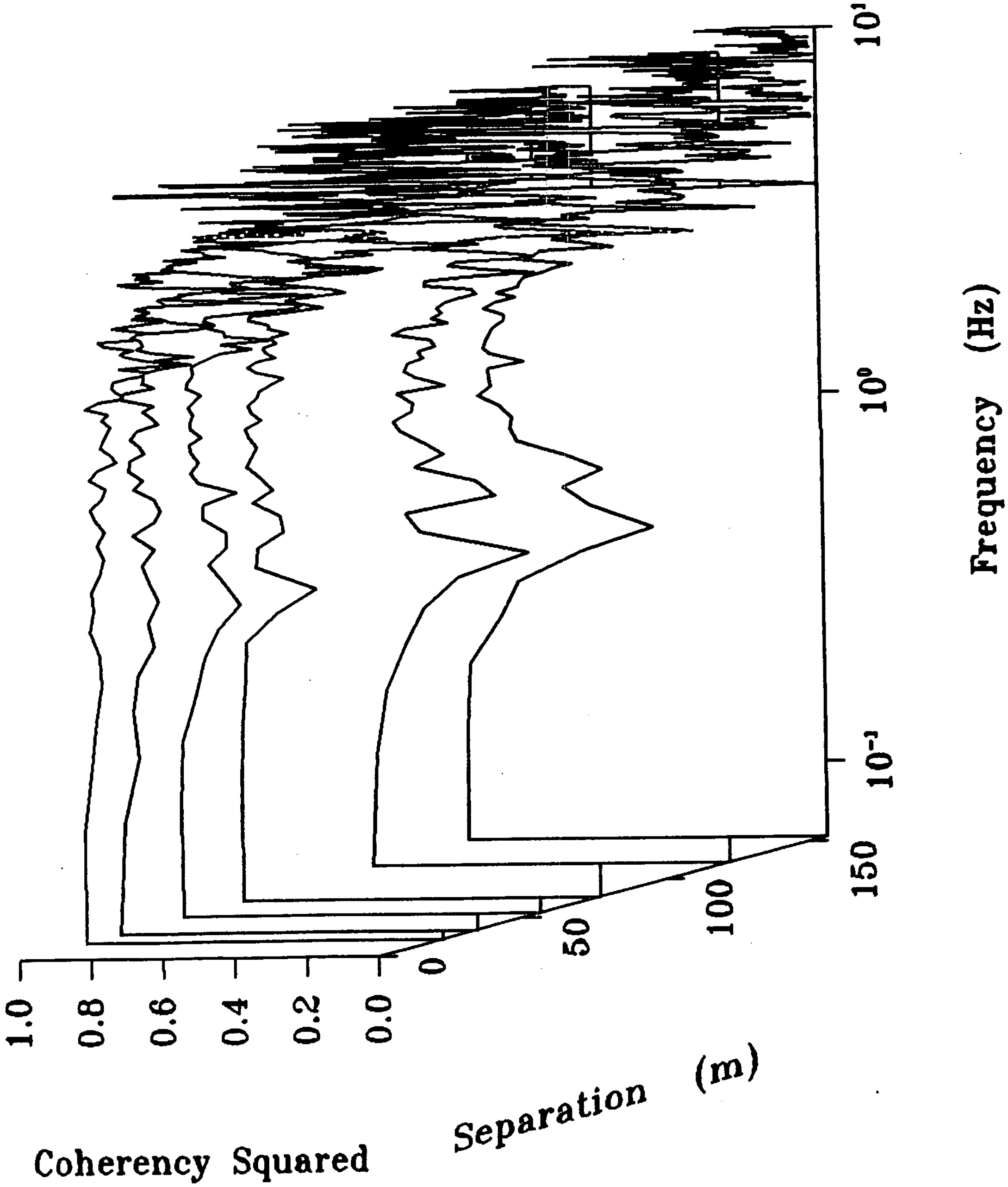


Figure 9. Coherence as a function of frequency and sensor separation.

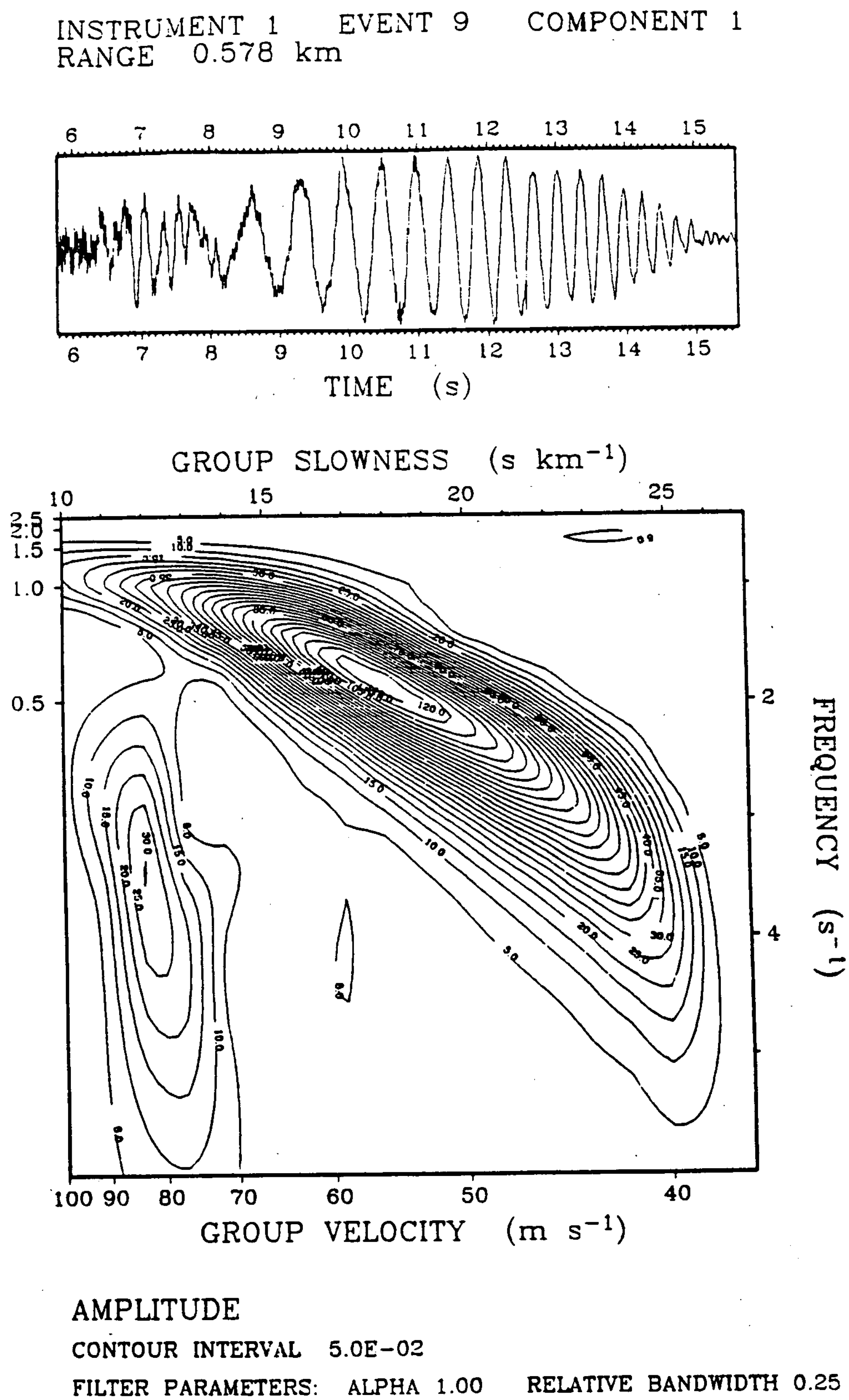


Figure 10. Time-frequency filter analysis of dispersed Stoneley wave train. The two ridges represent the fundamental (Mode 0) and second overtone (Mode 2).

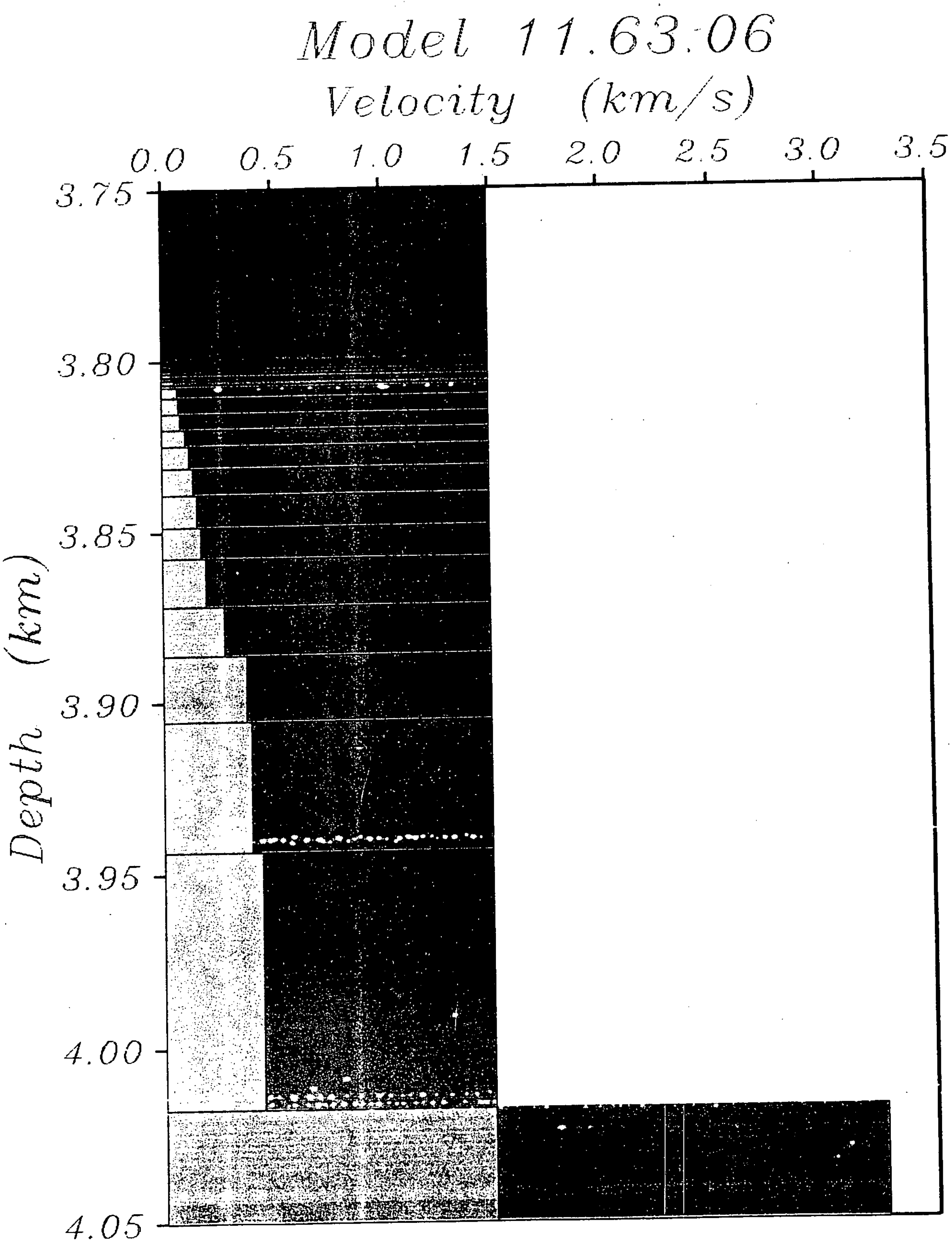
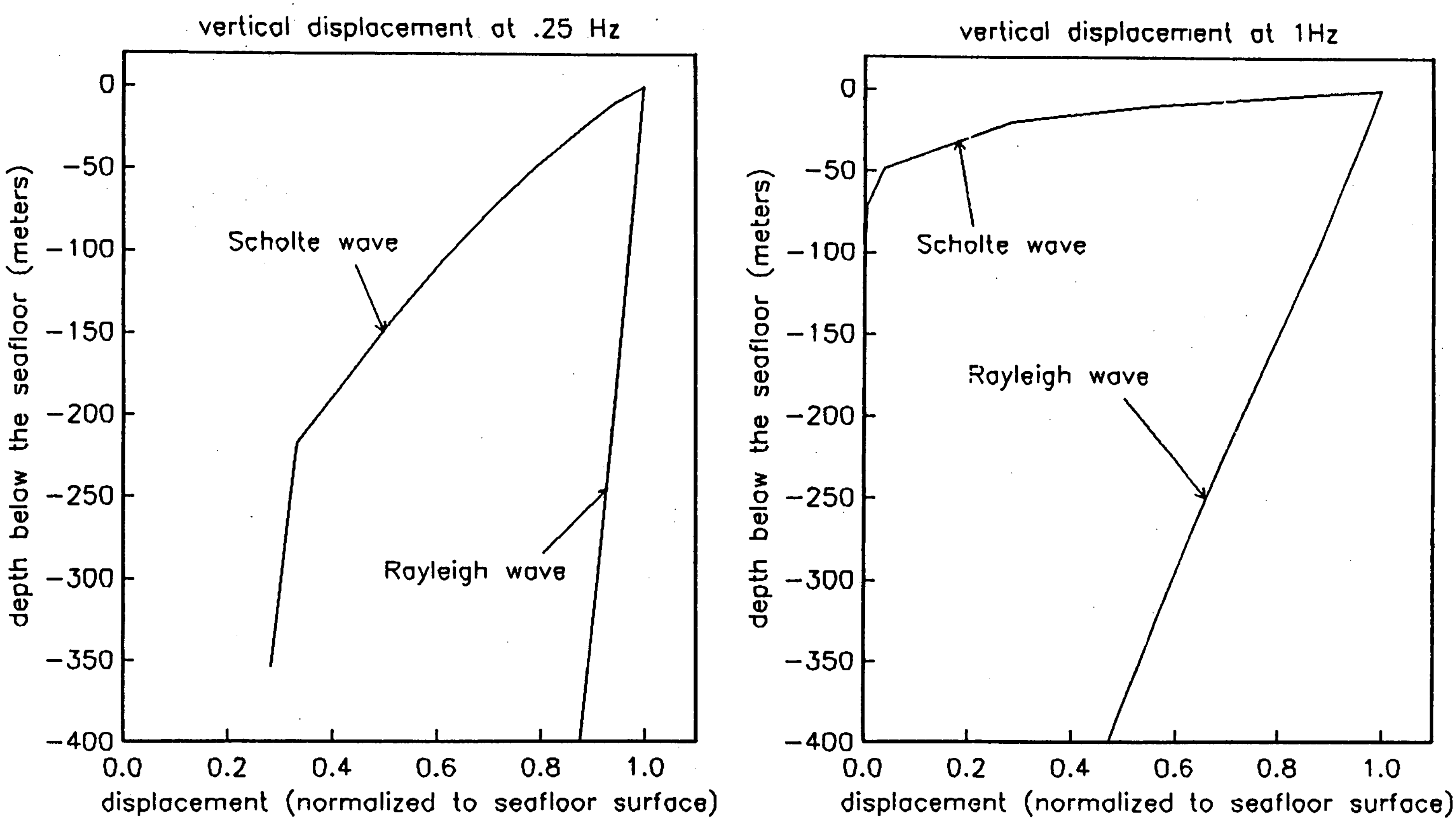


Figure 11. The Compressional (green) and shear (red) velocities at DSDP Site 469.

Figure 12. Wave function for Rayleigh and Scholte modes at two frequencies.



EFFECTS OF OBS BURIAL ON GROUND COUPLING AND S/N RATIO ENHANCEMENT

T. Yamamoto, A. Turgut, M. Trevorrow, D. Goodman, and M. Badiey
Rosenstiel School of Marine and Atmospheric Sciences
University of Miami
Miami, FL 33149

Abstract

A new OBS system has been developed to bury the OBS seismometer packages ~0.5 m below the seafloor. This system has been tested successfully in 1987 New Jersey Atlantic Generating Station Site for the burial of a 7-element ULF/VLF (.005-100 Hz) OBS array. In this paper, the effects of OBS burial on ground coupling and signal-to-noise ratio enhancement are examined by comparing buried and unburied OBSs at the same site. The comparisons show that burial of OBSs improves signal-to-noise ratio drastically, especially in the horizontal components.

A seven-element shallow water (0 to 200 m) OBS array has been designed to measure ULF/VLF seismic signals and pressure fluctuations on the continental shelves. Since one of the applications of the system is to determine shear modulus profiles of the marine sediments, it is named the Bottom Shear Profiler (BSMP). Each BSMP may be buried in the seabed by a newly developed hydraulic jet burial system (Figure 1). The array electronics are set up to make real-time measurements, which enable signal monitoring and processing on-site. Also, it is capable of determining directionally dependent spectral characteristics of gravity waves and seismic signals. The spacing between each element can be extended up to 200 m and the dimensions can be extended up to 800x400m (Figure 2). A single BSMP unit is capable of measuring acceleration of 10^{-6} m/s^2 at 10 mHz and pressure fluctuations of 10 pascal in the frequency range of between 5 mHz and 100 Hz. Each unit of the array contains: three orthogonally mounted medium-period (Teledyne Geotech, Model S-750) seismometers, a differential pressure sensor, two tiltmeters, and a compass. In principal, the Lopez Island Neutral Density Burial Seismometer is chosen to be a prototype of the BSMP (Sutton and Duennebier, 1987). A complete information on the instrumentation of the BSMP array is given in Turgut et al (1987).

In this paper, we present some results of the 1987 experiment to examine coupling and signal fidelity of a buried BSMP unit. In Figure 3a and 3b, measured acceleration and pressure power spectra from a buried (BSMP-301) and an unburied (BSMP-302) unit at Atlantic Generating Station site are given. Although the energy level of a horizontal seismometer signal at periods between 1 and 3 seconds are in the same order, the acceleration power level of a buried BSMP is 20 dB lower for the larger periods. This difference seems to be minor for vertical seismometer signals. A complete data analysis of 1987 New Jersey experiment is given Trevorrow et al (1988).

Because the S/N ratio enhancement is achieved by using a burial BSMP unit, Goodman et al (1988) were able to show that particle motion at 1-3 second is retrograde and apparent velocities of seismic waves are between 200 and 300m/s. This suggests that the waves at these periods are Sholte waves. Also they showed that for larger periods the particle motion is prograde and mostly gravity wave coupled. Using the measured seabed motion and seabed displacement in the gravity wave coupled frequency band, depth dependent shear modulus can be inverted. Inversion is based on the Singular Value

Decomposition (SVD) technique. In Fig. 4 a rather deep inversion (200m) for AMCOR-6010 site is shown.

In conclusion, noise level of a conventional OBS can be reduced drastically especially for the horizontal seismometers just by burying the seismometer package right under the seafloor. Of course, one would like to deploy the seismometer package in a borehole reaching the rock basement for any event recording. However, there is a trade off between cost and signal quality. We think that coupling and signal quality of OBSs can be improved by burying the seismometer package which requires very small amount of additional cost.

References

Goodman, D., et al., Directional Spectra Observations of Seafloor Microseisms from an Ocean Bottom Seismometer Array, submitted to JGR.

Sutton, G.H., and Duennebier, F.K., Optimum Design of Ocean Bottom Seismometers, Marine Geophysical Researches, v. 9, p. 47-65, 1987.

Trevorrow, M., et al., High Resolution Bottom Shear Modulus Profiler Experiments on the New Jersey Shelf, Summer 1987, GAL Report 1006, University of Miami, 1988.

Turgut, A., et al., High resolution Bottom Shear Modulus Profiler: A Shallow Water Real-time OBS Array, RSMAS TR-87-007, GAL Report 1005, University of Miami, 1987.

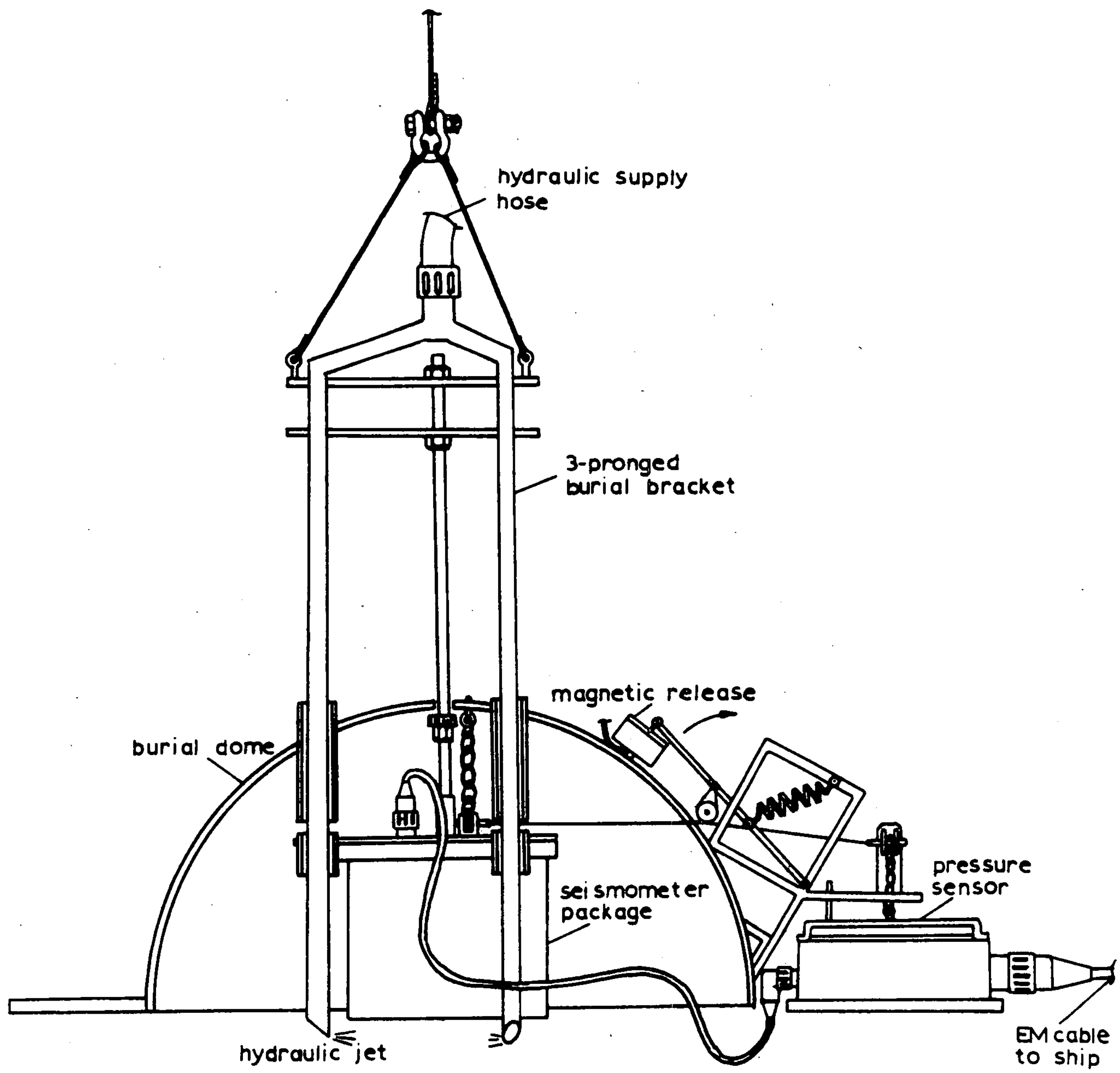
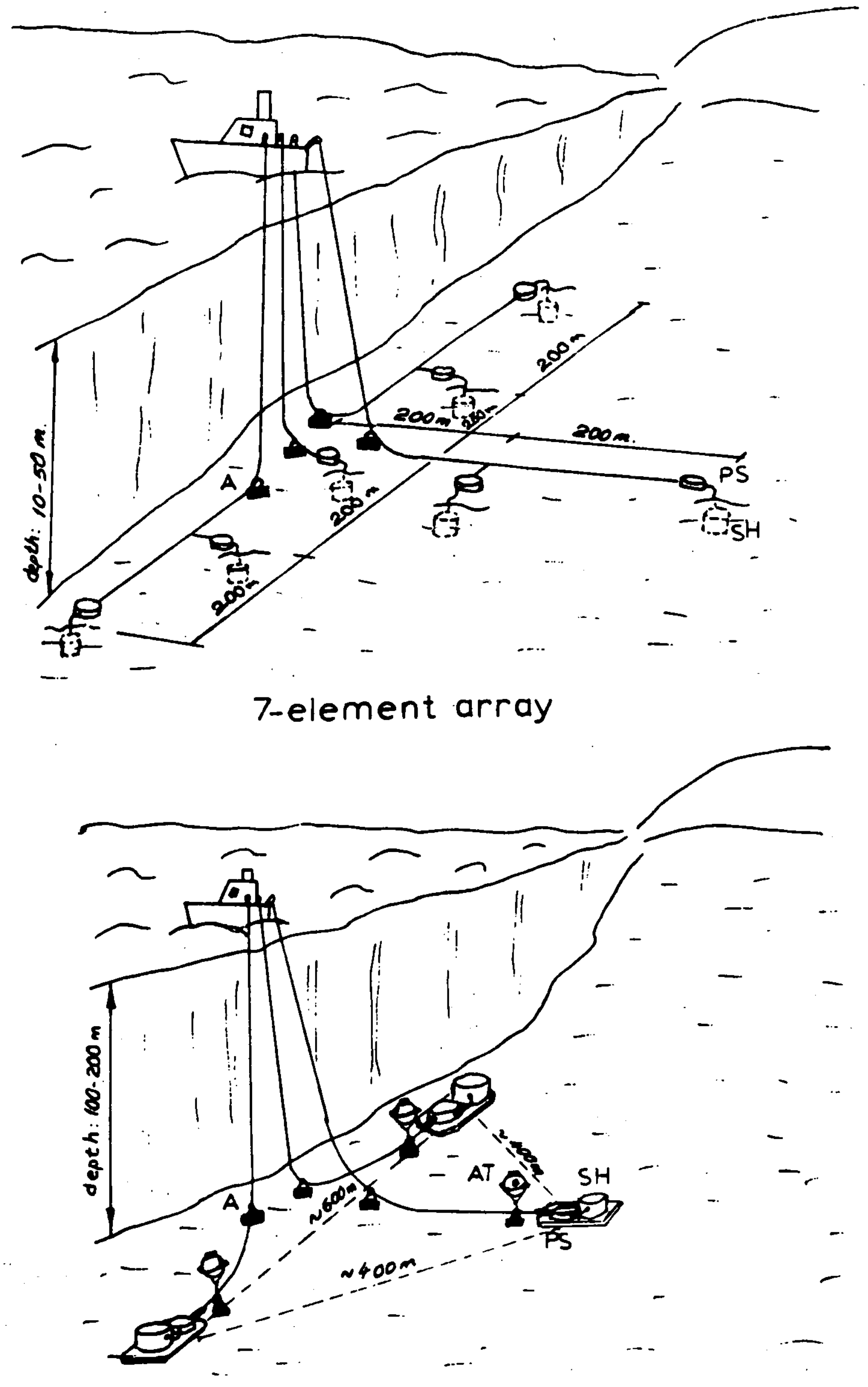


Fig. 1 Hydraulic jet BSMP burial system.



7-element array

3-element array

Fig. 2 Typical array configurations for two different depth ranges. (PS: pressure sensor, SH: seismometer housing, AT: acoustic transponder, A: anchor).

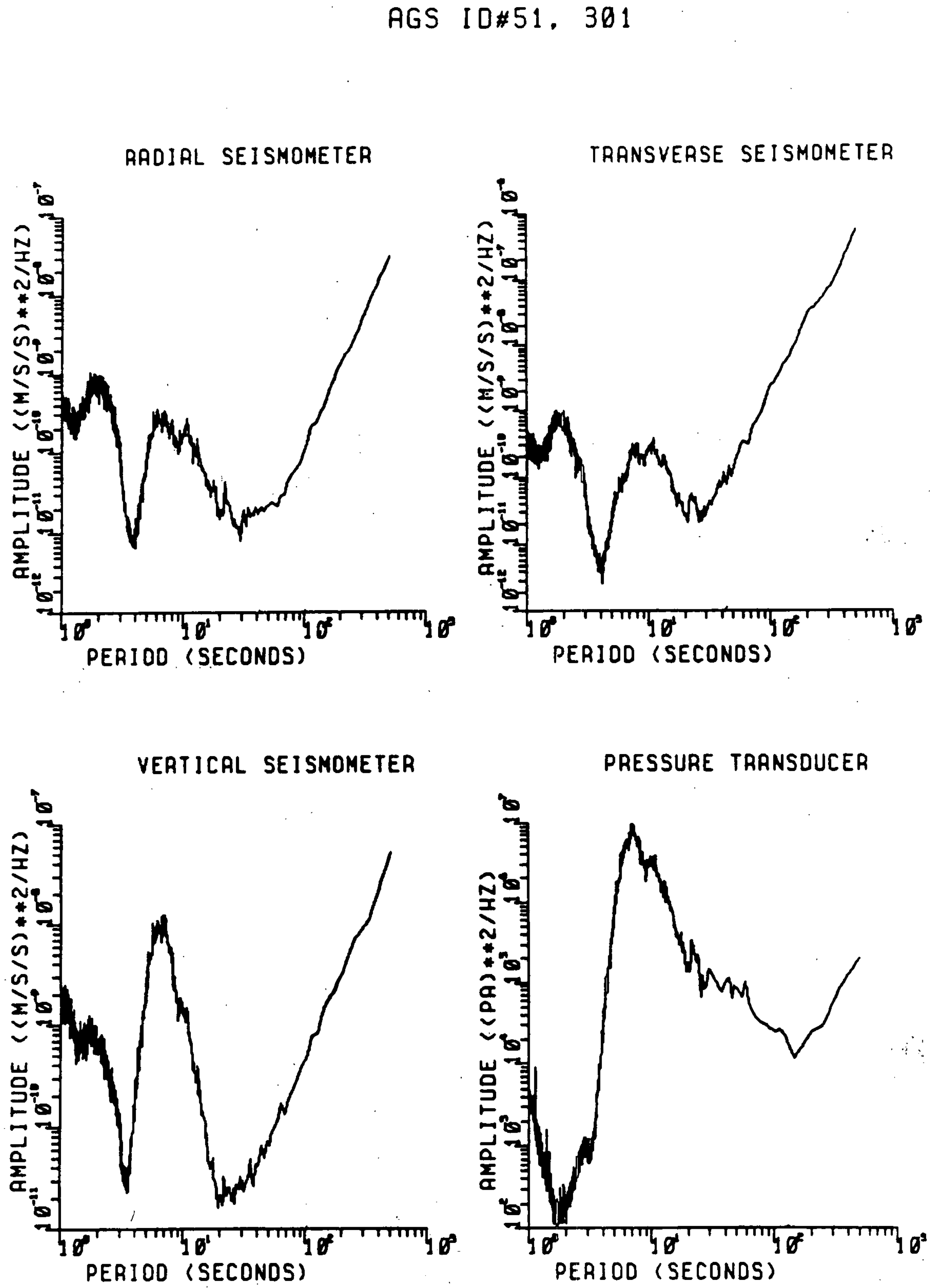


Fig. 3a. Acceleration and pressure power spectra from a buried BSMP at AGS site.

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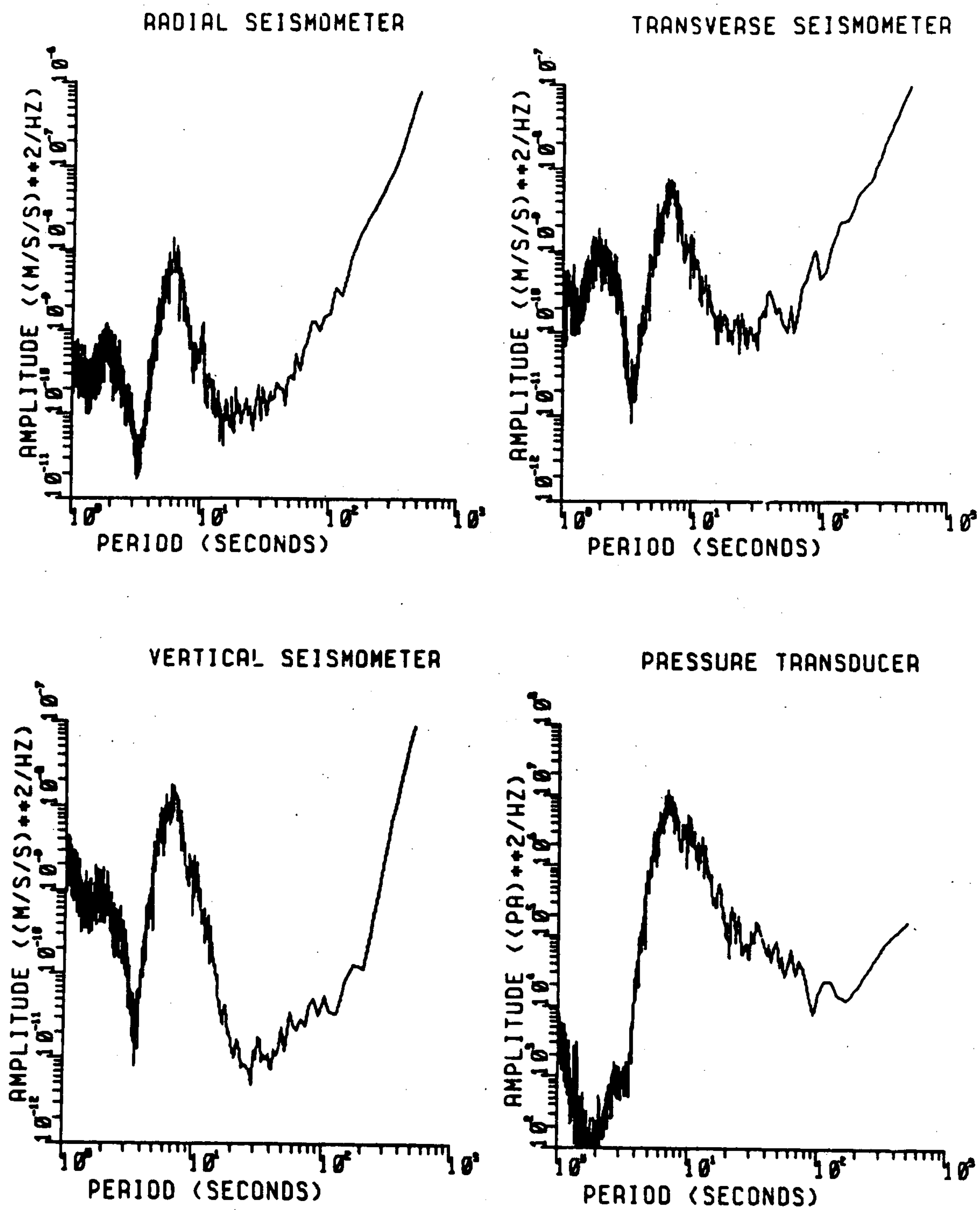

Fig. 3b. Acceleration and pressure power spectra from a plate-mounted BSMP at AGS site.

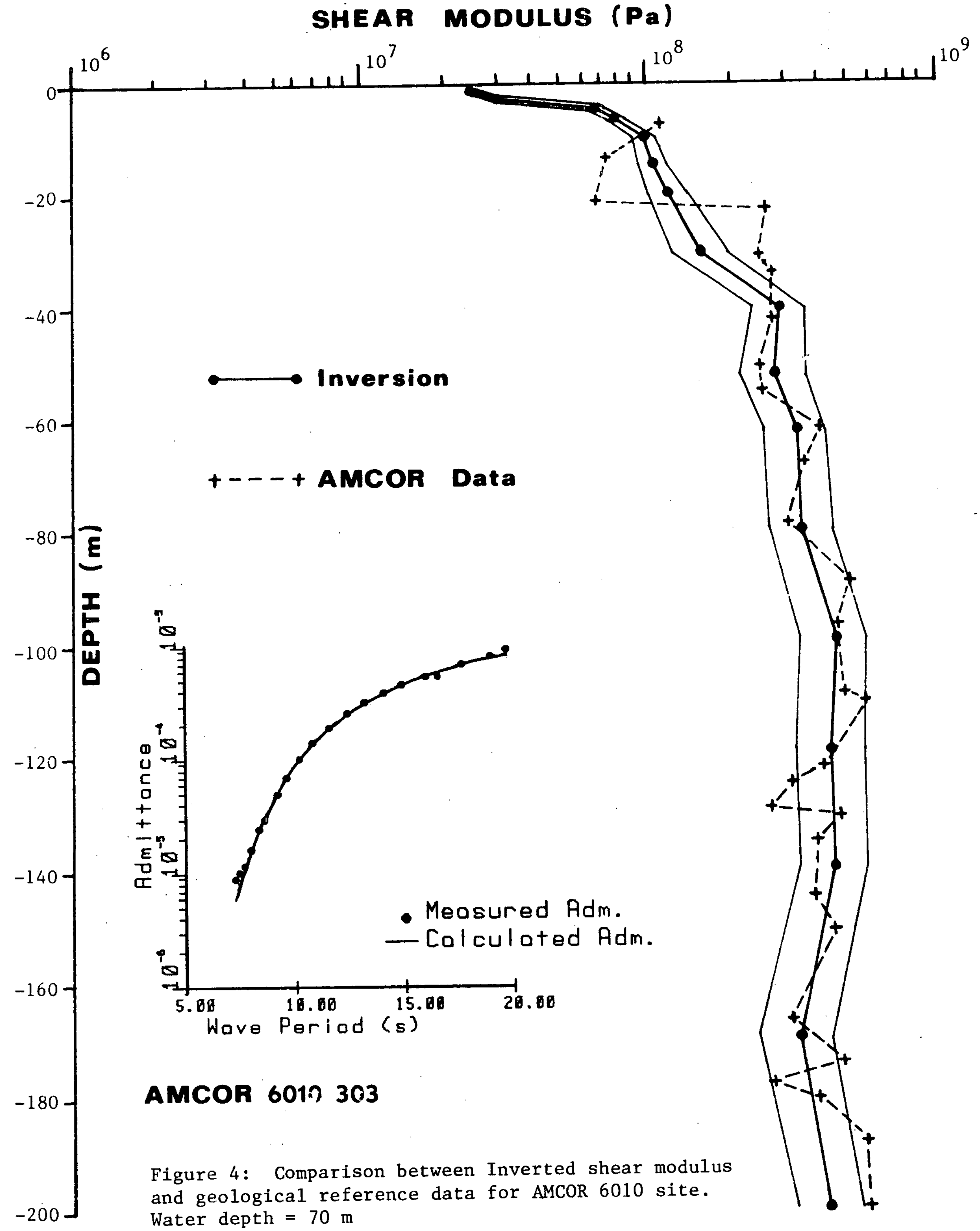

Figure 4: Comparison between Inverted shear modulus and geological reference data for AMCOR 6010 site. Water depth = 70 m

CONTINENTAL DOWNHOLE INSTALLATION

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The U.S. Geological Survey (USGS), Albuquerque Seismological Laboratory (ASL) has installed and operated about 100 stations comprising the Worldwide Standardized Seismograph Network (WWSSN) since 1962, 12 high-gain long-period (HGLP) stations during the 7O's and 12 Seismic Research Observatories (SRO) and 5 Modified HGLP (ASRO) stations since about 1975. The WWSSN system has been highly successful as a global seismograph system for years, but it was realized that the WWSSN LP system was not capable of resolving earth noise in a natural low in the earth's spectrum from 20 to 40 seconds period.

Lamont-Doherty Geological Observatory of Columbia University resolved this problem in the late 1960's and early 70's with the design and deployment of the HGLP Seismograph (Ref. 1, 3, 6, 7). This system used 30-second pendulums and electronic filtering to eliminate the dominant 4-8 second microseismic peak and resolve earth noise in the 20-40 second "window." However, the horizontal seismographs frequently suffered from tilt noise caused by air pressure variations, wind, and daily solar effects, depending on the depth of the seismometer vault.

Efforts to eliminate environmentally-caused tilt noise (which is indistinguishable from horizontal acceleration) and improve signal detection resulted in the development in the early 70's of a high dynamic range, low- noise, broadband, borehole-deployable seismometer, known as a Teledyne-Geotech (T-G) Model 36000, for the SRO project. The idea of using a borehole seismometer came from research showing that wind-caused tilt noise attenuates with depth (Ref. 8, 9, 10). Boreholes for the SRO project were drilled to a depth of 100 meters in rigid rock, such as granite, to take advantage of this and the fact that tilt noise also attenuates more greatly in more rigid rock. SRO systems were deployed at about 12 different sites worldwide by ASL.

The T-G Model 36000 seismometers are 5.5 inches in diameter and about six feet long. They can be installed in standard 7-inch diameter oil well casing. For the SRO installations, the casing is cemented into an 8- to 10-inch drilled hole (see Figure 1), and all SRO boreholes are dry, to reduce horizontal noise caused by convection. Experiments at ASL in installing 36000 seismometers in oil vs. dry environments show that, although LPZ data are unaffected (Figures 2A and 2B), the LPH data are much noisier when the seismometer is installed in oil (Figures 3A, B, C). Figures 3B and 3C show a difference of as much as 20db at periods longer than 20 seconds.

The borehole in Taiwan was originally not well sealed against ground water, and the seismometer gradually became immersed in water. Although these data are for different time periods, Figures 4A through 4E demonstrate the excessive long-period horizontal noise caused by the seismometer being under water in Taiwan. Again, there seems to be little or no effect on LPZ data. Figure 4A also contains a segment of LPZ data from ANMO for comparison to the TATO data. The increase in LPH noise when the seismometer is installed in water or oil is thought to be due to convection. Dry air environments probably also convect, but the mass per unit volume of the fluid is not as great, resulting in less influence on the seismometer.

Teledyne-Geotech has developed a new borehole seismometer known as the Model 54000, which has a flat velocity response from about 0.04 to 4.2 Hz and a more positive module leveling mechanism than the 36000. Figures 5 and 6 show the noise performance of the 54000 vertical and horizontal modules, compared to the ANMO 36000. As seen in Figure 5, the ANMO LPZ component is excessively noisy beyond 30 seconds period. This is a problem that has been recognized for some time. It is probably due to a bad capacitor in the feedback loop. The ANMO seismometer is scheduled for replacement as soon as another good Model 36000 has completed testing. Figure 6 indicates that the 54000 horizontal modules may have somewhat lower LP noise than the ANMO 36000 EW module.

If an instrument's outer casing is sealed, variations in atmospheric pressure acting on the case can distort it enough to cause noise if it is not stiff enough and the distortions are somehow coupled to the mass or suspension. Streckeisen STS-1 vault-type seismometers are used in the China Digital Seismograph Network (CDSN). They are installed on glass plates cemented to the vault floor and sealed inside a glass bell jar in a vacuum, as shown in Figure 7. Teledyne-Geotech is currently developing a broadband sensor based on their short-period S-13 seismometer, known as the BB-13. During April 1988, two BB-13 horizontal instruments were operating in the ASL vault with an STS-1 horizontal instrument. The BB-13 horizontal instrument cases were sealed, but the instrument itself was exposed to atmospheric pressure changes. This apparently resulted in the BB-13 horizontals responding to changes in pressure much more strongly than did the STS-1 horizontal, as seen in Figure 8. (The STS-1 seismometer is decoupled from the glass bell jar.) The conclusion is that the instrument must be protected from pressure changes both on its inertial mass and its outer casing. Distortions of the outer casing are the most plausible explanation of the BB-13's response to pressure changes.

At least one company other than Teledyne-Geotech builds broadband borehole seismometers: Guralp systems of Reading, England. The vault-installable version of the sensor modules are called CMG3 seismometers. Several sets of CMG3 vertical and horizontal sensors were tested at ASL in the summer of 1986 and 1987. The noise performance of the vertical and horizontal sensors in 1987 is shown in Figures 9 and 10. Borehole versions of the CMG3 were installed in Nevada during March 1988 by UCSD and data should be available soon.

New small diameter Streckeisen sensors are expected to be available for ASL vault testing in the fall of 1988. If these tests demonstrate good performance compared to the STS-1 instruments in the ASL vault, the smaller diameter sensors would be a good candidate for borehole packaging.

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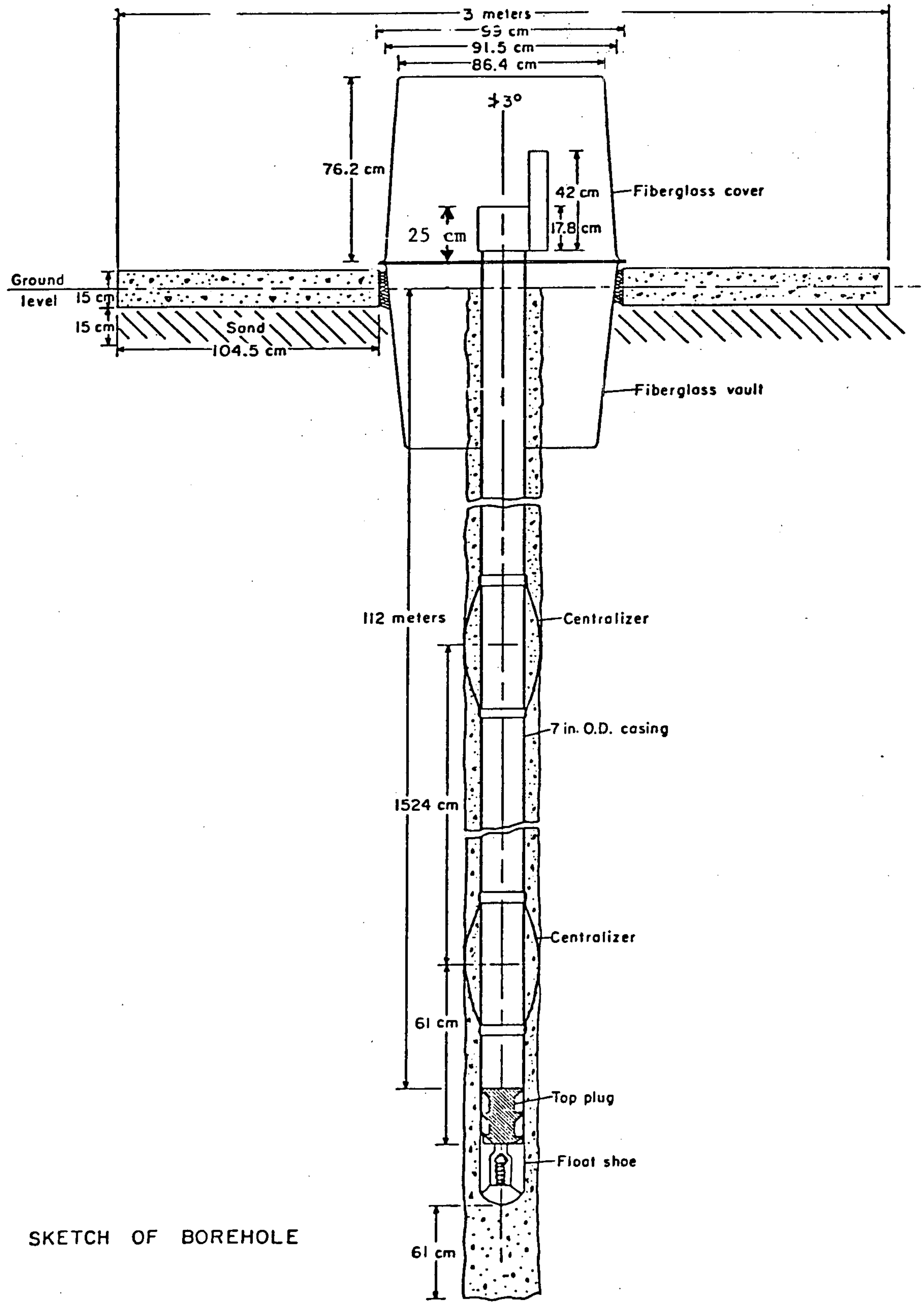


Figure 1 --Sketch of finished borehole.

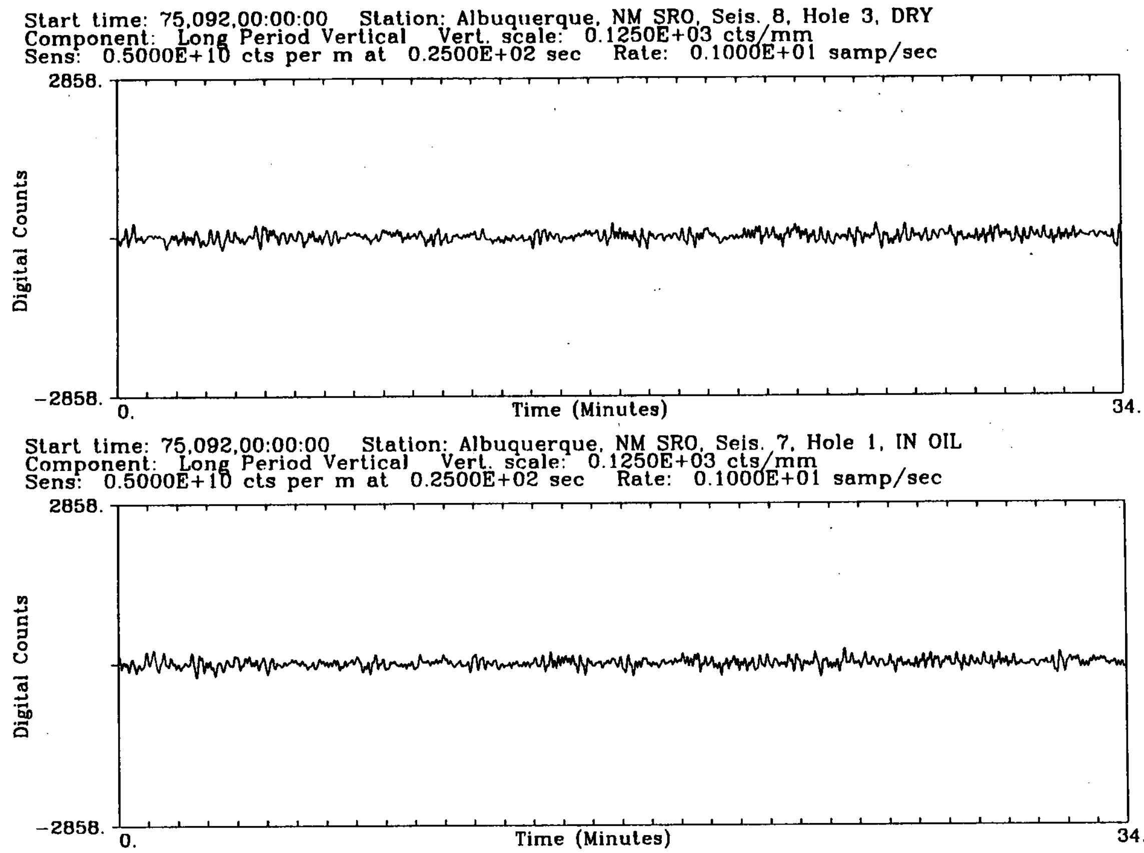
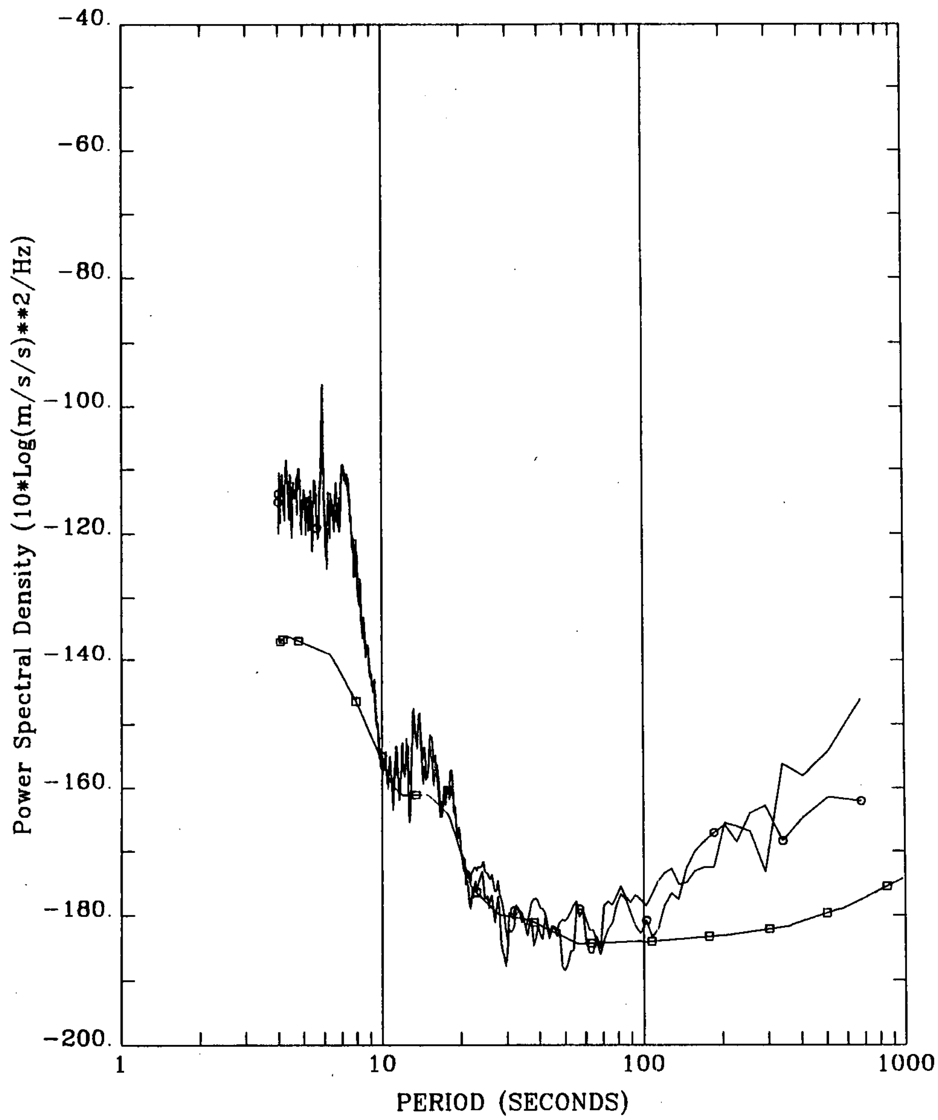


Figure 2A. Vertical LP data in dry vs. oil conditions in Albuquerque hole. Note: In this and all subsequent analog LP waveform plots, the original plotted magnification is 40K at 25 seconds period.



PSD for station: Albuquerque, NM SRO, Seis. 8, Hole 3, DRY
Component: Long Period Vertical Rate: 0.1000E+01 sps
Sensitivity: 0.7916E+11 counts per m/s/s at 0.2500E+02 seconds period.
Start time: 75,092,00:00:00 Segment length: 0hr 34min 8. 0sec

PSD for station: Albuquerque, NM SRO, Seis. 7, Hole 1, IN OIL
Component: Long Period Vertical Rate: 0.1000E+01 sps
Sensitivity: 0.7916E+11 counts per m/s/s at 0.2500E+02 seconds period.
Start time: 75,092,00:00:00 Segment length: 0hr 34min 8. 0sec

Low Noise Model (Acceleration)

Figure 2B. Vertical LP data in dry vs. oil conditions in Albuquerque hole.

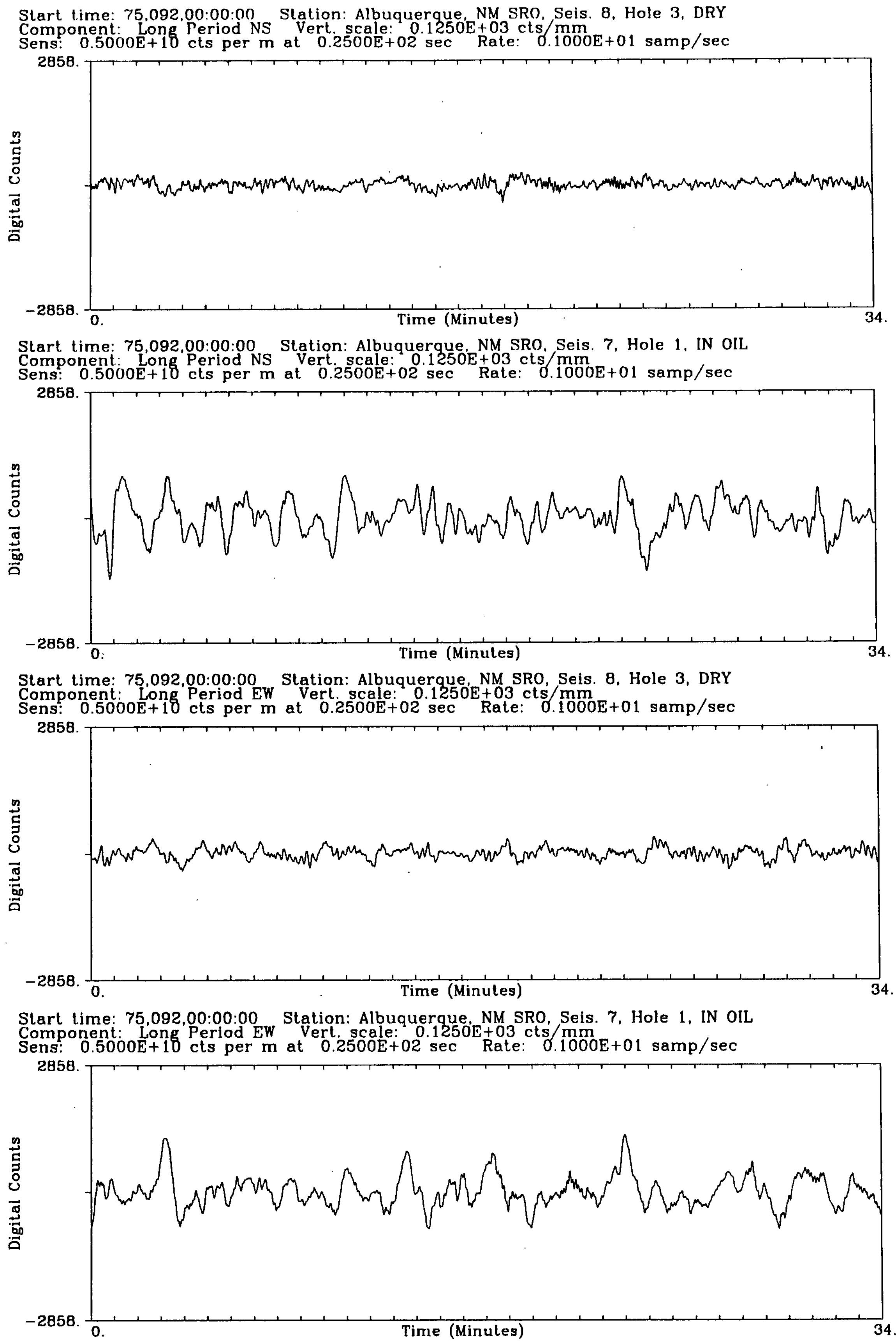
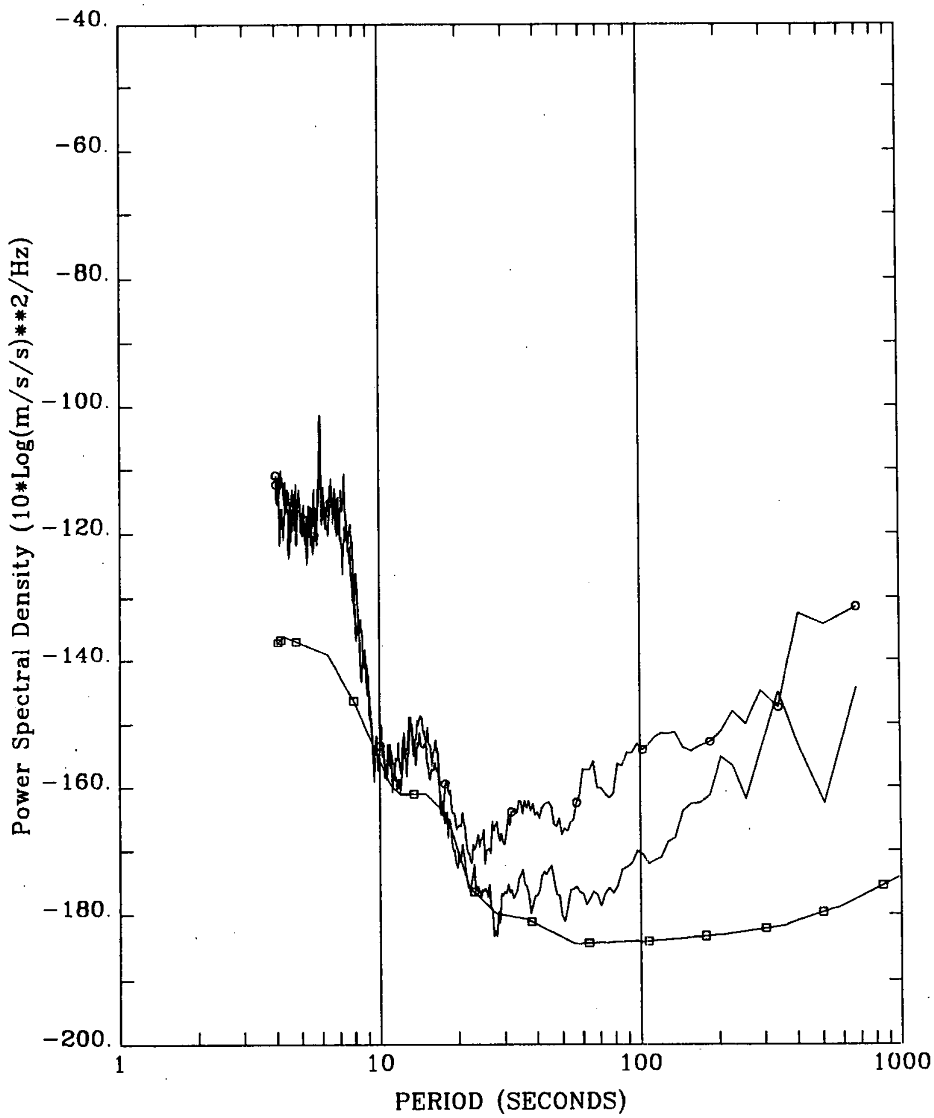


Figure 3A. Horizontal LP data in dry vs. oil conditions in Albuquerque hole.

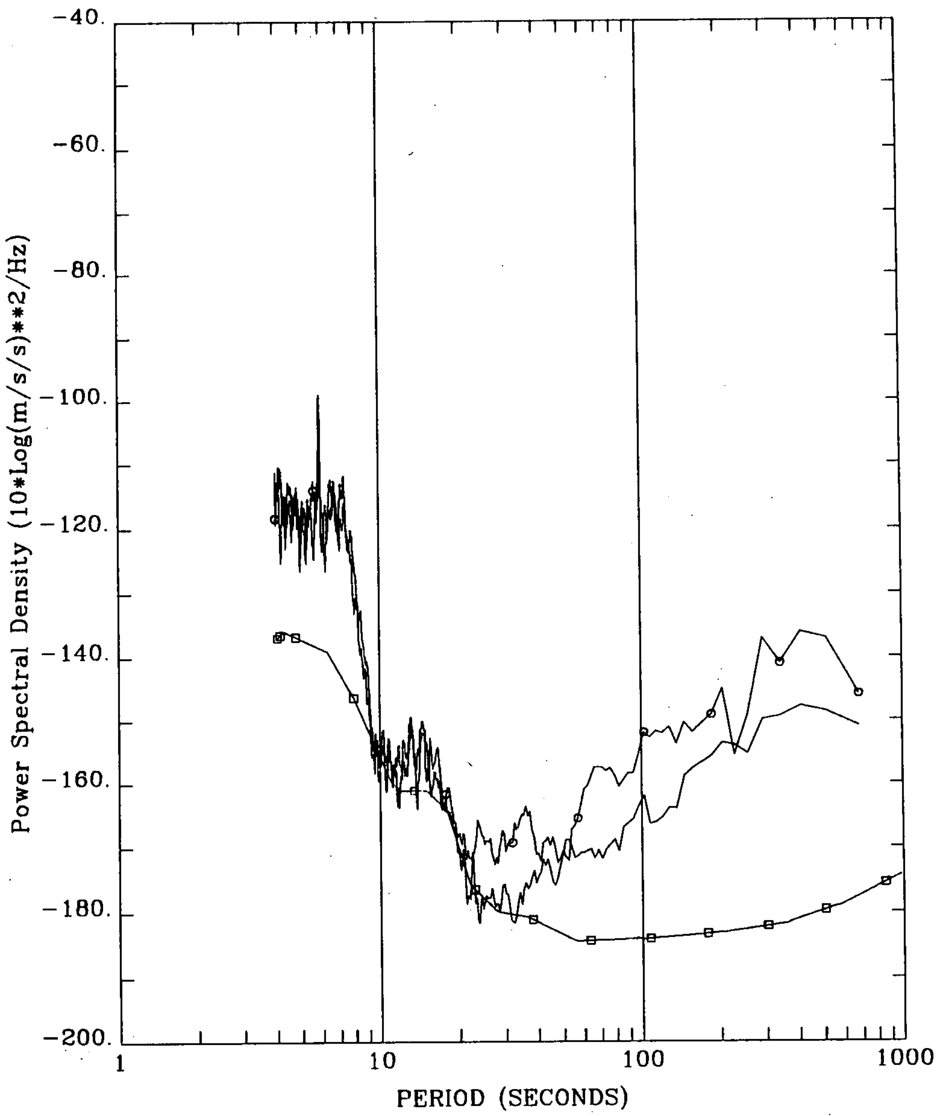


PSD for station: Albuquerque, NM SRO, Seis. 8, Hole 3, DRY
Component: Long Period NS Rate: 0.1000E+01 sps
Sensitivity: 0.7916E+11 counts per m/s/s at 0.2500E+02 seconds period.
Start time: 75,092,00:00:00 Segment length: 0hr 34min 8. 0sec

PSD for station: Albuquerque, NM SRO, Seis. 7, Hole 1, IN OIL
Component: Long Period NS Rate: 0.1000E+01 sps
Sensitivity: 0.7916E+11 counts per m/s/s at 0.2500E+02 seconds period.
Start time: 75,092,00:00:00 Segment length: 0hr 34min 8. 0sec

Low Noise Model (Acceleration)

Figure 3B. Horizontal LP data in dry vs. oil conditions in Albuquerque hole.



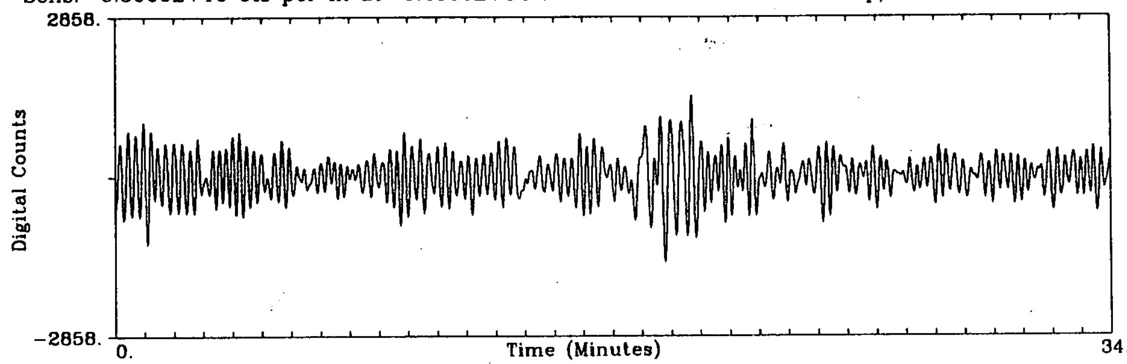
PSD for station: Albuquerque, NM SRO, Seis. 8, Hole 3, DRY
Component: Long Period EW Rate: 0.1000E+01 sps
Sensitivity: 0.7916E+11 counts per m/s/s at 0.2500E+02 seconds period.
Start time: 75,092,00:00:00 Segment length: 0hr 34min 8. 0sec

PSD for station: Albuquerque, NM SRO, Seis. 7, Hole 1, IN OIL
Component: Long Period EW Rate: 0.1000E+01 sps
Sensitivity: 0.7916E+11 counts per m/s/s at 0.2500E+02 seconds period.
Start time: 75,092,00:00:00 Segment length: 0hr 34min 8. 0sec

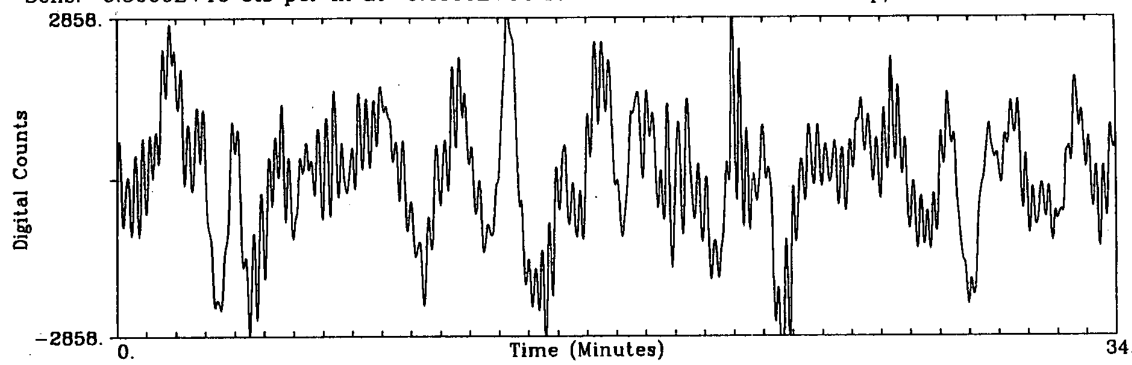
Low Noise Model (Acceleration)

Figure 3C. Horizontal LP data in dry vs. oil conditions in Albuquerque hole.

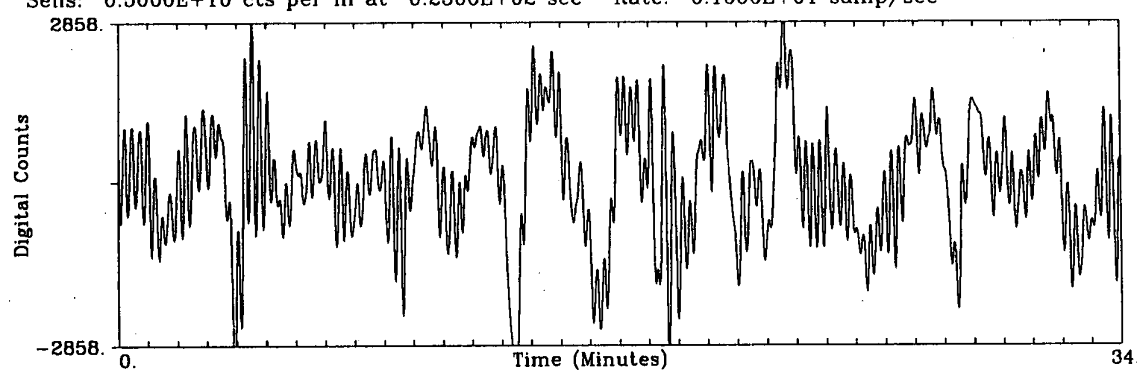
Start time: 76,222,02:00:00 Station: TATO Taipei, Taiwan SRO
Component: LPZ Seismometer under water. Vert. scale: 0.1250E+03 cts/mm
Sens: 0.5000E+10 cts per m at 0.2500E+02 sec Rate: 0.1000E+01 samp/sec



Start time: 76,222,02:00:00 Station: TATO Taipei, Taiwan SRO
Component: LPN Seismometer under water. Vert. scale: 0.1250E+03 cts/mm
Sens: 0.5000E+10 cts per m at 0.2500E+02 sec Rate: 0.1000E+01 samp/sec



Start time: 76,222,02:00:00 Station: TATO Taipei, Taiwan SRO
Component: LPE Seismometer under water. Vert. scale: 0.1250E+03 cts/mm
Sens: 0.5000E+10 cts per m at 0.2500E+02 sec Rate: 0.1000E+01 samp/sec



Start time: 88,036,12:00:00 Station: Albuquerque, NM SRO
Component: LPE Vert. scale: 0.1290E+03 cts/mm
Sens: 0.5170E+10 cts per m at 0.2500E+02 sec Rate: 0.1000E+01 samp/sec

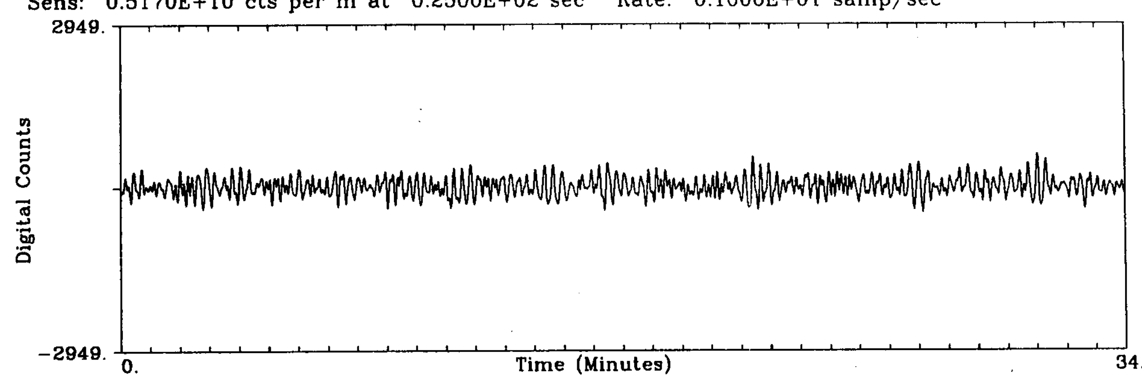


Figure 4A. Comparison of vertical and horizontal LP data in dry vs. underwater conditions in Taipei, Taiwan SRO borehole. This figure is for underwater conditions. Data for ANMO LPZ are shown for comparison.

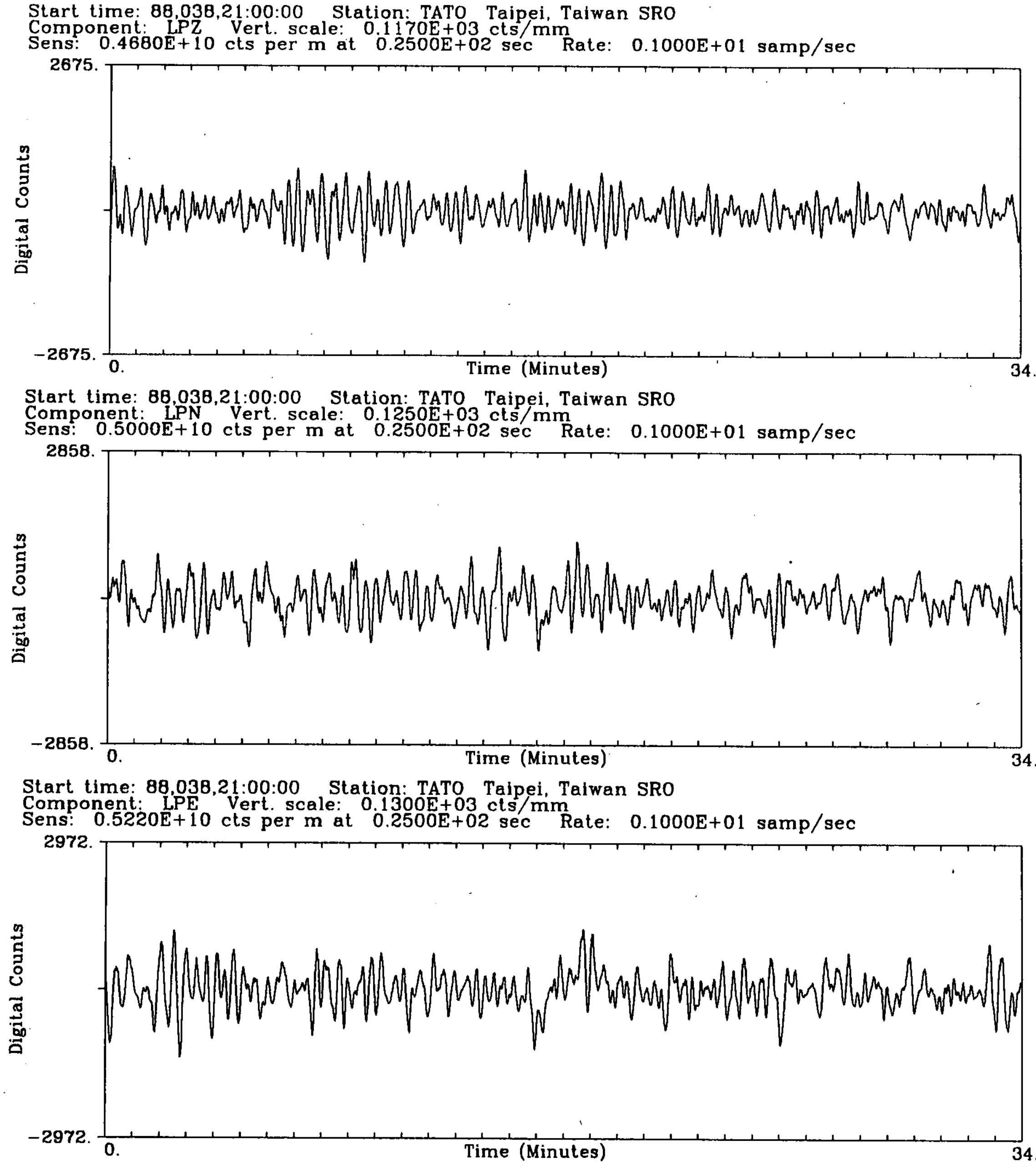
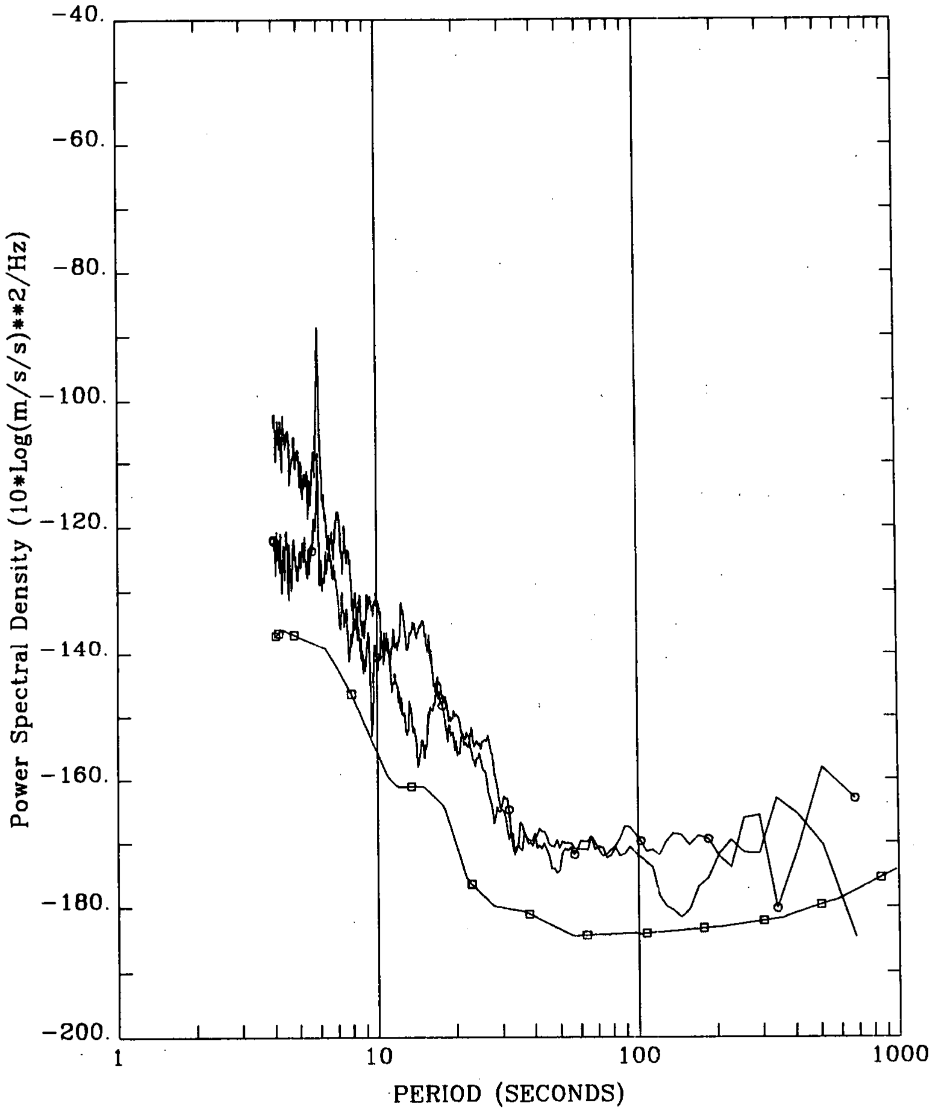


Figure 4B. Vertical and horizontal LP data in dry conditions in Taipei, Taiwan SRO borehole.

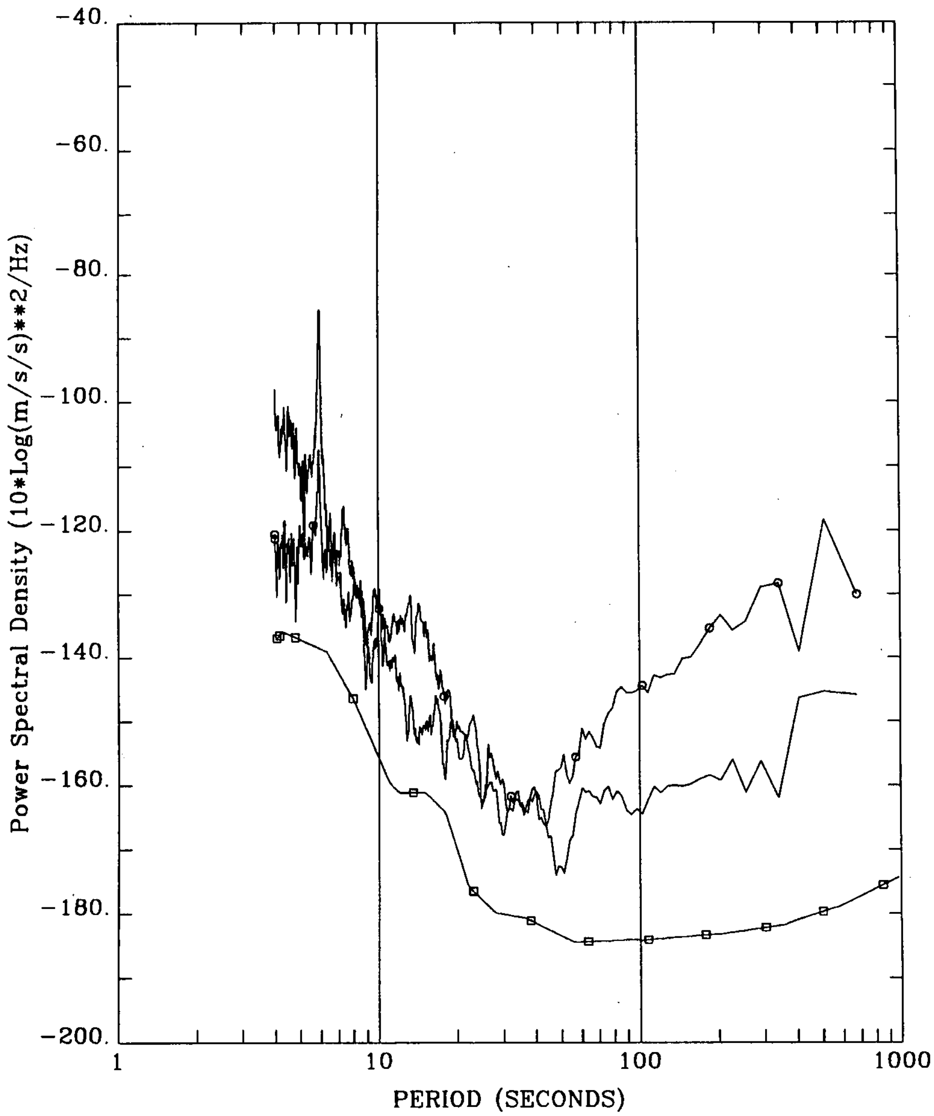


PSD for station: TATO Taipei, Taiwan SRO
Component: LPZ Rate: 0.1000E+01 sps
Sensitivity: 0.7409E+11 counts per m/s/s at 0.2500E+02 seconds period.
Start time: 88,038,21:00:00 Segment length: 0hr 34min 8. 0sec

PSD for station: TATO Taipei, Taiwan SRO
Component: LPZ Seismometer under water. Rate: 0.1000E+01 sps
Sensitivity: 0.7916E+11 counts per m/s/s at 0.2500E+02 seconds period.
Start time: 76,222,02:00:00 Segment length: 0hr 34min 8. 0sec

Low Noise Model (Acceleration)

Figure 4C. Comparison of vertical LP data in dry vs. underwater conditions in Taipei, Taiwan, SRO borehole.

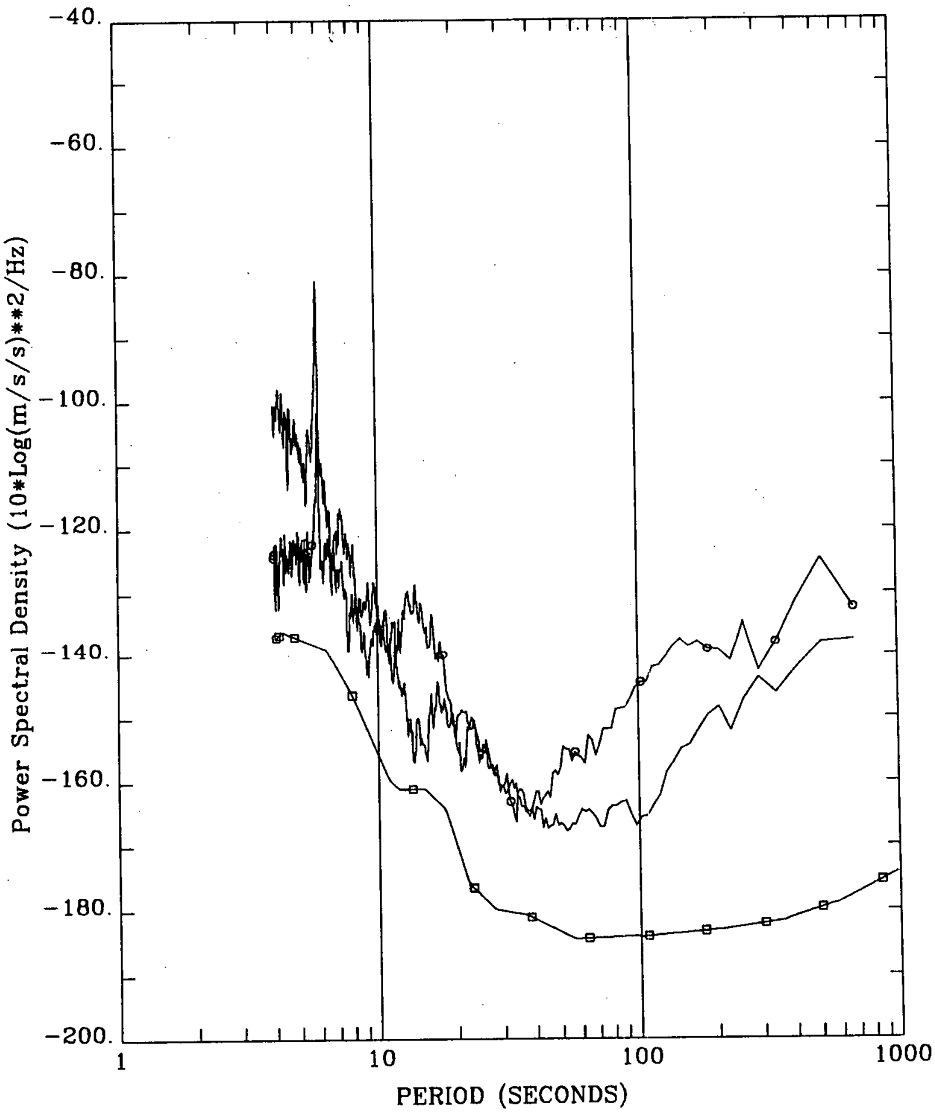


PSD for station: TATO Taipei, Taiwan SRO
Component: LPN Rate: 0.1000E+01 sps
Sensitivity: 0.7916E+11 counts per m/s/s at 0.2500E+02 seconds period.
Start time: 88,038,21:00:00 Segment length: 0hr 34min 8. 0sec

PSD for station: TATO Taipei, Taiwan SRO
Component: LPN Seismometer under water. Rate: 0.1000E+01 sps
Sensitivity: 0.7916E+11 counts per m/s/s at 0.2500E+02 seconds period.
Start time: 76,222,02:00:00 Segment length: 0hr 34min 8. 0sec

Low Noise Model (Acceleration)

Figure 4D. Comparison of horizontal (NS) LP data in dry vs. underwater conditions in Taipei, Taiwan, SRO borehole.

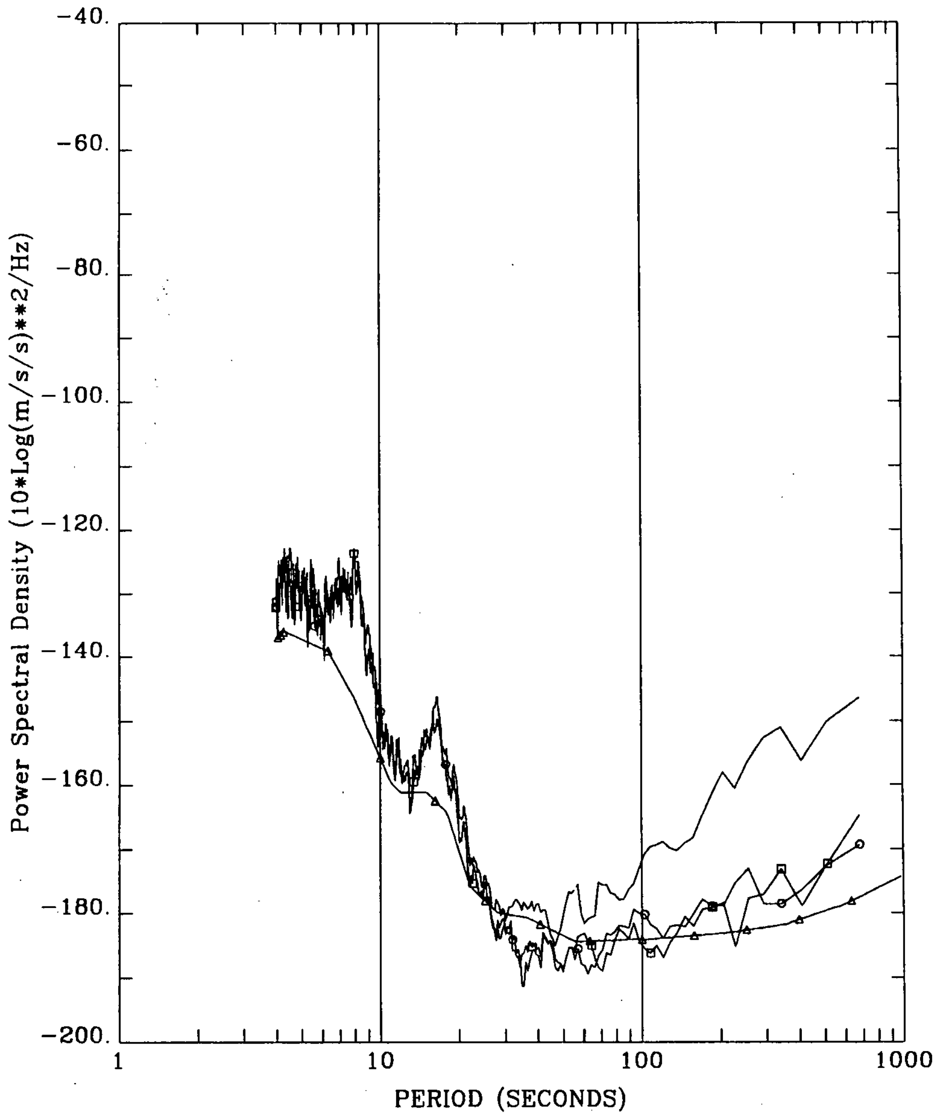


———— PSD for station: TATO Taipei, Taiwan SRO
Component: LPE Rate: 0.1000E+01 sps
Sensitivity: 0.8264E+11 counts per m/s/s at 0.2500E+02 seconds period.
Start time: 88,038,21:00:00 Segment length: 0hr 34min 8. 0sec

o——o——o PSD for station: TATO Taipei, Taiwan SRO
Component: LPE Seismometer under water. Rate: 0.1000E+01 sps
Sensitivity: 0.7916E+11 counts per m/s/s at 0.2500E+02 seconds period.
Start time: 76,222,02:00:00 Segment length: 0hr 34min 8. 0sec

□——□——□ Low Noise Model (Acceleration)

Figure 4E. Comparison of horizontal (EW) LP data in dry vs. underwater conditions in Taipei, Taiwan, SRO borehole.



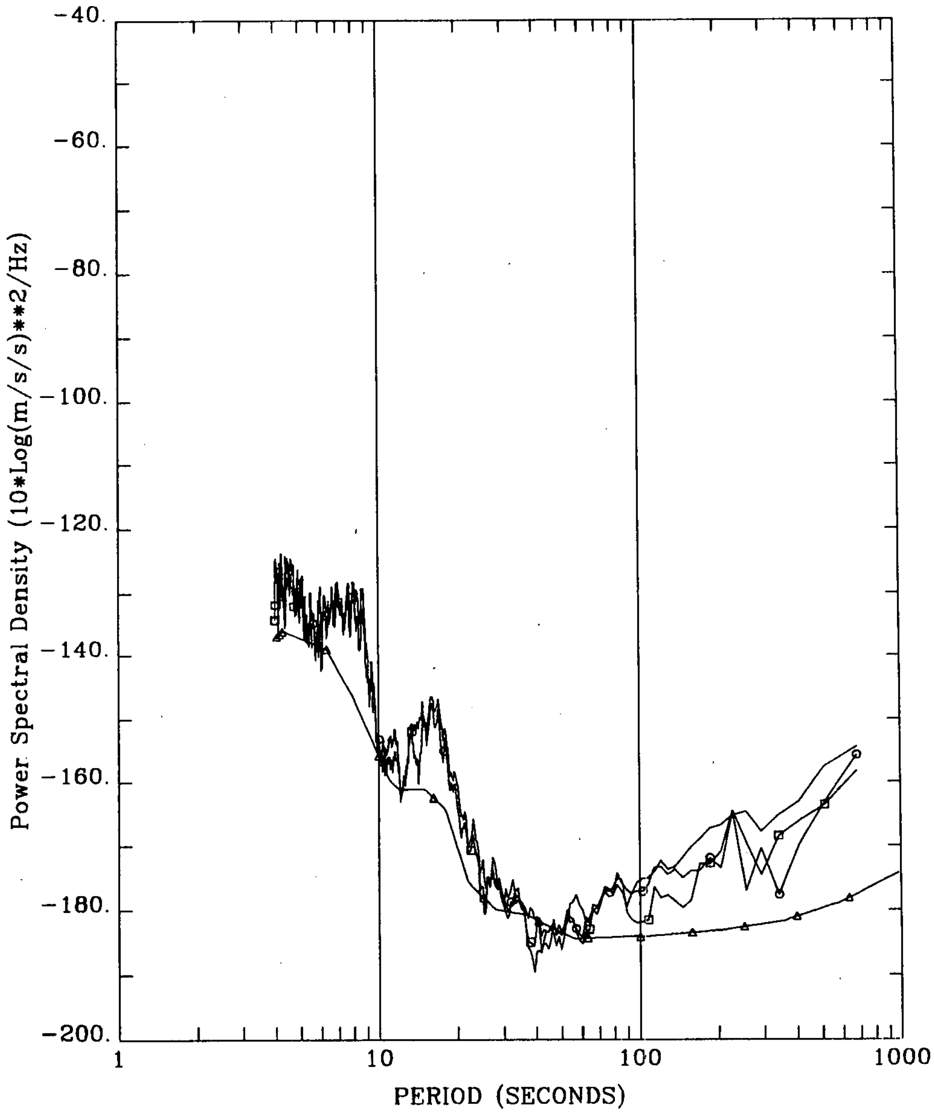
———— PSD for station: Albuquerque, NM SRO
Component: LPZ Rate: 0.1000E+01 sps
Sensitivity: 0.6806E+11 counts per m/s/s at 0.2500E+02 seconds period.
Start time: 88,036,12:00:00 Segment length: 0hr 34min 8. 0sec

o——o——o PSD for station: ALQX 54000 Seis. Tests
Component: lpz1.1 (N. hole) Rate: 0.1000E+01 sps
Sensitivity: 0.1001E+12 counts per m/s/s at 0.5000E+02 seconds period.
Start time: 88,036,12:00:00 Segment length: 0hr 34min 8. 0sec

□——□——□ PSD for station: ALQX 54000 Seis. Tests
Component: lpz2.2 (S. hole) Rate: 0.1000E+01 sps
Sensitivity: 0.1001E+12 counts per m/s/s at 0.5000E+02 seconds period.
Start time: 88,036,12:00:00 Segment length: 0hr 34min 8. 0sec

△——△——△ Low Noise Model (Acceleration)

Figure 5. LPZ noise performance of Teledyne-Geotech Model 54000 vs. ANMO SRO (Model 36000) seismometers. The ANMO LPZ component is known to be excessively noisy due to an electronic problem.



PSD for station: Albuquerque, NM SRO
Component: LPE Rate: 0.1000E+01 sps
Sensitivity: 0.8185E+11 counts per m/s/s at 0.2500E+02 seconds period.
Start time: 88,036,12:00:00 Segment length: 0hr 34min 8. 0sec

PSD for station: ALQX 54000 Seis. Tests
Component: lpe1.1 (N. hole) Rate: 0.1000E+01 sps
Sensitivity: 0.4641E+12 counts per m/s/s at 0.5000E+02 seconds period.
Start time: 88,036,12:00:00 Segment length: 0hr 34min 8. 0sec

PSD for station: ALQX 54000 Seis. Tests
Component: lpe1.2 (N. hole) Rate: 0.1000E+01 sps
Sensitivity: 0.4641E+12 counts per m/s/s at 0.5000E+02 seconds period.
Start time: 88,036,12:00:00 Segment length: 0hr 34min 8. 0sec

Low Noise Model (Acceleration)

Figure 6. LPH (EW) noise performance of Teledyne-Geotech Model 54000 vs. ANMO SRO (Model 36000) seismometers.

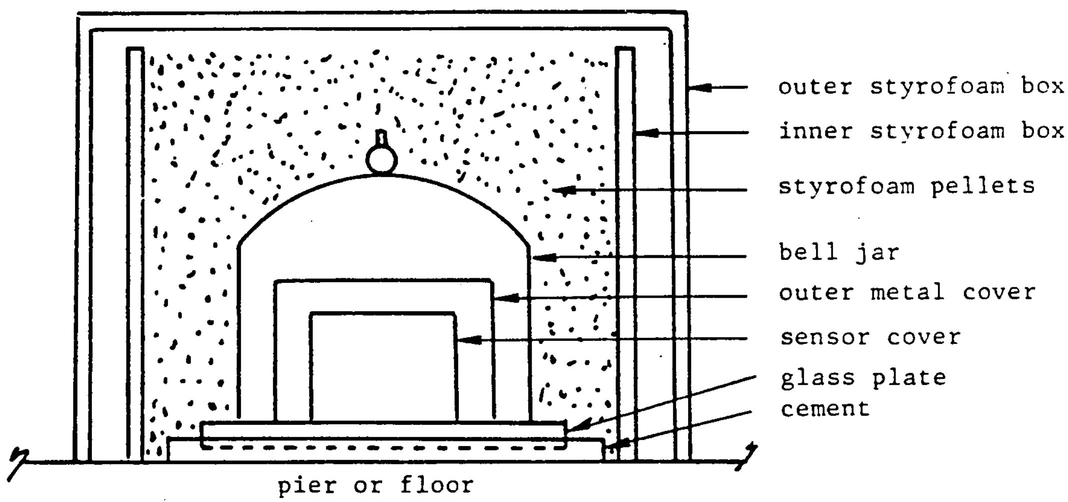


Figure 7 .--Illustration of STS-1 installation technique used in the CDSN.

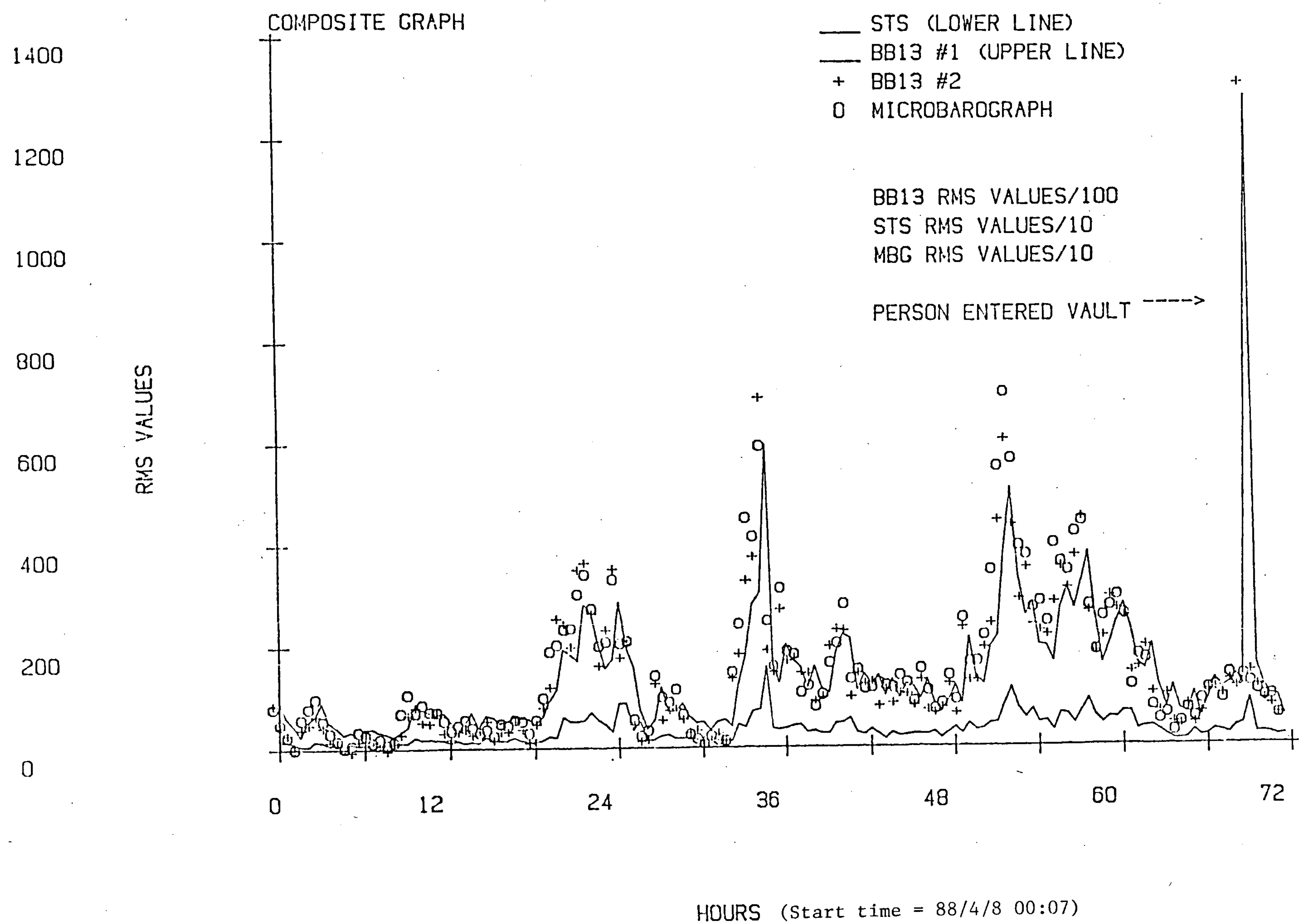


Figure 8. Comparison of LPH noise of STS-1 and BB-13 seismometers during windy conditions at Albuquerque. Microbarograph activity in the vault is also plotted.

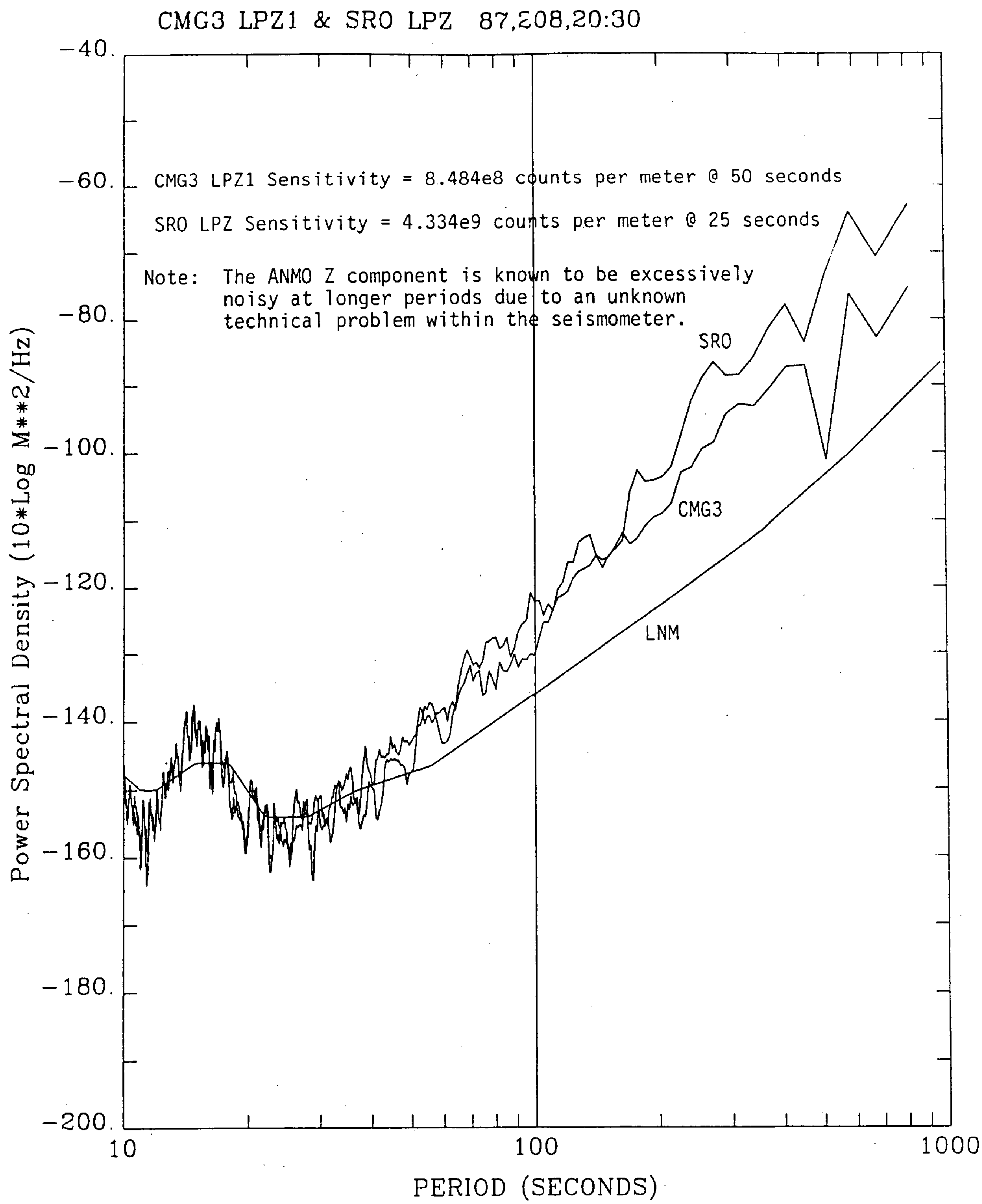


Figure 9. Comparison of CMG3 LPZ noise performance to that of ANMO SRO LPZ.

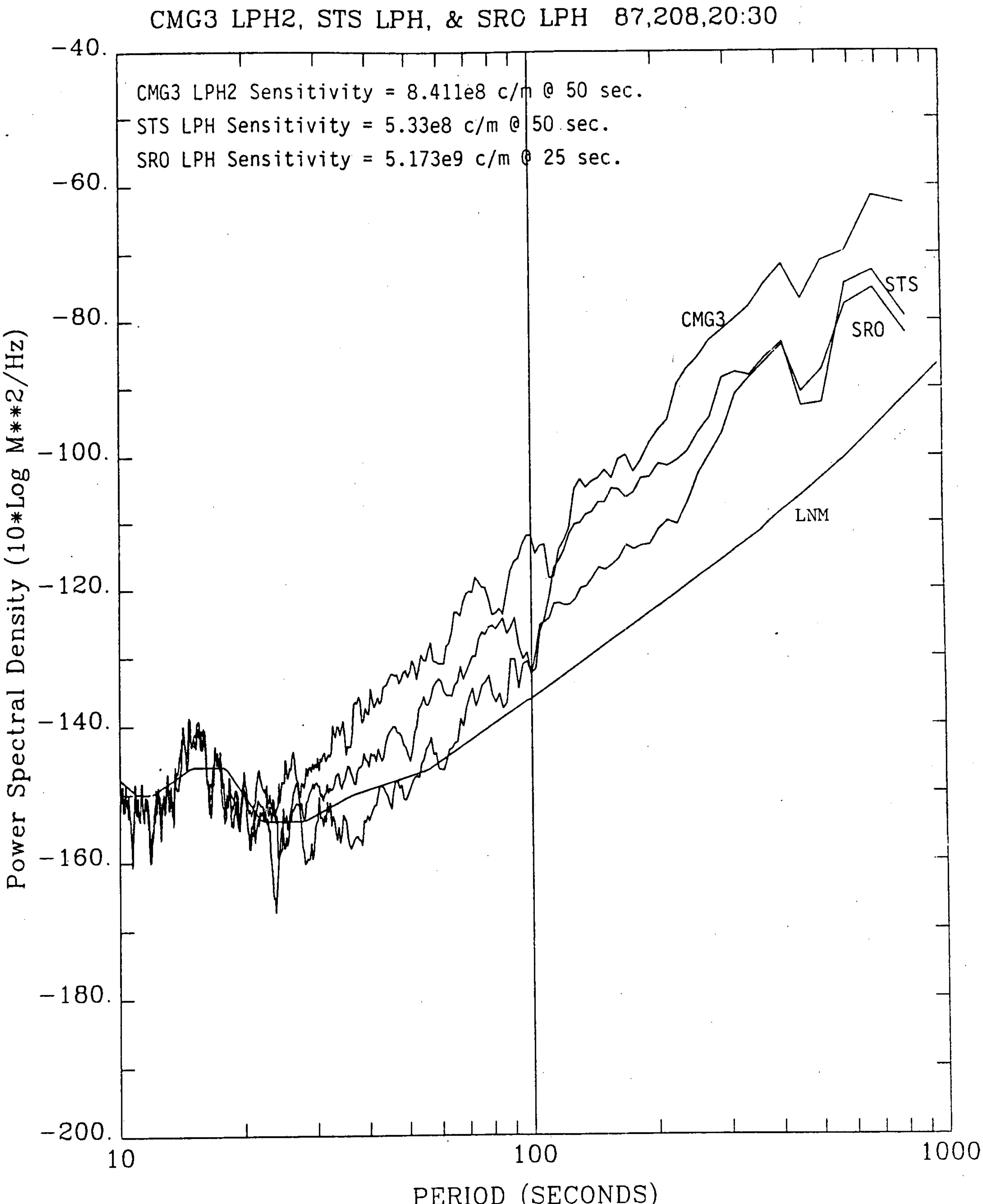


Figure 10. Comparison of CMG3 LPH noise performance to that of ANMO SRO LPH and STS-1 LPH.

SEISMIC BROADBAND SIGNAL AND NOISE LEVELS ON AND WITHIN THE SEAFLOOR AND ON ISLANDS

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John A. Orcutt, and Dalia Lahav
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Abstract

Ambient noise levels recorded on and beneath the seafloor are compared with the signal levels expected at an epicentral range of 30° from events ranging in M_w from 5.0 to 9.5. The results indicate that at this range events with M_w as low as 5.5 should be detectable by sensors on the seafloor in the Pacific. Because of lower ambient noise levels present in the Atlantic and beneath the seafloor in the Pacific, events with M_w as low as 5.0 should be detectable in these locations.

Noise level data collected by three land-based IDA/IRIS sensors (RPN, ESK and PFO) ranging in distance from 1 to 100 km from the shoreline are compared. Low frequency noise levels (below 50 mHz) are not sensitive to this distance although at frequencies above 50 mHz noise levels are clearly inversely proportional to the distance from the shoreline. At 250 mHz noise power spectral levels drop 40 dB as the observation point varies from 1 to 100 km from the coast.

Introduction

It is obvious that the global coverage which land based seismometers provide is limited since 71% of the Earth's surface is covered by water. A much more uniform coverage is feasible if instruments can be deployed on the seafloor in environments with signal-to-noise ratios comparable to those available on land. Many technical deployment problems have been overcome during the past 30 years as seismologists have deployed seismometers on and beneath the sea floor. A previous paper (Hedlin and Orcutt, 1989) has already dealt with the question of the relative noise levels on land and the seafloor. It was concluded that over the frequency band from 40 mHz to 10 Hz the ocean bottom and island noise levels are comparable. In this paper we look more deeply into the acceptability of seafloor noise levels by comparing them with signal levels expected from earthquakes of varying sizes at an epicentral range of 30°.

Although seafloor sites could potentially provide a uniform global coverage, island seismic stations still play an important role in the global seismographic network since they provide a means for deploying sensors in oceans relatively inexpensively and with relative ease. In this paper we will also examine the dependence of island and continental noise levels on the distance from the shoreline.

The Data

The sub-seafloor noise level data were collected by the Marine Seismic System (MSS) in February of 1983 during the Scripps' Ngendei Expedition and the noise analysis was originally published by Adair et al (1986). The MSS was deployed at 23.8°S, 165.5°W in DSDP drill hole 595b (leg 91), 124 meters below the sediment-water interface and 54 meters into the basement. The MSS data were obtained from short (40 samples/s) and mid-period (4 samples/s) channels with time series which were roughly 13 and 128 seconds long respectively. To reduce the variance of the power spectral estimates, the short and mid-period curves are based on the average of 10 spectra. The Ocean Bottom Seismometer (OBS) noise level bounds were computed from 25 noise level estimates obtained from data collected at the locations and dates detailed in table 1. Each individual estimate was made after stacking from 3 to 125 noise level curves although the individual time series collected were too short to obtain high resolution at frequencies approaching 0.2 Hz. Both MSS and OBS power estimates have been corrected for instrument response. The MSS and OBS instruments are discussed in detail in Adair et al (1986) and Orcutt et al (1987).

The data for the high-frequency seafloor pressure spectra were collected on two separate occasions. The Pacific data were collected in May 1988 during the Scripps' Nachos Expedition 350 km west of San Diego in 3850 m of water. A low frequency hydrophone (Cox et al, 1984) mounted on the side of an OBS was sampling at 32 samples/s and the recorder collected 15 minute records. The 28,800 point time series were divided into 30% overlapping 8192 point windows. Each series was Hanning tapered prior to Fourier transformation and then corrected for the instrument response. The final spectral estimate was obtained from a stack of the resulting spectra. The Atlantic seafloor pressure spectrum was computed from data collected by the Woods Hole Oceanographic Institution during the HEBBLE expedition in October of 1985. The HEBBLE pressure data were collected by a long period transducer (Cox et al, 1984) deployed at 4820 m depth over the Nova Scotia Rise. The hydrophone data were sampled at one sample/s during 6 hr time periods each day. To compute the power spectral estimate, the data were treated in a similar manner as the Atlantic high frequency data, the only difference being that 4 π prolate windows, rather than Hanning windows, were used to taper each sub-sequence.

The land-based data were collected exclusively at IDA/IRIS stations. Low (1 sample/s) and high frequency (20 samples/s) spectral estimates have been merged to provide the broadband spectra in figure 3. Each spectrum is the result of stacking, in the frequency domain, several spectral estimates calculated from distinct portions of the original noise sample. The very low frequency seafloor pressure data were obtained from figure 6 of Filloux (1980).

Comparison of Seafloor Signal and Noise Levels

Figure 1 illustrates three seafloor pressure power spectra spanning the band from 0.7 μHz to 20 Hz. The solid and finely-dashed curves represent data collected in the Pacific while the data for the lower power, coarsely-dashed curve were collected in the Atlantic. Power spectral levels in the lowest band, from 0.7 μHz to 20 mHz, decrease monotonically at 25 dB/decade except in the band from 10 to 200 μHz where several tidal lines are recorded. The microseism band in both spectra is dominated by the double frequency microseism peaks above 200 mHz. Between the microseism peaks and 2 Hz, power spectral levels decrease at 40 dB/decade.

In figure 2 the high-frequency data in Figure 1 have been converted to acceleration units and superimposed on curves which illustrate the acceleration expected at an epicentral range of 30° from the earthquakes spanning a range of magnitudes. OBS and MSS data have been plotted as well. The conversion from pressure to amplitude was made using a simple acoustic plane wave relationship assuming vertical incidence of seismic energy and continuity of vertical traction and displacement. Although the validity of this conversion is clearly restricted, we expect that the predicted accelerations are probably realistic estimates. The validity of the noise levels can only be verified through experimentation. This figure is very encouraging and indicates that with the exception of the noisiest OBS sites all events with M_W of 5.5 or greater lead to motions above the expected noise levels at a range of 30°. Such earthquakes or explosions should be detectable with seafloor sensors. Using hydrophones deployed in the Atlantic and the MSS buried in the substrate beneath the Pacific, where ambient noise levels appear to be somewhat lower, earthquakes with M_w as low as 5.0 should be detectable at a range of 30°. This conclusion is consistent with the results of Orcutt and Jordan (1985). Using data collected by the MSS during the Ngendei Expedition they found that no events with m_b less than 5.1 were detected at ranges greater than 30°. At lesser distances, where high frequency signals propagate through the high-Q lithospheric waveguide, they found that the detection threshold was far lower. Shallow-focus events at distances on the order of 15° with m_b as low as 3.7 could be detected by the MSS. At the noisiest sites, depending on the frequency band of interest, the smallest events detectable range in size from M_w of 6 to 8.

The Dependence of Land-Based Noise Levels on the Distance from the Shoreline

Figure 3 shows power spectra of noise data obtained from the IRIS/IDA seismographic stations on Easter Island (RPN), one km from the coastline, at Eskdalemuir (ESK), 40 km from the Irish Sea and at Pinon Flat (PFO), some 100 km from the California coast. The dashed lines show the level of minimum and maximum ambient noise at quiet continental locations. The high frequency portion of the spectrum for the RPN station was derived from data collected during the most noisy period experienced during a one-year period; it should be taken as representative of the maximum high frequency noise at an island site. The spectra demonstrate that for periods longer than the microseisms, there is only a small decrease in noise as distances to the coastline increase. Island stations should provide relatively low noise data in this frequency range. At frequencies higher than the microseisms, there is considerable reduction in noise as the distance from the coastline increases. At 250 mHz the noise levels drop 40 dB. Factors such as coastal shoaling, barrier reef coverage, and surf conditions can be expected to cause large variations in the ambient noise in this frequency range.

Conclusion

It appears that most locations on the seafloor are sufficiently quiet to permit the detection of moderate-sized earthquakes and explosions at teleseismic ranges. This result is particularly encouraging in view of the need for an increased uniformity of the global coverage of seismic sensors.

The acceptability of island sensor deployments depends on the frequency band of interest and on the distance of the sensor from the shoreline (and thus the size of the island). It appears that low frequency (below 50 mHz) noise levels are insensitive to the distance of the sensor from the shoreline. However, in high-frequency studies it appears that sites on large islands relatively far from the shoreline will have superior noise levels.

Acknowledgements

We would like to express our appreciation to Spahr Webb for providing us with the Atlantic hydrophone data in advance of publication, and J.H. Filloux for the low-frequency Pacific data.

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TABLE AND FIGURE CAPTIONS

Table 1. The geographic coordinates and dates at which the OBS noise samples used in this study were collected.

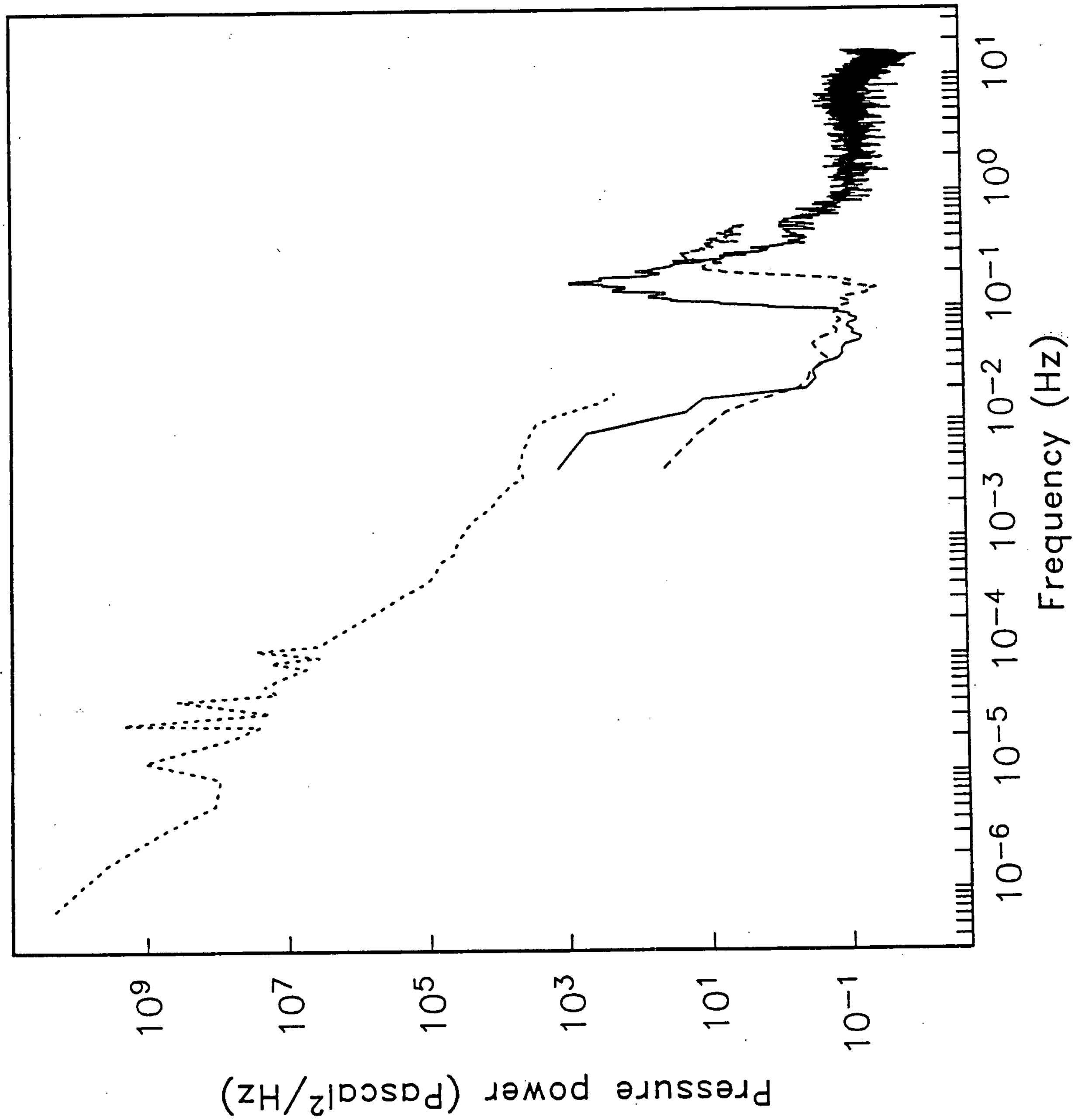
Figure 1. Noise power spectral density in pressure units for three seafloor sensors. The data for the solid and finely-dashed curves were collected in the Pacific Ocean, off the Coast of California (in 3850 m of water) and off the Gulf of California (at 3210 m depth) respectively. The data for the coarsely-dashed curve were collected in the Atlantic at the Nova Scotia Rise in 5000 m of water.

Figure 2. Acceleration amplitudes expected at an epicentral range of 30° from earthquakes ranging in M_W from 5.0 to 9.5. The data for the light-solid curve were collected in the Pacific Ocean, off the coast of California at 3850 m depth. The two heavy-solid curves represent the extreme noise level bounds detected at 25 sites in the Pacific (see table 1). The frequently dashed curve represents the noise level recorded by the Marine Seismic System deployed during the southwest Pacific Ngendei Expedition. The data for the infrequently dashed curve were collected in the Atlantic at the Nova Scotia Rise in 5000 m of water. The amplitudes plotted are 1.25 times the RMS values over 1/3 octave bandwidths.

Figure 3. Noise recordings made on an island (RPN-Easter Island) compared with measurements made on continental sites located near the ocean (ESK-Eskadalemuir; PFO-Pinon Flat Observatory). Note that the high frequency content of the microseism noise decreased as the distance from the shore increases. The amplitude of the microseism peak also decreases in the same sense as the high frequencies.

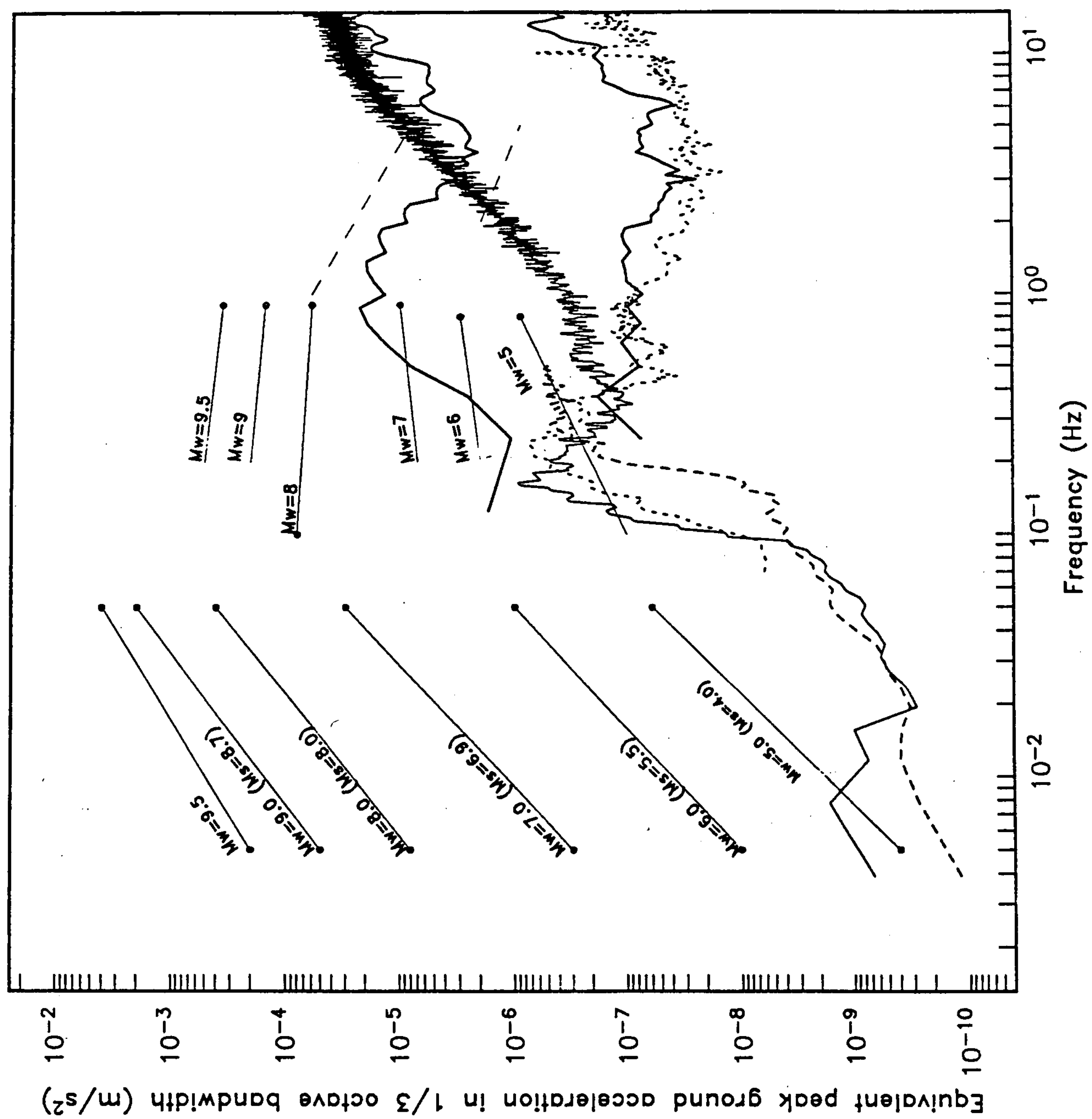
Table 1 OBS data

NUMBER OF SAMPLES	LATITUDE	LONGITUDE	DATE
6	41° 30' N	127° 20' W	JUN 1976
2	16° 0' N	145° 0' W	OCT 1977
1	29° 30' N	122° 0' W	OCT 1977
3	18° 0' N	145° 20' W	JAN 1976
8	16° 30' N	100° 30' W	JUN 1977
4	20° 50' N	109° 6' W	APR 1979
1	23° 48' S	165° 30' W	FEB 1983



Frequency (Hz)
Pressure power (Pascal²/Hz)
10⁻⁶
10⁻⁵
10⁻⁴
10⁻³
10⁻²
10⁻¹
10⁰
10¹
10⁹
10⁷
10⁵
10³
10¹
10⁻¹

Figure 1



Equivalent peak ground acceleration in 1/3 octave bandwidth (m/s²)
Frequency (Hz)
Mw=9.5
Mw=9.0 (Ms=8.7)
Mw=8.0 (Ms=8.0)
Mw=7.0 (Ms=6.9)
Mw=6.0 (Ms=5.5)
Mw=5.0 (Ms=4.0)
Mw=9.5
Mw=9
Mw=8
Mw=7
Mw=6
Mw=5

Figure 2

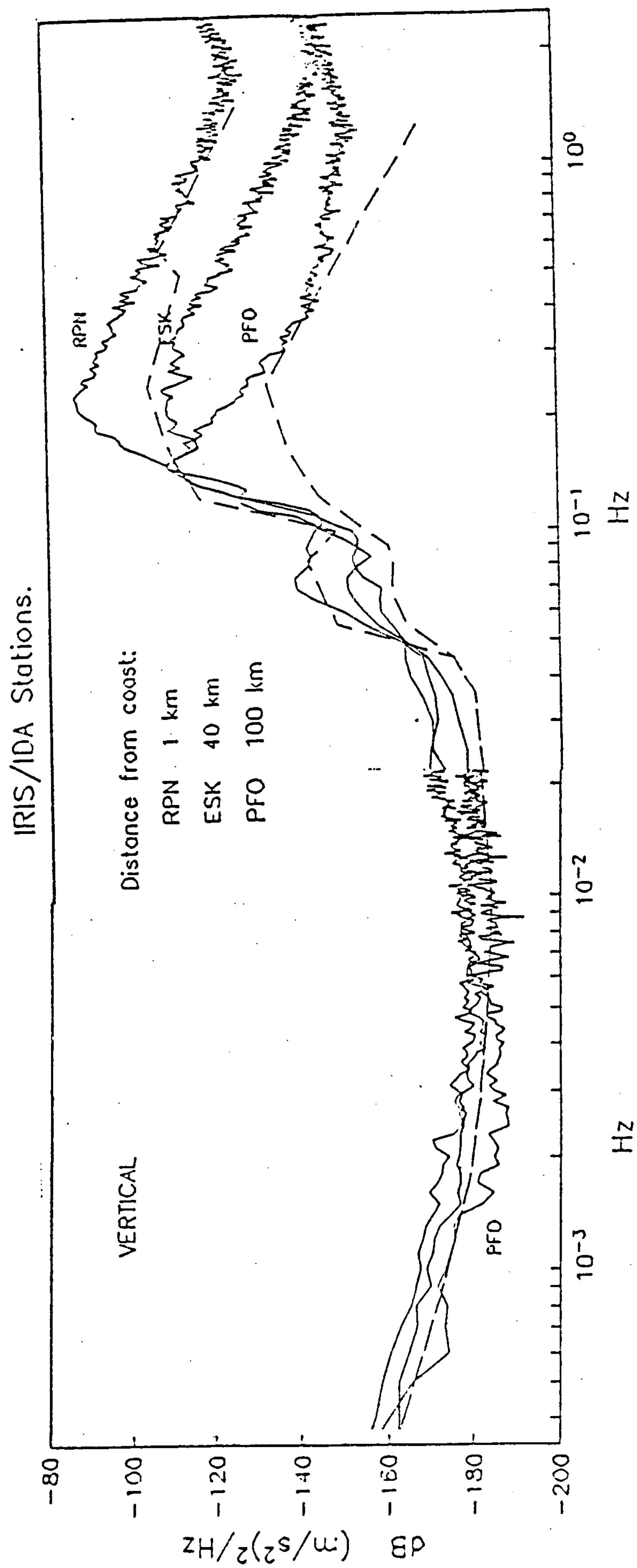

IRIS/IDA Stations.
Distance from coast:
RPN 1 km
ESK 40 km
PFO 100 km
VERTICAL
RPN
ESK
PFO
PFO
dB (m/s2)2/Hz
-80
-100
-120
-140
-160
-180
-200
10-3
10-2
10-1
100
Hz
Hz

Figure 3

A COMPARATIVE STUDY OF ISLAND, SEAFLOOR AND SUB-SEAFLOOR AMBIENT NOISE LEVELS

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ABSTRACT

A study of seafloor and island stations shows that for the frequency band 0.1 - 10 Hz, the seismic noise levels on islands are comparable to the levels on the seafloor. The microseism peak at the seafloor appears to be comparable to the highest levels observed on small islands. For this band, seafloor stations are realistic alternatives when island sites are not available.

Seven year averages of the ambient noise levels recorded by Seismic Research Observatory (SRO) stations on three islands (Guam [GUMO], Taiwan [TATO] and New Zealand's north island [SNZO] are compared to those recorded by the International Deployment of Accelerometers (IDA) station on Easter Island and on and beneath the ocean floor by Ocean Bottom Seismometers (OBS's) and the Marine Seismic System (MSS) deployed in a south Pacific DSDP drill hole at 23.8°S., 165.5°W (Adair et al, 1986). From 0.3 to 2 Hz the SRO displacement power levels fall in the range historically observed by the Scripps' OBS's (decreasing at 70 dB/decade from 1 x 10^6 nm^2/Hz at 0.3 Hz to 1 nm^2/Hz at 2 Hz) and are 10 to 15 dB above MSS levels. Above 2 Hz it appears that the same ratios hold (the SRO power levels decrease at 70 dB/decade to 1 x 10^{-3} nm^2/Hz at a frequency of 10 Hz), although this correlation is based on very limited, high gain, short period data. At frequencies below 0.3 Hz the SRO noise levels peak and decrease to approximately 2 x 10^3 nm^2/Hz at 40 mHz. The noise levels recorded at Easter Island are somewhat higher (decreasing at 70 dB/decade from 1 x 10^7 nm^2/Hz at 0.2 Hz to 1 nm^2/Hz at 10 Hz and to 1 x 10^5 nm^2/Hz at 50 mHz). At the microseism peak near 0.2 Hz the MSS levels are from 15 to 20 dB higher than observed by the SRO stations and equivalent to those recorded at Easter Island. There appears to be little dependence of the variance in noise level estimates on frequency. The upper 95% confidence limit generally lies 10 dB above the average noise levels for all island stations.

All island noise level curves are dominated by the broad double frequency microseism peak centered between 0.15 to 0.2 Hz. The single frequency peak ranges from absent (Easter Island) to discernable (Guam and New Zealand) to obvious (at Taiwan). The center frequency of this peak ranges from 0.07 Hz at Guam and New Zealand to 0.1 Hz at Taiwan. We speculate that the increased amplitude and frequency of the single frequency microseism peak is due to the interaction between the shallow continental shelf and surface gravity waves and/or the presence of Taiwan in a region of limited fetch.

INTRODUCTION

For the past 30 years, seismologists have deployed seismometers on and beneath the sea floor. The primary motivation for this work has been the desire to understand ocean crustal/lithospheric structure and seismicity. In addition, however, a long range goal of this research has been to use these instruments to improve the global coverage of seismometers. To demonstrate the feasibility of long term ocean bottom deployments it must be shown that noise levels at the seafloor are not significantly higher than those commonly observed on land.

This study is intended to assess the noise level change, if any, that can be expected by moving seismic sensors away from land. Toward this end this paper compares ambient noise levels on four islands with those at and below the sea-floor. In addition, the island noise level estimates are based on several noise observations spanning a number of years so that an estimate of the long term variance of these levels due to changing meterological and cultural conditions as a function of frequency will be obtained. The locations and relative magnitudes of the microseism peaks in the island spectra will also be discussed.

THE DATA

The SRO data used in this study were taken from day tapes of stations deployed on Guam, Taiwan and the north island of New Zealand, (GUMO, TATO and SNZO respectively, see table 1). Short and long period SRO data (with sampling rates of 20 and 1 sample per second respectively) have been merged to provide power estimates over the range from 40 mHz to 10 Hz. Only the vertical component data were studied since no horizontal short period SRO data are collected by the stations. Each power estimate is based on averaging 10 separate samples taken at random from 4 to 7 year time spans between 1976 and 1983 so that a wide range of meteorological conditions were incorporated. Prior to averaging, each trace was smoothed so that the final variance estimate would be largely due to changing meteorological conditions only.

Short period data are available only on event triggered records which are relatively infrequent. Each time series was examined on a visual display before transforming to ensure that the data corresponding to seismic arrivals were eliminated. Each long and short period sample consisted of 853 points padded to 1024 points with zeros. The time series were demeaned and then tapered with a Hanning window prior to Fourier transformation. The SRO seismometers are equipped with a notch filter centered at roughly 0.17 Hz which is designed to reduce the effect of the microseismic noise on the long period channels. The notch causes a sharp peak in the power spectrum following a correction for the instrument response. Since this peak is unlikely to be a real feature of the data, the portion of the power spectra between .15 and .19 Hz was removed and the plotted levels were obtained by interpolation. The power levels in this frequency band are clearly not valid and should be ignored. For a discussion of SRO instrumentation the reader is referred to Peterson et al (1976).

The IRIS/IDA data were collected by the vertical component of the Streckeisen seismometers deployed on Easter Island (station RPN, see table 1). These data were treated in the same manner as the SRO data although only 9 intermediate period (5 samples/s) and 4 short period (20 sample/s) samples, taken from June, 1987 to March, 1988, were included in the final noise level estimate. For a discussion of the instrument refer to Wielandt and Streckeisen (1982).

The MSS data were collected in February of 1983 during Scripps' Ngendei Expedition and the noise analysis was originally published by Adair et al (1986). The MSS was deployed at 23.8°S, 165.5°W in DSDP drill hole 595b (leg 91), 124 meters below the sediment-water interface and 54 meters into the basement. The MSS data were obtained from short (40 sample/s) and mid-period (4 sample/s) channels with time series which were roughly 13 and 128 seconds long respectively. To reduce the variance, the short and mid-period curves are based on the average of 10 spectra. The OBS noise level bounds were computed from 25 noise level estimates obtained from data collected at locations and dates detailed in table 2. Each individual estimate was made after stacking from 3 to 125 noise level curves although the individual time series collected were too short to obtain high resolution at frequencies approaching 0.2 Hz. Both MSS and OBS power estimates have been corrected for instrument response. The MSS and OBS instruments are discussed in detail in Adair et al (1986) and Orcutt et al (1987).

COMPARISON OF NOISE LEVELS

The four island stations exhibit similar time-averaged noise structure (figures 1, 2, 3 and 4). All records are dominated by the double frequency microseism peak centered between 0.15 and 0.2 Hz. The single frequency peak ranges from absent (Easter Island) to discernable (Guam and New Zealand) to obvious (at Taiwan). The center frequency of this peak ranges from 0.07 Hz at Guam and New Zealand to 0.1 Hz at Taiwan. At frequencies above the main microseism peak, all four spectra fall off at roughly 70 dB/decade. In all spectra the rate of decay decreases at frequencies above 2 Hz and the SRO station power levels actually increase above 5 Hz. It is felt that this apparent "leveling off" of the power spectra is not a real feature of the data but is due to digitization and instrumentation noise. The three SRO stations used in this report were selected because they are located on islands and thus are close to the shoreline. To combat the resultant high noise levels, the SRO short period station gains were reduced to a level 60 dB lower than those of all other SRO stations, (Peterson et al, 1980). As a result, these records are more susceptible to digitization and instrumentation noise at high frequencies. In the early days of 1976 at GUMO, however, the short period data were collected with high gain. The spectrum from 12 January 1976 is shown in figure 5. At frequencies above 2 Hz the power spectral levels continue to decay at roughly 70 dB/decade. This spectrum is clearly more representative of the true power level in this frequency band. Because the average long term power levels at all stations were desired, it was obviously not possible to base the average power level estimates solely on these high gain data.

At frequencies above 0.2 to 0.3 Hz the island power levels fall within the OBS noise level bounds and are 10 to 15 dB above MSS levels. At frequencies below 0.2 Hz the MSS levels are comparable to those recorded at Easter Island and 10 to 15 dB above noise levels recorded on the 3 SRO islands. This enhanced island noise level is probably a reflection of the oceanic origin of the microseisms. The upper 95% confidence limit, lying two standard deviations above the average, is plotted on figures 1 to 4 with the average power level curve. The lower confidence limit was excluded since it was negative in places. These curves track each other fairly well and indicate that each frequency band has approximately the same variance.

THE ISLAND MICROSEISM PEAKS

The traditional explanation of the origin of the double frequency microseism peak involves the non-linear interference of gravity waves traveling in opposite directions, Longuet-Higgins (1950). In deep water this non-linear interaction is effective in producing acoustic energy at twice the frequency of the interacting gravity waves. This acoustic energy subsequently excites microseismic activity in the basement rock at the same frequency. The explanation of the single frequency microseism peak is fundamentally different. Early observations of this peak were made by Wiechert (1904), Oliver and Ewing (1957) and Haubrich et al (1963). The generally accepted explanation holds that when surface gravity waves propagate through shallow water the decaying pressure oscillations they cause are "felt" directly by the substrate. These pressure oscillations thus excite seismic energy at the same frequency as the source surface gravity waves. A generally accepted rule of thumb holds that this interaction is significant when the water depth is less than 60% of the wavelength of the surface gravity waves. The pressure fluctuations caused by such a wave on the bottom is greater than 4.6% of the static water pressure. At 0.07 Hz a typical wavelength is on the order of 350 m. This crude calculation suggests that in water shallower than about 200 m, single frequency microseisms can be excited.

Adair (1985) discussed previous continental microseismic studies which have generally detected the small, single frequency microseism peak at 0.07 Hz and the larger amplitude, double frequency, peak at 0.14 Hz. The single frequency microseism peaks in the GUMO and SNZO spectra agree with each other and conform to these continental studies. However, at TATO the fundamental microseism peak is more apparent and both peaks is more apparent and both peaks are shifted to higher frequencies (0.1 and 0.2 Hz). The single frequency peak is absent from the Easter Island data.

Plausible explanations for these differences may be found by considering the bathymetric and geographic settings of the four islands. Taiwan is located on the continental shelf of China and is 150 km from the mainland. There is a broad expanse of water less than 200 m deep in the vicinity of this island (the shallowest water, generally less than 100 m deep, lies to the north and west). The north island of New Zealand, Guam and Eastern Island, however, are located far from major land masses. The local water depths increase rapidly away from the shore to greater than 200 m so that the generative area for primary microseisms is small (relative to Taiwan). This is especially true at Guam and Easter Island. It seems likely that the increased amplitude of the primary microseismic peak at TATO relative to GUMO, SNZO and especially RPN is due to the broad expanse of shallow water surrounding Taiwan which facilitates the direct conversion of gravity wave pressure fluctuations to seismic energy in the basement.

Hasselmann and Collins (1967) discussed the influence of shallow water on the frequency content of surface gravity waves. As these waves travel in shallow water, the long period components that interact directly with the sea floor are preferentially dissipated. In the vicinity of Taiwan where the surface gravity waves most likely have traveled a great distance over the shallow basement before approaching the island it seems that this mechanism would strip away the low frequencies preferentially and cause the apparent shift of the single frequency microseism peak to a slightly higher frequency.)

It is widely known that the frequency (f_m) of the main energy peak in the spectrum of surface gravity waves depends on the fetch and the local wind speed (commonly represented by U_{10}, the velocity of wind 10 meters above the sea surface). Hasselmann et al (1973) found that f_m was proportional to $(U_{10})^{-34}$ in the North Sea, a region of limited fetch. In theory fm is proportional to $(U_{10})^{-1.0}$ in unlimited fetch. The waves grow until the phase velocity, which is inversely proportional to the frequency, equals that of the wind. The proximity of Taiwan to mainland China restricts the local wind fetch and thus could plausibly explain the increased frequency of the single frequency microseism peak. The dependence of f_m on the wind speed itself could provide an alternative explanation for the shift in Taiwan's single frequency microseism peak provided that the long term average wind speed was significantly lower in the vicinity of Taiwan than near the other islands. However, wind speed data published by the U.S. Naval Weather service indicate little difference between the average scalar wind speeds at these locations. At Taiwan, Guam, New Zealand and Easter Island the average wind speeds are 14.0, 12.7, 14.5 and 13.0 knots.

CONCLUSION

We have computed and compared estimates of noise levels on four islands, on the seafloor and beneath the seafloor. Over a limited frequency band the ocean bottom and island power levels are comparable and thus no degradation of the signal to noise ratio should be expected by moving sensors from land to the seafloor. We have shown that over this frequency band a sensor located beneath the ocean bottom will display significantly lower noise levels than seismometers deployed on the ocean bottom or on islands. This is true except at the lowest frequencies (below 0.2 Hz). These observations are particularily encouraging with regard to seafloor stations over the bulk of the Pacific and Indian oceans where few islands are available for station siting.

It seems plausible that the increased amplitude and frequency of the single frequency microseismic peak in the TATO record relative to GUMO, SNZO and RPN is due to the interaction between the shallow continental shelf surrounding Taiwan and surface gravity waves and/or the presence of Taiwan in a region of limited fetch.

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TABLES AND FIGURE CAPTIONS

Table 1: The geographic coordinates of the four island-based seismometers incorporated in this study. The areas of the islands on which these seismometers are deployed are included.

Table 2: The geographic coordinates and dates at which the OBS noise samples used in this study were collected.

Figure 1: Noise power spectral density in displacement units for the SRO station on Guam. The lower solid curve is the average noise level obtained from seven years of data while the upper curve is a 95% confidence limit on the power. The frequently dashed curve is the power spectral density from the Marine Seismic System deployed during the southwest Pacific Ngendei Expedition. The broadly dashed bounds represent upper and lower limits of noise levels observed by the Scripps' OBSs over a twelve year period in the Pacific Basin.

Figure 2: Noise power spectral density in displacement units for the SRO station on the north island of New Zealand. The other curves are explained in the caption for Figure 1.

Figure 3: Noise power spectral density in displacement units for the SRO station on Taiwan. The other curves are explained in the caption for Figure 1.

Figure 4: Noise power spectral density in displacement units for the IRIS/IDA station on Easter Island. The other curves are explained in the caption for Figure 1.

Figure 5: Power spectral levels for displacement for GUMO for records in 1976 when the SRO station was operating at higher gain. The other curves are explained in the caption for Figure 1. Note that the noise levels continue to decay at higher frequencies indicating that the higher noise levels observed at this and the other island stations are probably associated with digitization noise.

Table 1 Island Data

SITE	LATITUDE	LONGITUDE	ISLAND AREA (Square Km)
GUMO	13° 35' 17" N	144° 51' 58" W	541
SNZO	41° 18' 36" S	174° 42' 16" W	115,000
TATO	24° 58' 32" N	121° 29' 18" W	36,000
RPN	27° 08' 37" S	109° 26' 10" W	118

Table 2 OBS data

NUMBER OF SAMPLES	LATITUDE	LONGITUDE	DATE
6	41° 30' N	127° 20' W	JUN 1976
2	16° 0' N	145° 0' W	OCT 1977
1	29° 30' N	122° 0' W	OCT 1977
3	18° 0' N	145° 20' W	JAN 1976
8	16° 30' N	100° 30' W	JUN 1977
4	20° 50' N	109° 6' W	APR 1979
1	23° 48' S	165° 30' W	FEB 1983

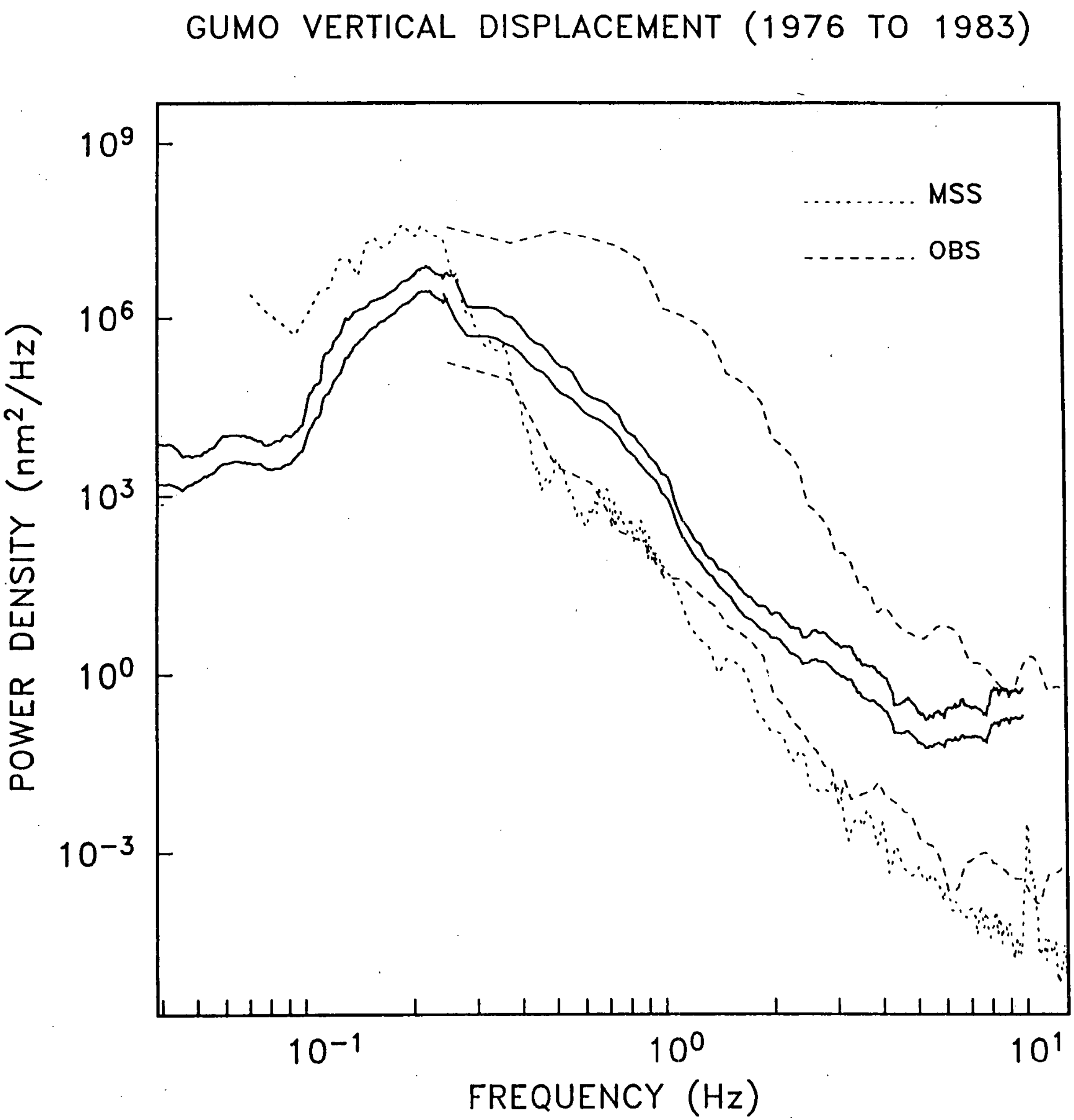
GUMO VERTICAL DISPLACEMENT (1976 TO 1983)
MSS
OBS
POWER DENSITY (nm²/Hz)
10⁹
10⁶
10³
10⁰
10⁻³
10⁻¹
10⁰
10¹
FREQUENCY (Hz)

Figure 1

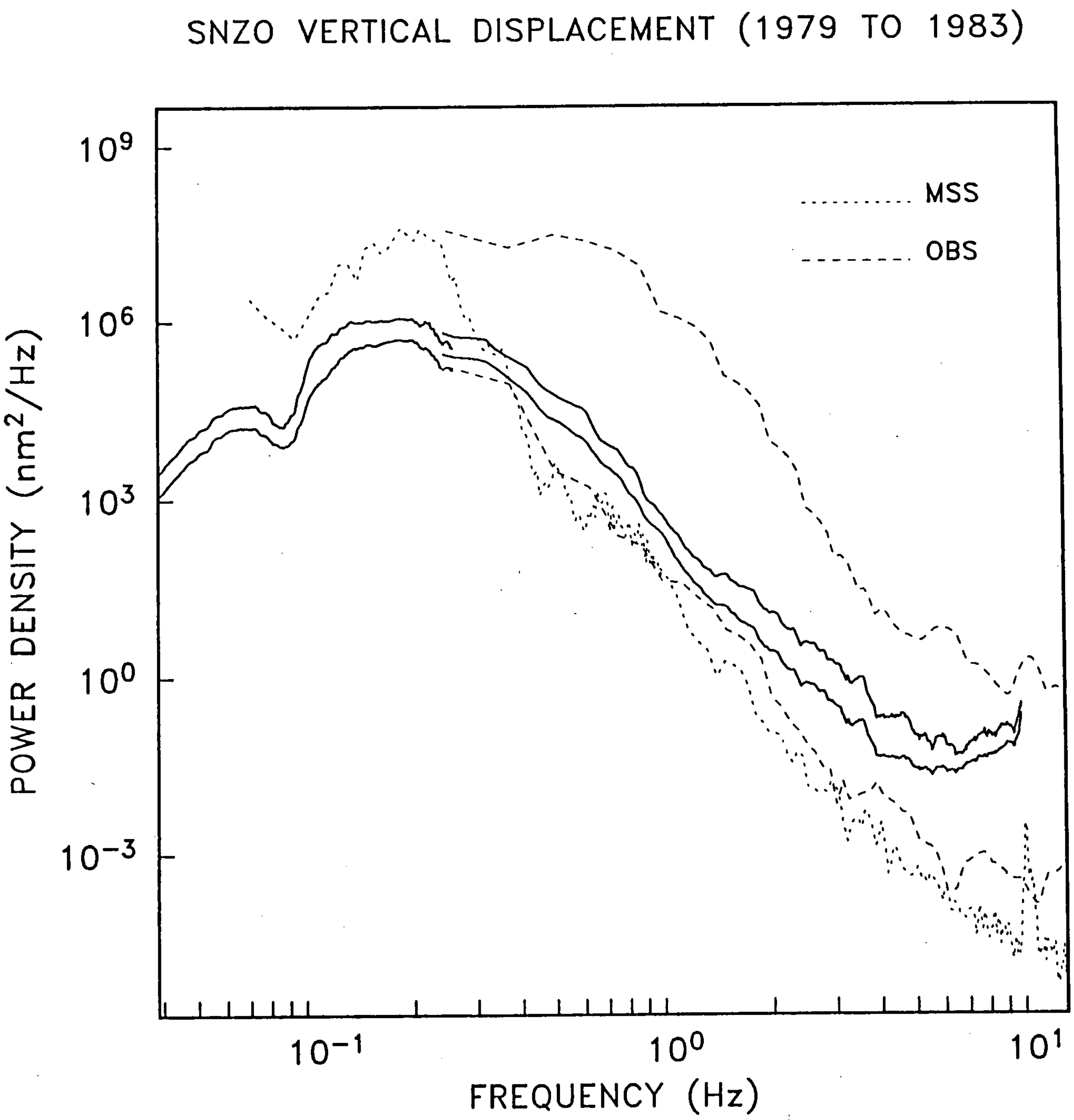
SNZO VERTICAL DISPLACEMENT (1979 TO 1983)
MSS
OBS
POWER DENSITY (nm²/Hz)
10⁹
10⁶
10³
10⁰
10⁻³
10⁻¹
10⁰
10¹
FREQUENCY (Hz)

Figure 2

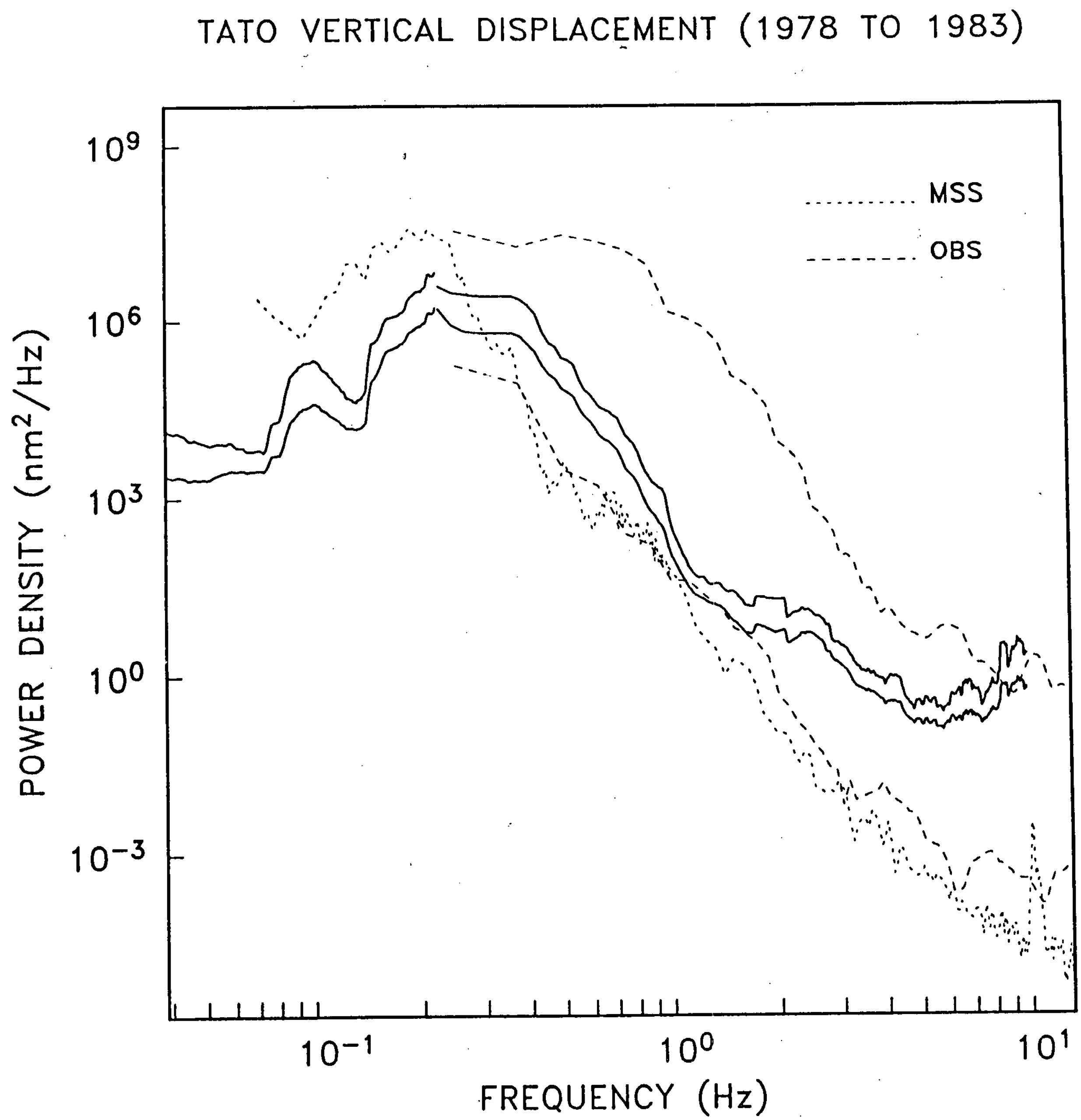
TATO VERTICAL DISPLACEMENT (1978 TO 1983)
MSS
OBS
POWER DENSITY (nm²/Hz)
10⁹
10⁶
10³
10⁰
10⁻³
10⁻¹
10⁰
10¹
FREQUENCY (Hz)

Figure 3

RPN VERTICAL DISPLACEMENT (JUNE 1987 TO MARCH 1988)

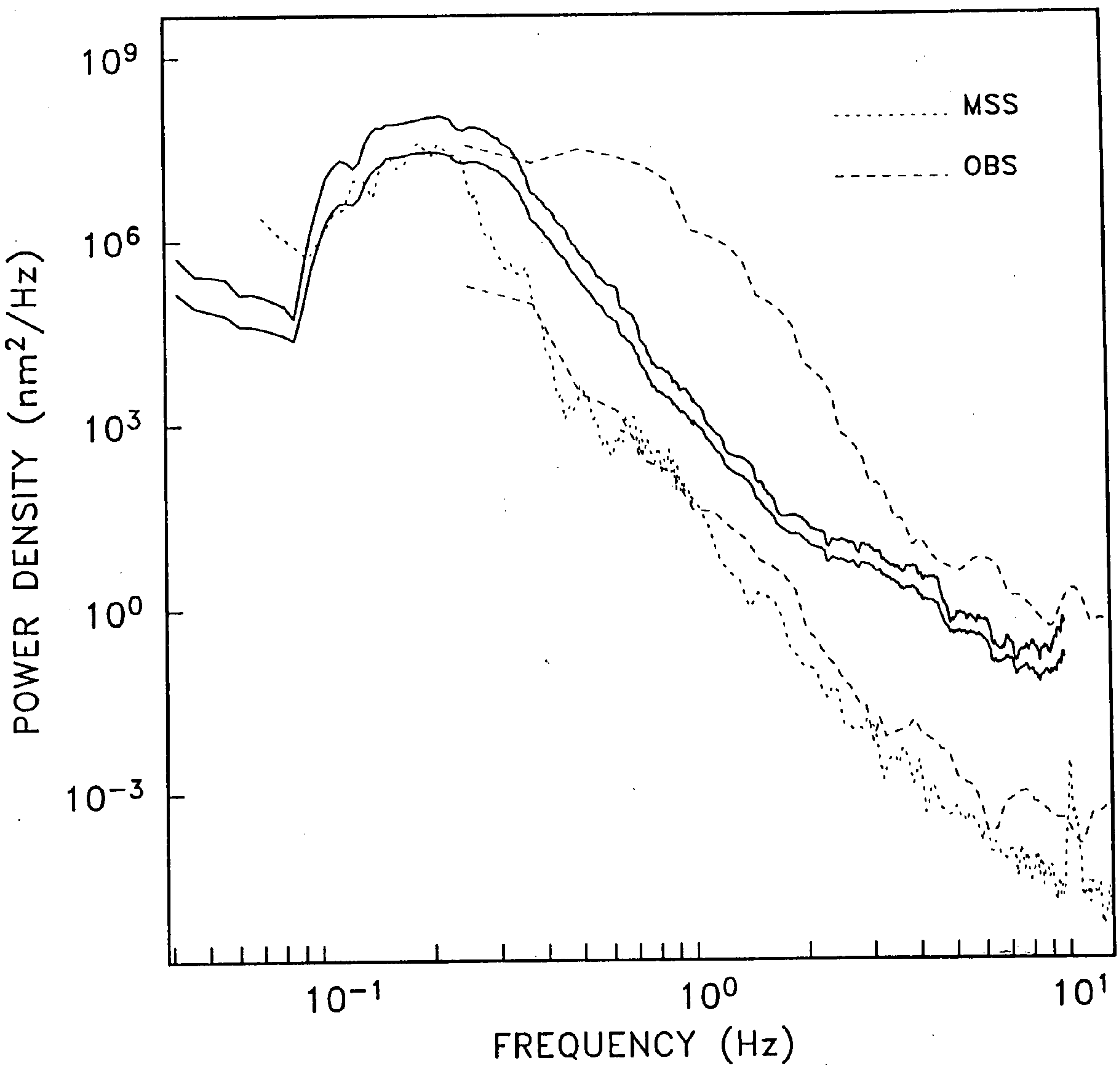

Figure 4

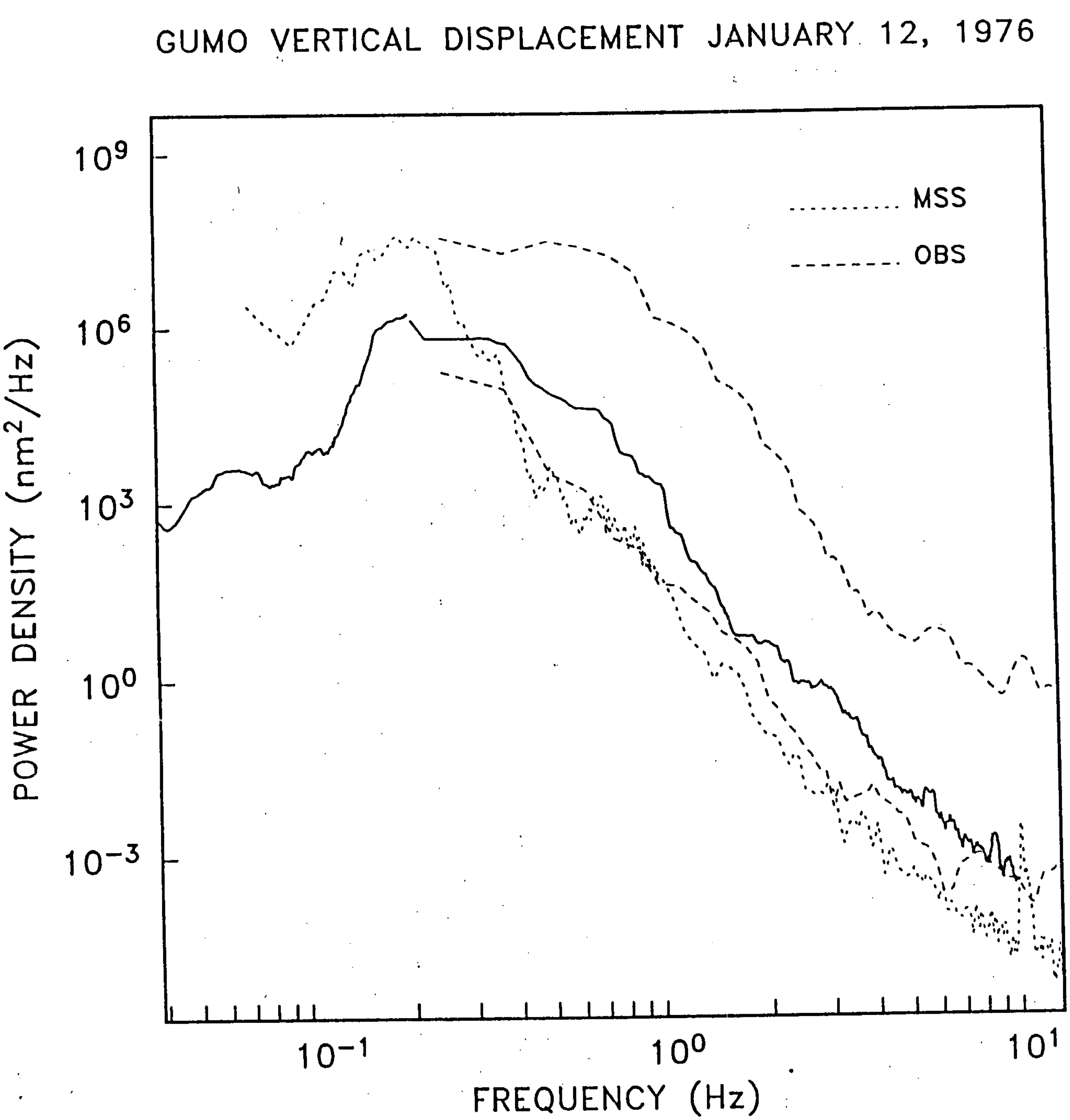
GUMO VERTICAL DISPLACEMENT JANUARY 12, 1976
MSS
OBS
10^9
10^6
10^3
10^0
10^-3
POWER DENSITY (nm^2/Hz)
10^-1
10^0
10^1
FREQUENCY (Hz)

Figure 5

Section B4: Marine Downhole Seismic Experiments

1. Borehole Seismic Experiments and the Structure of Upper Ocean Crust: R.A. Stephen
2. Ambient Noise as a Function of Depth in the Seafloor: R.A. Stephen and J.A. Orcutt
3. Low Frequency Acoustic Seismic Experiment (LFASE): R.A. Stephen, J.A. Orcutt, H. Berteaux, D. Koelsch and R. Turpening
4. Downhole Seismometer Experiment in the Sea of Japan: K. Suyehiro
5. Long-term Deep-Ocean Borehole Seismology - Is it worth the Cost?: F. Duennebier
6. Signal Behavior in Boreholes: G.J. Tango
7. Downhole and Seafloor Seismic Measurements made by the Hawaii Institute of Geophysics: Past, Present, and Future: F. Duennebier and C. McCreery

BOREHOLE SEISMIC EXPERIMENTS AND THE STRUCTURE OF UPPER OCEAN CRUST

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Woods Hole, MA 02543

Introduction

During the past eleven years it has been possible to carry out seismic experiments in the deep sea with clamped borehole geophones (Figure 1). Ten oblique seismic experiments were carried out from the D/V GLOMAR CHALLENGER with the assistance of 'shooting ships' which fired small explosive charges and airguns out to ranges of 12 km or more (Figure 2, Table 1). One oblique seismic experiment and two normal incidence VSPs have been carried out on the JOIDES Resolution. The first experiment was carried out in the Western Atlantic Ocean south of Bermuda in 1977. Excellent quality three component data were collected at two depths in the hole from shots out to 12 km range at four azimuths. Traveltime, synthetic seismogram and particle motion analyses were applied to the data. In the second experiment in the Gulf of California only vertical component data were obtained and traveltimes and amplitudes were analyzed. Three experiments have been carried out in the Costa Rica Rift area (DSDP Hole 504B).

The original objective of the borehole experiments were established on the assumption of two kilometers penetration into oceanic crust. The objectives were 1) to determine the lateral extend of the velocity structure intersected by the borehole, 2) to analyze the role of fissures and large cracks (greater than centimeter size) in the velocity structure of ocean crust, 3) to look for seismic anisotropy in upper oceanic crust and 4) to obtain a measurement of attenuation. By comparing the refraction velocities from the borehole experiment to sonic log velocities and laboratory sonic measurements on recovered core material, an estimate of the large-scale porosity can be made. Seismic anisotropy can be studied effectively by particle motion analysis of three component data as shown by the site 504B data. The observed anisotropy is attributed to the preferred orientation of large scale fissures and faults. Attenuation has remained an elusive objective because of the small penetration into into basement material.

Although not originally an objective, the borehole seismic results have proved useful in determining shallow basement velocity structure. The structure of the upper 500 m of basement is poorly resolved using ocean bottom or ocean surface receivers. Unfortunately, travel time inversion schemes require an estimate of the uppermost velocity in a profile. This can be measured directly at in situ conditions and seismic frequencies from borehole receiver data and the velocity-depth profile in the uppermost crust is then considerably better resolved. Traveltime data can be inverted using the zeta tau inversion method modified for borehole receivers. Amplitude analysis for borehole receivers by trial and error fitting of synthetic seismograms generated by the reflectivity and finite difference methods has also been carried out.

Offset VSPs and Seismic Anisotropy in the Upper Oceanic Crust

Over the past ten years there have been four reported observations of seismic azimuthal anisotropy in the upper 2 km of ocean crust. Three of these observations have been based on offset VSP data. The most probable cause of anisotropy in the upper crust is the preferred orientation of large scale fractures which are created during crustal

Leg	52	65	65	67	70	78A
Hole	417D	482C	485A	494A	504B	543A
Date	March 1977	February 1979	March 1979	May 1979	December 1979	March 1981
Coordinates	25°06.69'N, 68°02.82'W	22°47.34'N, 107°59.57'W	22°44.92'N, 107°54.23'W	12°43.01'N, 90°55.97'W	1°13.6'N, 83°43.8'W	15°42.74'N, 58°39.22'W
Location	Bermuda Rise	Gulf of California	Gulf of California	Middle America Trench	Costa Rica Rift	Barbados Ridge
Age of crust	110 my	0.5 my	1 my	Upper Cretaceous?	6 my	75 my
Water depth	5479 m	2998 m	2981 m	5472 m	3462 m	5633 m
Sediment thickness	343 m	137 m	153 m	322 m	275 m	411 m
1. Geophone depth	8 m	175 m	71 m	297 m	52 m	44 m?
Azimuths of shot lines	22.5°	none	30°, 15°, 75°, 60°	?	0°, 90°, 180°, 270°	none
Number of shots	45	none	220	?	96	none
Radii of shot circles	none	none	none	?	none	none
Number of shots	none	none	none	?	none	none
2. Geophone depth	228 m				542 m	
Azimuths of shot lines	22.5°, 67.5°				0°, 90°, 180°, 270°	
Number of shots	120				79	
Radii of shot circles	none				none	
Number of shots	none				none	
3. Geophone depth						
Azimuths of shot lines						
Number of shots						
Radii of shot circles						
Number of shots						
4. Geophone depth						
Azimuths of shot lines						
Number of shots						
Radii of shot circles						
Number of shots						
5. Geophone depth						
Azimuths of shot lines						
Number of shots						
Radii of shot circles						
Number of shots						
Ocean bottom hydrophones	none	1	none	5	3	5
Ocean bottom seismometers	none	1	none	5	none	5
Basement topography	unknown	unknown	known	unknown	unknown	known
Spreading direction	22.5°	125°	30°	45°?	0°	110°
Shear wave analysis	yes		no		no	
Amplitude analysis	yes		yes		no	
Shooting ship	*Virginia Key*		*Kana Keoki*	*Kana Keoki*	*Gillis*	*North Star*
Institution	Cambridge	HIG	WHOI	HIG	WHOI	HIG

Table 1(a) Summary of borehole seismic work carried out on DSDP and ODP programs. Normal incidence VSP's have not been included.

Leg	78B	88	91	92	102
Hole	395A	581	595B	504B	418A
Date	March 1981	September 1982	February 1983	April 1983	April 1985
Coordinates	22°45.35'N, 46°04.90'W	43°55.44'N, 159°47.84'E	23°49.34'S, 161°31.61'W	1°13.6'N, 83°43.8'E	25°02.10'S, 68°03.44'W
Location	Mid-Atlantic Ridge	northwest Pacific	southwest Pacific	Costa Rica Rift	Bermuda Rise
Age of crust	7.2 my	115 my	140 my	6 my	110 my
Water depth	4483 m	5467 m	5701 m	3462 m	5511 m
Sediment thickness	53 m	356 m	70 m	275 m	324 m
1. Geophone depth	556 m	22 m	54 m	42 m	430 m
Azimuths of shot lines	55°, 225°, 280°	70°, 165°, 260°, 345°	135°, 225°, 315°	0°, 45°, 90°, 135°, 180°, 225°, 270°, 315°	20°, 60°, 70°, 155°, 200°, 240°, 290°, 335°
Number of shots	112	385	250	100	315
Radii of shot circles	none	none	none	2, 4, 6, 8 km	2, 4, 6, 8 km
Number of shots	none	none	none	218	380
2. Geophone depth				272 m	330 m
Azimuths of shot lines				0°, 5°, 90°, 135° 180°, 225°, 270°, 315°	20°, 65°, 110°, 155°, 200°, 240°, 335°
Number of shots				111	313
Radii of shot circles				2, 4, 6, 8 km	2, 4, 6, 8 km
Number of shots				81	350
3. Geophone depth				452 m	230 m
Azimuths of shot lines				none	20°, 65°, 110°, 155°, 200°, 245°, 290°, 340°
Number of shots				none	310
Radii of shot circles				6 km	2, 4, 6, 8 km
Number of shots				66	374
4. Geophone depth				667 m	81 m
Azimuths of shot lines				0°, 90°, 180°, 270°	20°, 110°, 200°, 290°
Number of shots				56	118
Radii of shot circles				6 km	4, 6 km
Number of shots				64	115
5. Geophone depth					41 m
Azimuths of shot lines					20°, 200°
Number of shots					28
Radii of shot circles					6 km
Number of shots					11
Ocean bottom hydrophones	none	5	6	none	none
Ocean bottom seismometers	4	10	6	none	none
Basement topography	known	unknown	known	known	known
Spreading direction	90°	?	30°	0°	115°
Shear wave analysis	no	no	yes	yes	no
Amplitude analysis	no	yes	yes	yes	yes
Shooting ship	*Lynch*	Steiger, Dimitri Mendeleev	*Melville*	*Ellen B. Scripps*	*Fred Moore*
Institution	Scripps	HIG	Scripps	WHOI	WHOI

Table 1(b) Summary of borehole seismic work carried out on DSDP and ODP programs. Normal incidence VSP's have not been included.

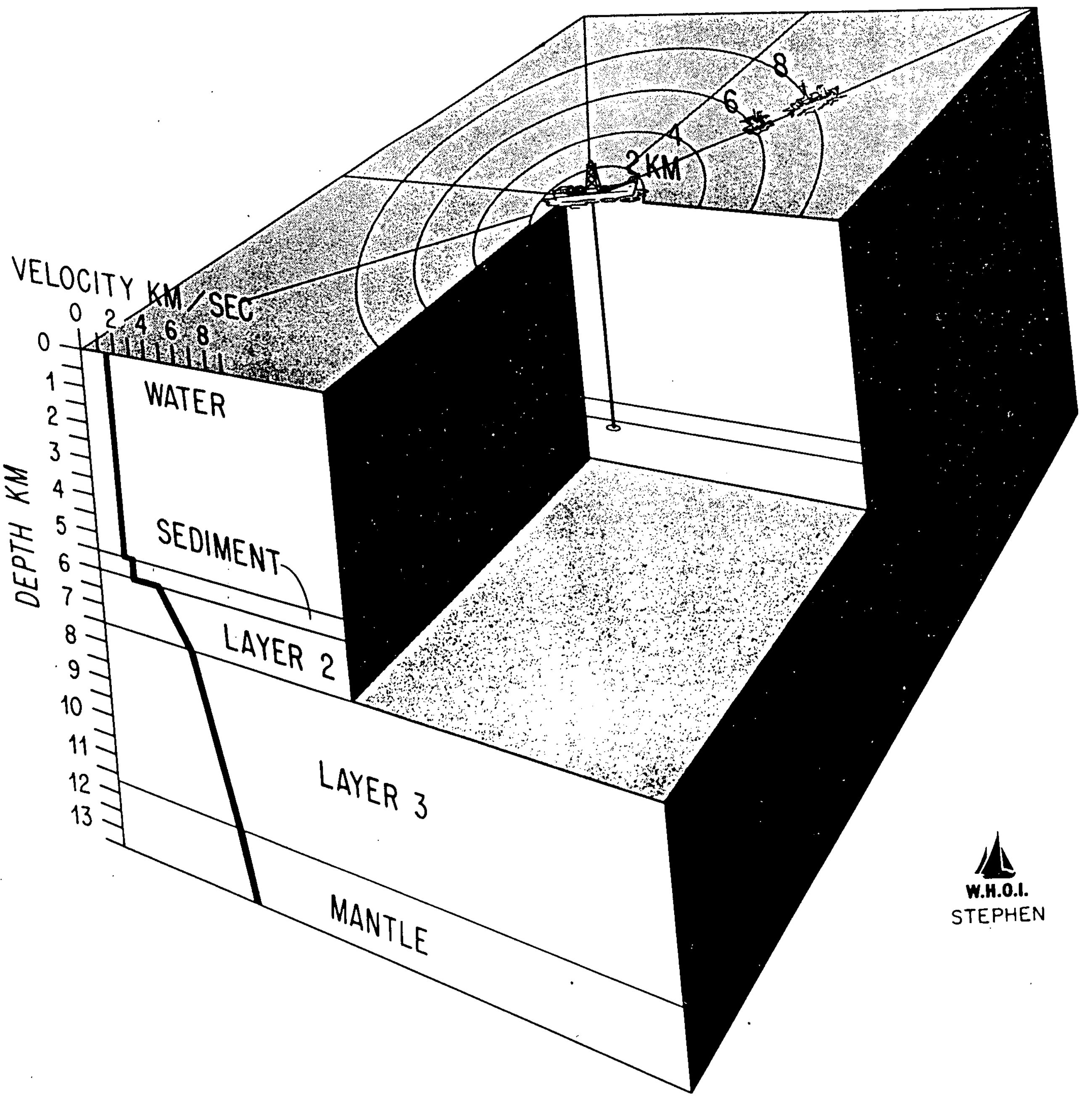

Figure 1. Schematic diagram of an offset VSP (or oblique seismic experiment). The velocity-depth function for a typical deep water site is also shown.

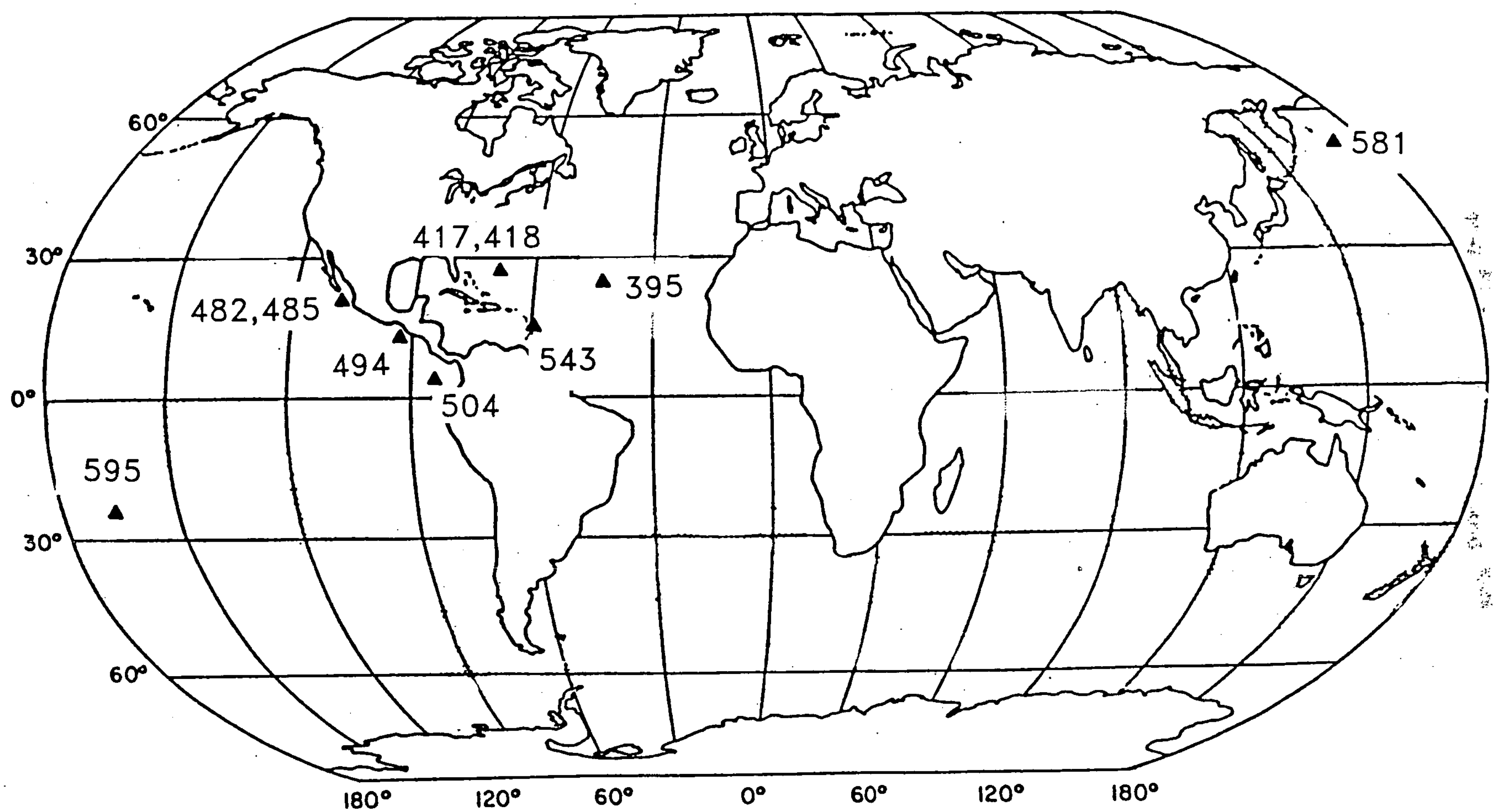

Figure 2. Location of DSDP and ODP holes in which previous borehole seismic work has been carried out.

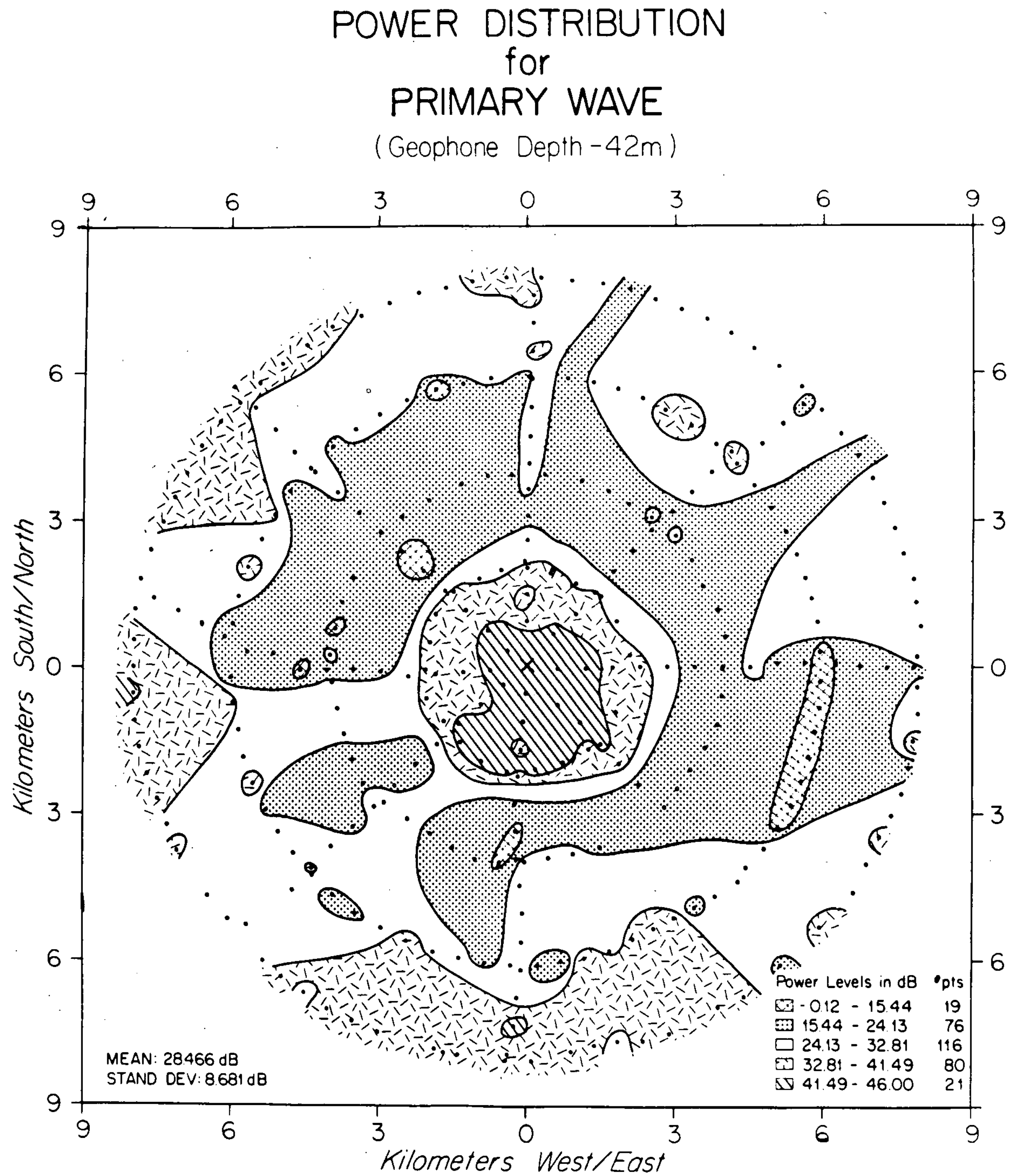

Figure 3. Power distribution plot for primary wave energy from an offset VSP experiment at DSDP site 504. The borehole receiver is at the origin and the dots indicate shot locations. If the seafloor were flat, isotropic and laterally homogeneous, this plot would consist of concentric circles. Deviations from a concentric pattern (greater than 20 db in some instances) indicate that these frequently made assumptions are invalid at the site. Synthetic seismogram modelling based on the observed bathymetry at the site shows that significant lateral heterogeneity (in addition to the bathymetry) is necessary to explain the anomalies.

generation at the ocean ridges. Seismic velocities are slow parallel to the spreading direction and fast perpendicular to it. In some instances clear azimuthally dependent traveltime anomalies are observed and, in others, there is evidence for shear wave splitting. At one site, both shear wave splitting and traveltime anomalies gave consistent results. The challenge in all studies, however, is to separate the effects of anisotropy from the effects of scattering and lateral heterogeneity. Current modelling capability is inadequate to include the effects of both anisotropy and heterogeneity at the scales observed in the seafloor. We need a definitive study on the effects of anisotropy and heterogeneity which will determine which observations are best for resolving given parameters. In addition, high quality three component seismic stations are required to provide the data for particle motion analysis.

Offset VSPs and the Lateral Variability of Ocean Crust

The lateral variability of the upper crust can be studied by mapping the amplitude (or power) content of arrivals from a dense array of identical sources to a single receiver (offset VSP). The results from a borehole seismic experiment at Deep Sea Drilling Project (DSDP) site 504 in the eastern equatorial Pacific show significant lateral variability. Seismic amplitude anomalies of 20 db and more are present with a horizontal correlation length of 1-3 km (Figure 3). Finite difference synthetic seismogram techniques can be used to predict seismic wave propagation through realistic bathymetry, basement relief and lateral variations within basement and a three dimensional model of the velocity structure around the borehole can be constructed. The resultant model has horizontal gradients greater than 2.0 sec^{-1} in the upper crust. These gradients are comparable to the vertical gradients normally associated with oceanic crust. Assuming that velocity anomalies in the upper crust reflect changes in porosity, the resultant structure provides constraints on the hydrothermal circulation at the site.

AMBIENT NOISE AS A FUNCTION OF DEPTH IN THE SEAFLOOR

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Woods Hole Oceanographic Institution*, Scripps Institution of Oceanography**

The Ngendie experiment in the southwest Pacific was a borehole seismic experiment aimed at comparing noise levels between a seismometer clamped in basement and a seismometer on the seafloor. Signal to noise ratios for both earthquakes and shots are better on the borehole seismometer by as much as 10 dB in the frequency range from 5 to 20 Hz. The improvement was considerably more significant on the horizontal components. The observations are consistent with a model in which the seafloor noise propagates as interface waves on the ocean bottom. Signal to noise gains can be expected if one places receivers below the skin depth for the interface phases.

Clearly, a seafloor seismic installation aimed at monitoring earthquake activity will benefit from being located in a borehole away from the noise interface. However questions still remain. At what depth does optimum or acceptable signal-to-noise ratio occur? What are the specific mechanisms of seafloor noise generation and propagation? Does the improve noise level warrant the additional expense of a borehole installation? What are the lateral variations in noise level on the seafloor? What is the effect of regional geologic environment and the extremely local geology at the receiver? How do noise levels vary as a function of frequency (50 Hz to 1000 sec.) below the seafloor? How does ambient noise vary between vertical and horizontal seismometers and pressure sensors? How does noise depend on sea state and proximity to shorelines, tectonic regimes, and sea lanes? These questions can only be answered by further, careful borehole seismic experiments in the seafloor.

LOW FREQUENCY ACOUSTIC SEISMIC EXPERIMENT (LFASE)

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Woods Hole Oceanographic Institution*, Scripps Institution of Oceanography**,
Massachusetts Institute of Technology***

The objective of the LFASE project is to understand the physics of excitation and propagation of low frequency noise (2-50 Hz) immediately above, at, and below the seafloor. The experiment is planned to be carried out in DSDP Hole 418 in the summer of 1989.

The experiment will first deploy an array of ocean bottom seismometers and a vertical hydrophone array near the borehole. Then the borehole array will be deployed using a borehole re-entry guide and thruster package (Figure 1). Airgun and explosive shots will be fired in a pattern around the borehole by a second ship while the re-entry ship stays attached to the borehole array and records data on board. After this phase, the thruster will release from the bottom package and the bottom unit will autonomously record array data in preprogrammed windows for up to a month (Figure 2). The borehole array will then be recovered by grappling from the thruster package.

The LFASE borehole seismic array will be a modified version of a commercially available array sold by Compagnie Generale Geophysique. The array has four nodes with variable separation between nodes of from 10 to 100 m. Each node consists of a three component seismometer and can be clamped in holes from 7" to 16" diameter. There is the option of putting a hydrophone in the top node at the expense of one horizontal geophone channel in the lower nodes. The geophones are Mark Products L15-B's with 4.5 Hz natural frequency and 60 percent clamping. The hydrophone is an Ocean and Atmospheric Science, Inc. E2-SD. The preamplifier gain is 66 dB for each geophone channel and 54 dB for the hydrophone channel. The filters have a pass band between 2.5 and 40 Hz with a low cut roll off of 12 dB/octave and a high cut roll off of 60 dB/octave. The 12 channels are digitized at a 2 msec sampling rate with gain ranging in 11 steps at 6 dB/step and an 11 bit (plus sign) analog to digital converter.

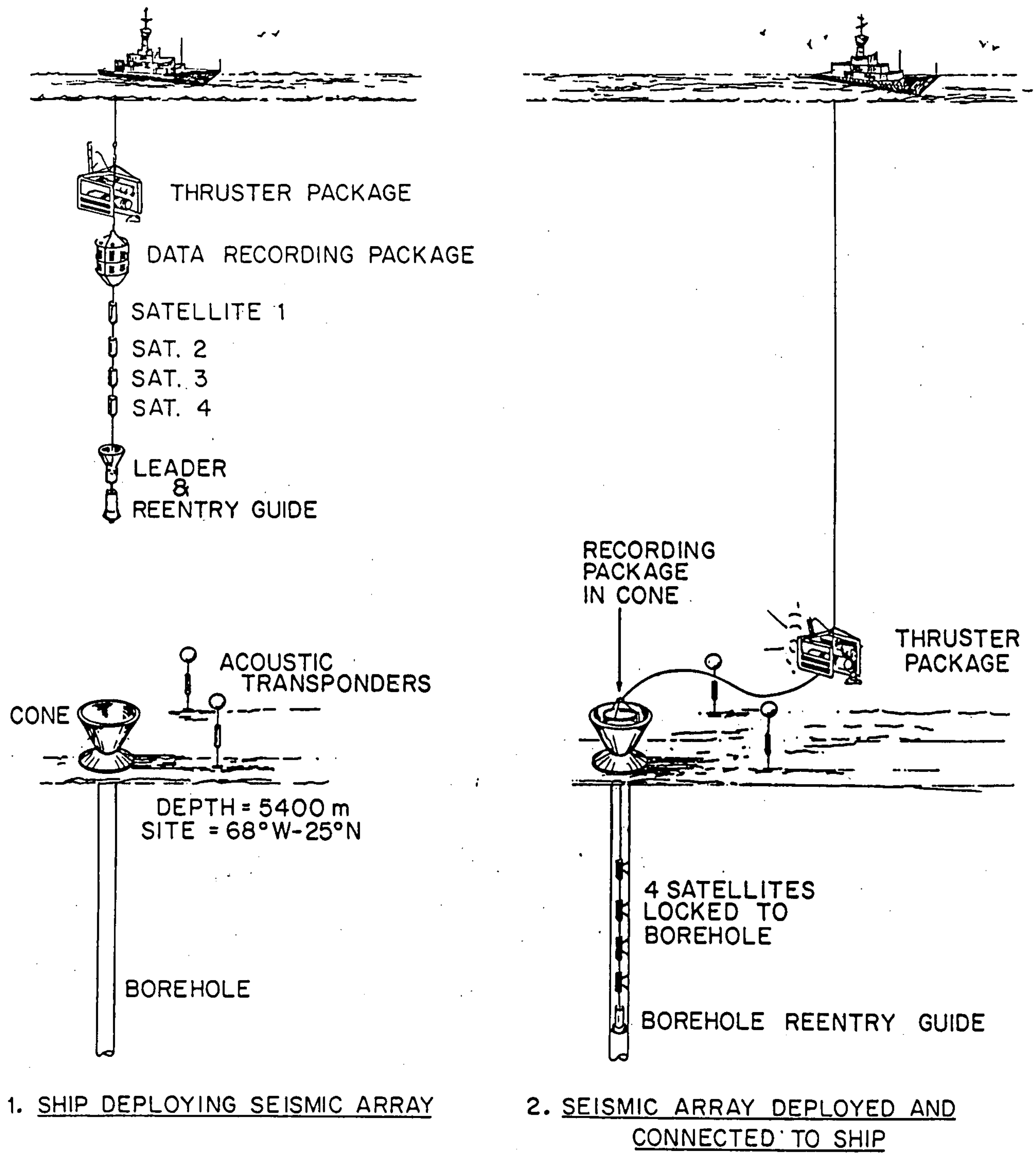

Figure 1. Deployment scenario for the borehole seismic array in LFASE.

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BERTEAUX

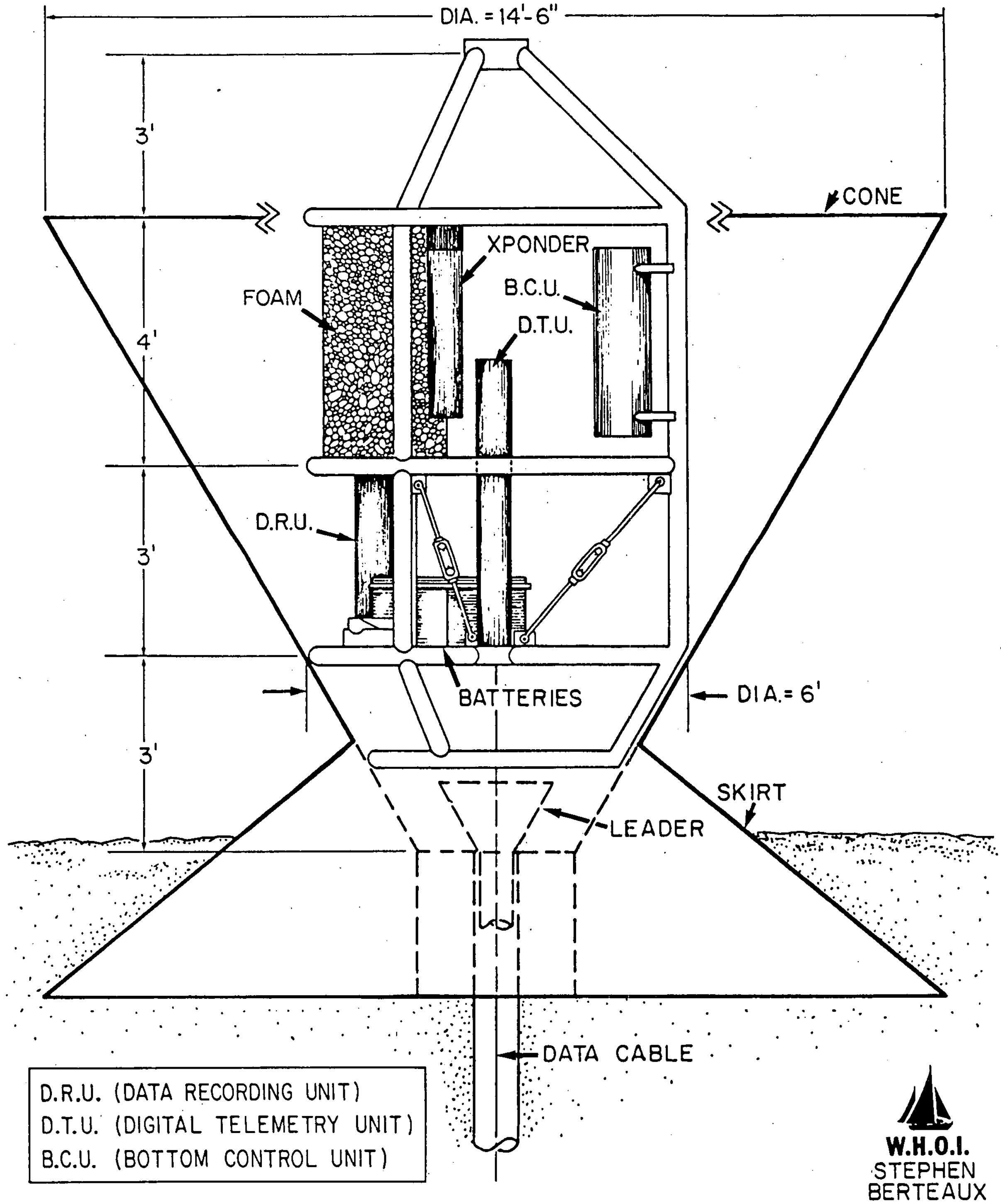

Figure 2. Schematic diagram of the LFASE Bottom Package in the DSDP Re-entry Cone.

DOWNHOLE SEISMOMETER EXPERIMENT IN THE SEA OF JAPAN

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Introduction

We are currently planning to install a downhole seismometer package in the Sea of Japan as a part of the ongoing Ocean Drilling Program. The experiment will be carried out on Leg 128 scheduled in August-October, 1989, based on the ODP proposal, #155F: Island arc to back arc basin transition" (Suyehiro et al., 1987).

The main difference of our plan from previous downhole seismometer experiments, utilizing DSDP/IPOD drill holes, is that it aims for long-term observation of large earthquakes to offer a unique seismic station beneath the seafloor to aid in mapping lateral heterogeneity of seismic structure of the earth.

On a local scale, the transitional structure of the crust from island arc to back arc is of interest. The change in thickness and wavespeed of each layer will provide an important clue in understanding the tectonic significance of large earthquakes that occur at the eastern margin of the Sea of Japan. The most recent case was 1983 Japan Sea earthquake (M=7.7).

Regionally, the subducted Pacific plate is penetrating the upper mantle beneath the Sea of Japan which is generally considered to possess low wavespeed and attenuative material. Uncertainties in their magnitudes and geometry must be removed to understand the plate subduction tectonics. This requires a station in the back arc basin. It is hoped that global seismology will benefit also from the data together with those from existing and planned broad-band land stations.

Site

The proposed site (J1b) is at the northern end of the Yamato Basin where the water depth is 2780 m (40°14.5'N, 138°15.1'E). The sediment thickness is 700 m. The drilling aims for 100 m into the basement and will perform vertical seismic profiling (Leg 127) and a large scale electrical resistivity experiment before downhole seismometer installation.

Experiment

We will install the instrument following the experience of the group at the Hawaii Institute of Geophysics. The sensor package (3-component) is clamped in the basement rock part of the hole about 800 m below seafloor. The digital data are cabled to the ship for on-line experiment and to the recording package resting on the seafloor for off-line experiment. The former takes place while the drill ship is at the site, while the latter is for long-term observation. One end of a long tight rope is attached to the recording package for retrieval. The other end holds a weight to be released by acoustic command. The difference from HIG installation is that the drill ship is different.

Online Experiment: The drill ship JOIDES RESOLUTION offsets about 1/2 mile from the cone as it pays out the electrical cable. Offset is necessary to avoid ship noise. A controlled source refraction seismic experiment is performed with a second vessel after deploying ocean bottom seismographs. Two or more days will be spent for this experiment to study the crustal structure. This will be a normal high frequency (i.e., 4-30 Hz) type experiment.

Offline Experiment: After the active experiment, natural earthquake observation starts. Because of the extreme environment, it is not possible to construct a system that maintains specifications of broadband digital seismic stations on land without considerable technological efforts. Given little lead time before installation, we have to rely on existing technology for this experiment. Seismic signals from earthquakes with M>5 within 10 degrees, M>6 within 40 degrees, or M>7 at all distances are to be recorded. Seismicity is not high near the site as compared to the Pacific side. So, nearby small events will not saturate the recorder. Dynamic range of 16 bits and frequency range of 0.01-20 Hz are sought.

Servicing: On-line wiring to land has to cover about 120 km distance avoiding fishery problems. At present, it is not feasible costwise. We chose off-line recording. The data must be retrieved after a certain period of time. With time, the recorder package will improve in terms of memory capacity and power consumption. But the first package will seek to run for about half a year. At 50 Hz sampling of six channels data, about 50 events with 30 minute record each can be stored on a commercially available digital tape recorder.

LONG-TERM DEEP-OCEAN BOREHOLE SEISMOLOGY - IS IT WORTH THE COST?

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What can be gained by placing seismic sensors in deep-ocean boreholes? Do we know enough about noise sources, noise levels, and propagation to answer this question? Are the advantages of placement in boreholes offset by the cost? Would it be more cost-effective to put arrays on the ocean floor? How deep below the bottom is best? How deep is too much? How do the answers to these questions change with frequency? In this 'white paper' I address these questions based on experience with deep ocean borehole instrumentation from DSDP legs 65, 67, 78, and 88, as well as experience with ocean bottom seismometers (OBS).

Noise Sources

The leg 88 experiment (OSSIV, Byrne *et al,* 1987) provides a continuous 64-day record of noise (and signals) at a depth of about 340m below the ocean floor just above basement in 5.5 km of water. Several conclusions have been drawn from these data that are discussed below.

The noise spectrum over the bandwidth of the data (0.1 Hz to 20 Hz) is dominated by energy originating at the surface of the ocean (Duennebier, *et al,* 1986). The spectrum of the noise from 0.2 to 5 Hz has characteristics that suggest that the energy source is ocean waves near the site. As the ocean wave spectrum changes in response to weather, a similar response is seen in the borehole noise level. As the wave spectrum saturates at high frequencies, so does the noise level in the borehole. As the seas become calm, the noise level in the borehole decreases by as much as 20 dB at some frequencies. Similar data have been observed on hydrophones and ocean bottom seismometers (OBS) on the ocean floor (McCreery, in revision, BSSA). These observations imply that some fraction of the ocean wave energy is transmitted to the ocean floor (and below) as a pressure disturbance, which is responsible in large part for the background noise observed from 0.2 to 5 Hz.

At higher frequencies (above 5 Hz) the OSSIV noise levels are below system noise when the weather is calm, but levels increase with with sea state along a different slope than the low frequency noise, implying a different noise generation mechanism. We believe that the higher frequency noise is generated by acoustic energy from breaking waves (Duennebier, *et al,* 1986). Its spectrum is nearly white to velocity, whereas the lower frequency noise is strongly red with a slope of about f^{-5} (particle displacement) when saturated.

While these data provide no information at periods below about 5s, other data do exist, and will certainly be discussed in this workshop. Of particular interest is recent work by Spahr Webb (submitted to Jnl. Oce. Eng) in the HEBBLE area, where he correlates pressure variations with bottom current flow.

Noise Levels

How do these noise levels change with depth in the bottom? The leg 67 and leg 88 data yield comparisons between the OSS borehole instrument and OBS's above them on the ocean floor. In the leg 78 OSSII data (Carter, *et al*, 1984), the vertical noise levels from 1 to 25 Hz are nearly the same when a correction is made for the impedance change from one site to the other. OSSII on Leg 67 was emplaced about 194m into the sediment, but well above basement on the inner slope of the Peru trench.

The vertical components of the OSSIV (Leg 88, Duennebier, *et al*, 1987) and an OBS directly above it had nearly identical spectra above 2 Hz, including spectral peaks from the nearby ships. The noise levels on the horizontal geophones are about 20 dB lower in the borehole than on the ocean floor, however, after making corrections for impedance changes. The horizontal particle motion spectrum also changes shape from the ocean floor to the borehole sensors indicating different noise sources. These changes are almost certainly the result of the very strong shear waveguide at the ocean floor. All shear energy coming up from the basement is refracted into paths within a few degrees of vertical, thus shear particle motion is almost completely restricted to the horizontal plane. Any local slope in the ocean floor will further trap shear energy near the ocean floor.

At frequencies below 2Hz, the best data comparing borehole and ocean bottom seismic data are from the Ngendei MSS Experiment on Leg 91 (Adair, *et al*, 1987). Their results indicate that noise levels at the microseism peak (between 5 and 10s) are nearly the same in the borehole and on the ocean floor. This observation implies that there is little to be gained by emplacement in a borehole over emplacement at the ocean floor for observations at long periods.

Several questions remain: 1. How far below the ocean floor is the horizontal noise level high? 2. How far below the ocean floor is ocean surface noise important? 3. What factors control noise level changes below 1 Hz? 4. How do noise levels change with sediment thickness and lithology?

Coupling

A seismometer should sense the motion of the ground with a known (and relatively simple) transfer function, but particle motion measured by seismometers placed on the ocean floor (and on other compliant surfaces) are often distorted by motion of the instrument package relative to the material they rest on. This problem is particularly serious where shear velocities are low and where relative motion of the medium above and below is strong. If the OBS has any cross section in the water, the part in the water will attempt to move with the water, and the part in the bottom will try to move with the bottom causing distortion by tilting not only when currents are present, but also in response to shear wave arrivals from below. The problems of coupling at frequencies above 1 Hz are covered in several papers, the most recent being Sutton & Duennebier, 1988. They conclude that the best coupled OBS's will have small mass, radius in contact with the bottom proportional to the cube root of the mass, very low profile in the water, density equal to the density of the bottom, and no density gradients in the package. The ideal OBS is buried in the sediments, has the sediment density, and is smaller than one quarter of a wavelength of the highest frequency wave to be measured. As shear velocities at the ocean floor can be less than 30 m/s, the latter is a severe constraint for high frequencies.

Ocean bottom currents can be coupled into the ocean floor (a problem addressed by Spahr Webb in his new paper), causing tilts that appear to be ground motion. These distortions can be lessened if the OBS is buried at sufficient depth, and package tilt should be minimized if its dimensions are small relative to the wavelength of the distortion.

Signals

Based on the above, it isn't obvious that there is significant advantage in deep borehole emplacements relative to shallow burial. A statistical comparison of the relative sensitivity of OSSIV and an OBS at Site 581 (leg 88) indicates that the borehole (OSS) system was more sensitive to regional earthquakes (between 6 and 30 degrees) than the OBS (Cessaro and Duennebier, 1987), and the increase in fidelity and S/N in the borehole made it possible to make sense out of particle motion, which is difficult (at best) on OBS's. There is no doubt that the signal quality in the borehole was superior to the that on the ocean floor at this site. Results from Ngendei (Shearer, *et al,* 1987) on Leg 91 show that the vertical particle motion is virtually identical in the borehole and on the ocean floor, but better S/N in the borehole made it the better instrument for detection. The question still remains, however - how deep is deep enough?

Cost

Dollar cost of emplacing long-term borehole experiments is *at least* an order of magnitude more than equivalent ocean floor experiments. These costs result from the necessities of getting the sensors in the holes and the data back out. In addition, the drilling of a hole suitable for emplacement is not a high probability task. Many holes (particularly deep ones) never reach their target depth. If emplacement requires the drill ship, as it did for the OSS (Byrne, *et al.* 1987) and MSS (Harris, *et al,* 1987, 1988) experiments, then the scientists must ride the ship for about 50 days or meet it at sea to emplace their experiment. If emplacement is to be done in an existing hole, (wire line reentry) then a ship with excellent station keeping ability and the necessary reentry hardware must be used. Hardware for wireline reentry is now being developed, but to my knowledge no deep-ocean borehole has yet been reentered with a wire. An experiment designed for wireline reentry needs to be designed not only take the desired data, but also to transmit data on wire tension and depth in the hole. It must be rugged enough to survive possible punching through bridges in the hole, and seismic sensors must be locked into the hole when the desired depth is reached. Although the costs are high, deep-ocean borehole experiments are attractive because significant improvement over ocean floor experiments is expected, and, in the near future, funding for ocean drilling related experiments will almost certainly be more available than funding for ocean floor experiments.

Logistics

Drilling deep-ocean holes requires competing with other scientists for drill ship time. This competition is fierce, and particularly difficult when holes are required in regions of little interest to other scientists, or in places with logistical problems, such as deep, reentry, bare rock, and high latitude sites. Experiments such as the OSS and MSS require an additional research ship in support of the experiment, again increasing cost and logistical problems.

Arrays

Arrays of seismic instruments have great advantages over single instruments for many studies. Instrumenting an array of holes in the deep ocean is a particularly appealing prospect, but one that would be difficult to justify to the science community because of the drilling time and expense involved. It is probably more reasonable to emplace an array of instruments on or just below the ocean floor for much less cost (or ten times more elements).

Needed Experiments

To answer the question "how deep is enough?" we need to emplace seismometers in drill holes at several depths in several different environments and evaluate signal and noise characteristics over a broad frequency band. One experiment requires an *uncased* borehole through at least 300 m of sediment and 50 m or more of basement. An instrument package could be emplaced at the bottom and pulled up through the hole, stopping to take data at many levels. This should be done without the drill ship in the area, since it is a large source of noise, but the experiment will require a ship on site. This experiment could be done with an OSS-type package, although broader band sensors are desirable. A MSS-type package is probably too massive, and would have coupling problems in the softer sediments. Some problems that could be encountered in this experiment include noise from the cable attached to the ship, coupling in soft sediment near the top of the hole, and difficulty in clamping and keeping the package vertical as hole diameter changes. A well-designed OBS at the top of the hole should be part of this experiment.

Planned Experiments

Multi-node Borehole Seismometer. The Navy is currently funding a deep-ocean borehole seismic system for wireline reentry. Many people know more about this system than I do, but I proposed one such system several years ago and will express my concerns. Each node must be mechanically coupled to the wall of the hole and de-coupled from the cables above and below it to obtain reasonable fidelity of ground motion. Care should be taken to isolate each node from the tension pulling up from above and down from below. Any nodes emplaced inside casing are unlikely to obtain reasonable fidelity, since no attempt is made during drilling to ensure that the casing is coupled to the surrounding hole, and the stiffness of the casing is usually much greater than the stiffness of the surrounding rock. Casing is pipe hung from the reentry cone to reduce the diameter of the hole in soft sediment and prevent caving in unstable sediments. Casing is cemented in at the bottom to stabilize it in the hole, but not along its length.

Shallow Burial OBS. Also funded by the Navy, this instrument (to be constructed at HIG this year) will have two broad-band seismic packages buried in the sediment in shallow water (~300m) as well as current meters and pressure sensors. Data will be transmitted over electro-optical (E/O) cable to a ship about 20 km from the sensors. Data from this experiment will address the question of how fidelity of data increases as sensors are buried in sediment, and how much their susceptibility to ocean bottom current noise is reduced. This experiment will be run in late 1989 or early 1990.

Geophone Streamer. A 30-node geophone streamer with 1 m spacing between nodes is being constructed at HIG for Rondout Assoc. and WHOI (George Sutton and John Ewing) with ONR funding. Each node contains orthogonal accelerometers, a hydrophone, and an 'up' direction sensor. Data are digitized and transmitted over an E/O cable for recording on ship. The array is designed to be laid on the bottom for shallow water sediment studies. First experiments are scheduled for summer, 1988. Data should address questions of shear velocity changes with depth, noise propagation, and anisotropy.

HUGO

The Hawaii Undersea Geo-Observatory is a deep-ocean laboratory to study Loihi Seamount, an active submarine hotspot volcano 35 km SE of the island of Hawaii. The project is proposed (Duennebier) to NSF. The instrument systems planned include a hydrophone network, three broad-band seismic stations, and seven short period seismic stations near the summit of the volcano, with capacity to add many more experiments in the future using submersible or ROV emplacement of instruments. Continuous real-time data will be transmitted to shore over E/O cable. Japanese Experiment. Suyehiro and others are planning to emplace a broad band seismic experiment in the Japan Sea on leg 129 using a method similar to the OSS Experiment.

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SIGNAL BEHAVIOR IN BOREHOLES

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The question of seismic signal behavior in ocean (as well as land) sub-surface drillholes is of great interest to a number of geophysical research, seismic exploration, and military applications groups, in the VLF (50-5 Hz), ELF (5 - .5 Hz), and ULF (.5-.005 Hz) bands. In marine crustal and oil exploration, marine vertical seismic profiles (VSPs) and cross-hole tomography using controlled sources provide signal amplitude and coherency data from which *in situ* P and S wave velocity and attenuation vs depth can be determined [2,5,8-11,15,18,43]. Another exploration interest are the source and geo-environmental factors controlling signal-generated noise specific to the cylindrical near-borehole volume ("tube waves" [3,13,19,24]). The DARPA "Vela Uniform" Project [22 & references], and similar Soviet efforts [10 & references], for over 13 years extensively examined the relative merits of 1-dimensional borehole vs 2-dimensional surface seismometer arrays to improve discrimination and localization of man-made vs natural teleseismic signals. It is believed by many [46b,c,j,l] that the post-1975 oil-exploration and Vela Uniform work define the current state-of-the-art in experimentally and theoretically understanding seismic signal generation, propagation, noise corruption, reception, and inversion in and around both shallow and deep boreholes.

What is currently known concerning theoretical signal behaviors [6,8-9,11,16,20,26,30,37-43] as well as experimental results supporting improvements to sensor-design and signal processing technologies [2,13,23,43] has to date largely been limited to thickly-sedimented near-shore environments, in the 2000+ marine boreholes utilized to date for marine VSPs in water depths from 10-to-1500+ m, in areas of commercial petroleum interest (Gulf of Mexico, North Sea, offshore West Africa, offshore Indonesia). Unfortunately, only rarely do petroleum exploration interest regions overlap with those of interest to crustal structure, earthquake, and geodynamic studies (Gulf of California, Gulf of Alaska, Spitzbergen [46a,c,d,g,h,n]). The prospect of utilizing existing or future exploration VSP data (recorded at 15-25+ seconds at each depth) for crustal seismic exploration studies has been shown [8,11,39,46r], from theoretical and experimental [46d,r] studies, as a cost-effective and potentially important supplement to dedicated scientific boreholes for continental boreholes, and in some cases also for marine seismic studies, also thereby potentially furthering the existing borehole signal behaviour database.

Another area of major technology/results transfer of potential interest to seismic instrumentation requirements for marine borehole studies are the major improvements in multi-sensor sondes and sonde/borehole locking technologies and quality control, yielded by 11+ years of proprietary R&D in the petroleum exploration industry (notably by CGG [46a], Gulf-Chevron [46c], Phillips [46h], Arco [46g], and Schlumberger [46e]). It is felt that despite the difficulties in obtaining detailed access to these results and equipment, the potential payoff for more rapidly-achieved and improved designs in scientific marine borehole seismic studies merits further dedicated pursuit [46a,o]. From the perspective of fundamental science, in the oil exploration industry and to a lesser extent in the present Vela Uniform project, standard EDP and numerical modeling methods have been continually evolving to better address fundamental wave-propagational questions such as the relative magnitude, geologic controls, and frequency/depth behaviors of *in-situ* signal vs ambient noise, and viscous vs stratigraphic vs porous attenuative seismic loss mechanisms. However, with the exceptions of the original "Vela Uniform" work by Romney [references to 22], Sax and Hartenberger [28,29], Douze [6], Roden [25], and others [12,26], the above studies have to date been carried out mainly for sedimentary (with few hard-rock) sites on land or to a lesser extent at sea. The results of marine experiments since 1979 using DSDP holes [33-35], to determine (e.g.) the comparative behaviours of hydrophone vs geophone sensors, the depth/geology dependence of downhole signal and noise levels, and the detailed decomposition of the seismic-acoustic propagating wavefields, have been preliminary and in some aspects contradictory [41]. General borehole seismic signal behavior (see **Figures 1a-e**; below) can be summarized as usually exhibiting (a) a nearly-linear trend of increase in net transmission loss ("TL") with depth that on average increases with increasing frequency, and (b) superposed on (a) are localized increases and decreases in coherent TL, and likewise frequency- and angle-dependent resonance effects [46q], more clearly seen in TL than net-energy measures [38,42]. With the still inconclusive results of the previous MSS experiment, the DSDP 418 LFASE experiment of 1989 will be the first to employ state-of-the-art (quiet) multi-node sensor sondes to permit a more detailed assessment of seismic-acoustic signal and noise energy and coherency behaviour vs borehole depth in a well-characterized ocean/bottom environment [41,46e]. Where the above results are not well known to the marine geophysical research communities at large, to the extent that they can be made at least in part available free of classification or proprietary restrictions [46e], it is strongly felt that their careful consideration, prior to future proposed experiments with permanent sub-bottom seismometer observatories, will maximize new science results and minimize repetition of old experimental and engineering errors.

Notwithstanding 30+ years of investigation, many fundamental questions remain, on basic signal propagation and reception effects in both the near- and far-fields of shallow and deep (land and marine) boreholes. Such questions include:

- the depth-dependence laws of signal amplitude/energy and phase behaviours vs frequency;
- " " vs local layer elasto-acoustic moduli (in addition to (LVL's)Vp and Vs, P and S wave attenuation, porosity and permeability, sediment thickness and stratigraphic layer disposition, interface asperity, etc);
- the relative roles of local and near borehole vs regional earth media inhomogeneities (scattering effects) in governing dominant ("optimum"?) signal propagation frequency, spectral and coherency characters;
- "optimum" sensor types and configurations, for given applications, as a function of depth below seabed, central signal frequency, consolidated/unconsolidated near-surface sediment, source depth and distance, etc;
- use of *apriori* and *insitu* geophysical and geoacoustic data for given boreholes to make "site (transfer function) corrections" to observed seismic instrument response better than simple 'rc' approximations;
- the role and occurrence conditions of borehole "standing waves" as components of overall downhole seismic response;
- the relative partitioning and identification of seismic signal and noise into (low and high-order) surface, interface, inhomogeneous, and multiply-reflected body waves;
- the relative importance of distributed vs specific noise ambient, and signal-generated noise mechanisms in complicating reception of specific desired seismic signals, and the manner(s) in which specific water-to-bottom noise source contributions are "channeled", multi-ordered modes are "coupled" and ultimately reach effective "equilibrium" levels at (what?) source-receiver offsets;
- the relative merit of shallow vs deep, soft vs hard rock, remote vs near-culture, and cased vs additionally-conditioned boreholes in maximizing specific signal/minimizing specific noise reception;
- the availability, implementability, and accuracy of theoretical and numerical seismic propagation models, to predict signal and noise behaviours for all geologies, signals, sensors and sensor depths (nb., "node points") , and noise types?

The above and other questions are equally important for surface-wave dominant as well as body-wave and mixed (body/surface) wave signal types. It is believed by many experimentalists

[46a,d,e,m,o,p] as well as seismic model developers [46b,f,j,p] that a key factor in calibrating, quantitatively predicting, and extrapolating scientific borehole results will Not be primarily the compiling of a larger number of new site specific results (since even the most optimistic eventual number of deep ocean boreholes will be insufficient to provide a statistically-significant database for definitive borehole signal propagation and noise characterization), but will also require the use of the best possible integrated signal+noise numerical models, routinely used to quantitatively assess competing environmental (geologic-hydrodynamic), experimental-engineering (sensor-type/position/coupling/source behaviour and directionality), and elastic wave propagation factors governing near and farfield borehole signal propagation, over a wide range of subseafloor sites of interest. It is also clear that fundamental progress in understanding and exploiting the 3D characteristics of seismic signal propagation to and within (deep ocean) boreholes will not be achieved without corresponding attention to the detailed physics of seismic-acoustic signal conversion, scattering, (re)generation, and propagation [10,18,26,28-29,36,38], and embodying applicable theories into practical working numerical prediction models. Ongoing results in these areas from other exploration, academic research, and military areas should be comprehensively identified, to technically support (and avoid "wheel" or error re-discoveries) in the proposed objectives of deep (sub)sea permanent seismic observatories.

As a single example, although evanescent (Rayleigh, pseudo-Rayleigh, Scholte, Stoneley, and inhomogeneous-root) wave amplitude and particle motion behaviours in range and depth away from a generating interface are well known for simple few-layer cases [35], despite 30+ years prior study, it is still not well known how source depth/near-source and farfield in-earth stratigraphy and structure affect the complexly-changing mix and equilibrium of surface and body waves in both ambient as well as signal-generated noise [26,29]. This latter could be particularly problematic for proposed shallow-burial deep ocean seismic experiments: in these water depths, although a plane-stratified bottom yields interface waves circa the microseism peak frequency (~.15 Hz [31]), for many kinds of seafloor meso- and micro-roughness, body-to-interface wave conversion via rough surface scattering and re-radiation has been theoretically and experimentally shown to yield additional interface waves of the same short or long period frequency content as the incident source signal [46e]. This heretofore largely unexamined factor could assist or invalidate desired subbottom seismic sensor advantages, and an improved understanding of these phenomena through improved numerical modeling capabilities should precede and accompany any proposed high-risk/high-cost field experiments. Another persistent problem, common to both oil exploration sedimentary environments [46k], and teleseismic cratonic propagation [10], is the common and to date largely-inexplicable observation of strong ~15 Hz tonals, as well as other "high VLF" frequency "optimum" propagation frequency and/or industrial-specific noise mechanisms, frequently observed in many land and marine borehole seismic measurements, apparently independent of local geology [46d,k] (exploited to date in real-time seismic tracking of oil well drill bits *in situ*). It is suspected that the probable existence of these "optimum" signal propagation frequencies, and "equilibration" processes for noise, in the crustal waveguide poses a strong challenge, both to experimental and theoretical seismologists, on how the many environmental-propagational parameters in combination actually control seismic signal and noise generation and propagation, in the context of an "effective" or "equivalent" parameter model [21,39,46m] holding an improved answer to the questions of how much or how little local and regional pre-experimental site geosurveying for detailed earth moduli and structure is practically needed for predicting optimum borehole location and depth, sensor type number and depth, etc. Unlike some short-period seismic signal and noise decay problems, it is not felt that simple rules-of-thumb (thin unconsolidated low shear modulus sediments = fast depth decay; high near surface asperity = higher noise) will suffice for envisioned subseafloor seismic observatories, and that only serious conjoint numerical and field experiments within a statistical framework will allow meaningful further progress to a problem that has been recognized but repeatedly resisted recognition and solution for over 30 years.

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DOWNHOLE AND SEAFLOOR SEISMIC MEASUREMENTS MADE BY THE HAWAII INSTITUTE OF GEOPHYSICS: PAST, PRESENT, AND FUTURE

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Abstract

Various measurements of seismic signals and ambient noise in the oceanic water column and on or below the ocean bottom have been made by the Hawaii Institute of Geophysics (HIG) over the past two decades. Below the ocean bottom, data have been collected by HIG's Ocean Sub-bottom Seismometer (OSS) from deployments down four Deep Sea Drilling Project (DSDP) drillholes; measurements on the seafloor have been made using HIG Ocean Bottom Seismometers (OBS's) and using the Wake Island Hydrophone Array (WIA); and water-column measurements have been made using WIA hydrophones. These data can answer some questions relevant to the planning of future long-term seismic instrumentation in the oceans. The first question is where to site sensors within the ocean-sediment-basement column in order to maximize the signal-to-noise ratio and the signal fidelity of various seismic and acoustic signals of interest. The data show that for frequencies between about 0.5 and 30 Hz, signal-to-noise can be increased and signal fidelity can be improved by siting the sensors below the ocean-sediment interface. These changes are especially pronounced on horizontal sensors. Although siting the sensors below the ocean bottom is clearly advantageous, it is not clear from the data just how deep into the sediments or basement it is necessary to go in order to achieve a significant improvement in signal-to-noise and signal fidelity. Another important question is where to site future ocean seismometers regionally in order to avoid sources of ambient ocean noise that can reduce signal-to-noise ratios. The long-term data provided by WIA, OSS, and the OBS's show that there are many significant sources of noise between 0.1 and 30 Hz that can be avoided by careful siting. These noise sources include: (1) high surface winds, (2) coastlines, (3) high bottom currents, (4) shipping, (5) hurricanes, typhoons, and cyclones, and (6) whales. In addition, the successes and failures of the electrical and mechanical design of HIG instruments provide useful information for the design of future ocean-environment seismic systems.

INTRODUCTION

Data collected and experience gained by the Hawaii Institute of Geophysics (HIG) over the past 10 years are applicable to the current goal of collecting new broadband long-term seismic data in the oceans to fill the gap in the worldwide data set. Existing HIG data include measurements made by seismometers both on and below the ocean floor, and by ocean bottom hydrophones. The data are generally in the short-period seismic band, ranging from the microseism peak at approximately 0.25 Hz to the upper frequency limit of teleseismic Po, So, and T phases at around 30 Hz. These data provide information about short-period ambient noise levels in different oceanic regions as well as at different depths in the water column and below the sea floor. They also provide information about signal-to-noise ratios as a function of depth below the sea floor for various signal sources. HIG experiences in designing, building, deploying, and recovering ocean bottom and sub-bottom seismic instruments have demonstrated numerous advantages and disadvantages. This information should be applied, when applicable, to future instrumentation and field work. Thus, it is the goal of this report to summarize the HIG ocean seismic work from both a science and engineering standpoint, especially as it pertains to evolving plans for future seismic data collection in the oceans.

INSTRUMENTATION

Ocean-Bottom Seismometers (OBS's)

The HIG ocean bottom seismometer program began in the early 1970's. By the mid-1970's a design called the Pop-up OBS or POBS had evolved and was being used on a regular basis for refraction and earthquake studies. The POBS consisted of: (1) an explosive-bolt-releasable lead anchor with two (redundant) preset release timers to fire and bolts, (2) a main pressure case containing a vertical and horizontal seismometer, preamplifiers and gain-ranging electronics, a clock, and up to three slow-speed four-channel cassette tape recorders, (3) a hydrophone connected by cable to the main pressure case, (4) recovery floats (glass spheres), (5) two surface-activated recovery strobe lights, and (6) two surface-activated recovery radio transmitters. The instrument was deployed from a ship by dropping it into the water and letting it sink freely to the ocean bottom. It would be recovered at a later pre-determined time, when the release timers would fire the explosive bolts, releasing the lead anchor and causing the package to rise to the surface. The floating package would then be located from the ship by using an onboard radio-direction-finding receiver tuned to the recovery radio transmitters and by scanning the ocean surface for the strobe lights (recovery was generally done at night). The POBS was capable of collecting up to six weeks of data at frequencies up to 30 Hz, with a dynamic range of about 72 dB (30 dB on the tape plus 42 dB for seven 6-dB gain steps). Although the POB's worked fairly well, the ambient noise was often high, and signals were sometimes monochromatic in character. After some investigation, the high noise was attributed primarily to ocean-bottom currents strumming the high-profile, inverted-pendulum package, and to coupling between mechanically noisy tape-drives and the seismometers housed in the same pressure case. The monochromatic signals were attributed to poor coupling between the POBS package and the bottom.

These problems were rectified to some extent in a new design called the "isolated sensor OBS" or ISOBS. The electronics and recovery systems in this design were nearly identical to those in the POBS. The seismometers, however, were deployed in a separate small pressure case that was mechanically decoupled from the rest of the ISOBS (Fig. 1). This configuration eliminated tape recorder noise and greatly reduced current strumming noise. And although coupling between the seismometers and the ocean-bottom was improved, it still remained a significant problem. The ISOBS instruments were in use until 1987. A more detailed summary of the POBS and ISOBS design characteristics has been written by Byrne et al. (1983).

Ocean Sub-bottom Seismometer (OSS)

HIG began development of the ocean sub-bottom seismometer in 1978, and between 1979 and 1982 four OSS deployments were made from the D/V Glomar Challenger at sites in the Pacific and Atlantic. Technical problems limited the amount and quality of data from the first three OSS's. OSSIV, however, had high-quality, real-time seismic data during deployment, high-quality, long-term recorded data during the first two-month remote operating period, and high-quality, real-time data during recovery and redeployment. OSSIV was (and still is) located down Deep Sea Drilling Project Hole 581C near the Kuril Islands in the northwestern Pacific, with 60 days of data waiting to be recovered. Water depth at the site is 5467 m, and the sediment cover is 357 m on top of basement basalts. The OSSIV sensor package was deployed in the sediment just above basement. Data from

OSSIV, and from OBS's deployed nearby during the deployment period are discussed later.

A general diagram of the deployed OSS system is given in Fig. 2. The borehole sensor package contains geophones stacked along three axes, a temperature sensor, a 2-component tiltmeter, and floating point analog-to-digital electronics. It is clamped inside the drillhole by a remote controlled extension arm. The data are transmitted in digital format up an electromechanical cable to a recorder package suspended above the bottom nearby. The recorder package converts the digital seismic data back to three analog signals for recording on 4-channel, slow-speed, analog cassette tape recorders similar to those used in the HIG OBS's. The temperature and tilt data are encoded in a time code signal and recorded on the remaining tape channel. Five tape recorders provide the capability to record continuously for up to 66 days with frequencies up to at least 30 Hz. Attached to the recorder package is a long, positively-buoyant, polypropylene rope that can be grappled for as a backup recovery procedure. (The excess positive buoyancy of this rope suspends the recording package above the bottom.) At the far end of the rope is the primary recovery system. It consists of an anchor assembly, recovery buoy, and two redundant acoustic releases.

The OSS system must be deployed by a drillship. The sensor package (tool) is lowered from the drillship through the drillstring (positioned above the bottom of the hole) using electromechanical logging cable. When the tool is at the proper depth it is clamped in position by remote command and tested for proper operation. Then the cable is cut and the drillstring is stripped off around the cable by a special procedure. At this point, real-time recording of the digital sensor signals can be made onboard the drillship, and active seismic experiments can be conducted. For final deployment, the recording package, recovery system, and anchor are attached and deployed over the side. Data recovery is accomplished at a later time using a standard research ship. An acoustic signal is sent from the ship to the acoustic releases, causing the anchor to be released and the recovery buoy to float to the surface. If this doesn't work, the polypropylene rope is grappled for. The recording package can then be pulled up from below and brought onboard. Tapes and batteries can be replaced; real-time experiments can be performed; and the recording system redeployed without disturbing the tool. If removal of the tool is desired, it can also be pulled out of the hole, although this is risky since the electromechanical cable may break if the took gets stuck. A more complete description of the OSS system -- its specifications and procedures for deployment and recovery -- is given in a report by Harris et al. (1988).

Wake Island Hydrophone Array (WIA)

The Wake Island Hydrophone Array is an array of twelve hydrophones located near Wake Island in the northwestern Pacific Ocean (Fig. 3). Six of the hydrophones are on the ocean bottom at 5.5-km depth, at the center and vertices of a 40-km-wide pentagon located approximately 100 km to the north of Wake. The other six hydrophones are located at three sites to the south and west of Wake and are at the depth of the SOFAR (SOund Fixing and Ranging) channel axis at about 0.8-km depth. The entire array spans an area that is about 100 by 300 km. The hydrophones are passive, moving-coil type, and signals from the hydrophones are transmitted to Wake via long undersea cables. The array was installed in the late 1950's as part of the US Missile Impact Location System, but it was abandoned by that project after many years of use. Since 1976, HIG has been using the array to record seismic signals. In 1982, a digital recording system was installed to record signals from eight of the WIA hydrophones (five bottom, three SOFAR) on a continuous basis. This recording system is still in operation, and the reduced data, consisting primarily of ambient noise samples and time intervals containing seismic signals from earthquakes and explosions, are stored at HIG and at the DARPA Center for Seismic Studies in Rosslyn, Virginia.

ADVANTAGES AND DISADVANTAGES OF LOCATING SENSORS AT DIFFERENT DEPTHS WITHIN THE OCEAN-SEDIMENT-BASEMENT COLUMN

SOFAR Depth

The SOFAR channel is a low-velocity zone in the water column that acts as a very efficient acoustic waveguide. It is found in all the world's oceans, and its axis is generally at a depth of 1 km or less. Sound waves within this channel can propagate with little loss over thousands of kilometers. Seismic measurements near the SOFAR channel axis can be made using hydrophones suspended above deeper ocean floor, or using seismometers (or hydrophones) deployed on seamounts that rise to axis depth.

The primary advantage of locating sensors at SOFAR channel depth is the abundance of long-range water-borne signals that will be observed. Using an array of SOFAR hydrophones in the Pacific (including the WIA SOFAR hydrophones), Duennebier and Johnson (1967) located 6.5 times more northwestern Pacific earthquakes than were located during the same one-year period by the worldwide seismic network. SOFAR sensors can detect signals from events other than earthquakes -- for example, submarine volcanoes and nuclear test explosions. MacDonald Volcano at the southeast end of the Austral seamounts was discovered by triangulating on underwater sounds produced by its eruptions (Norris and Johnson, 1969), and underwater volcanism from Kaitoku Seamount was observed by Walker et al. (1985) using WIA SOFAR hydrophones. Extremely strong acoustic signals are also routinely recorded at WIA from French nuclear test explosions at Mururoa Atoll, more than 10,000 km away (McCreery and Walker, 1988). One of the important advantages of recording water-borne signals is that they travel quite slowly (1.5 km/sec) compared to the phase velocities of the other oceanic short-period phases P, Po, or So. This property makes it somewhat easier to locate distant events with small arrays, since time offsets between sensors are relatively large.

The primary disadvantage of locating sensors at SOFAR depth is high noise. The same properties that make the channel efficient for transmitting signals also make it efficient for transmitting noise. The primary noise sources in the frequency band from 1-30 Hz are ocean surface wind waves and shipping. Other noise is produced by scientific and military underwater explosions, and by biological sources such as whales. Signals from even small explosions can propagate in the SOFAR channel over an entire ocean. A typical one-day helicorder record of a WIA SOFAR hydrophone (Fig. 4) has many tens of discrete signals, most of unknown origin. Figure 5 from McCreery (1988) shows a comparison between the one-year-average ambient noise on four WIA hydrophones, two at 5.5-km depth (74 and 76) and two at the SOFAR axis depth of 0.8-km (10 and 20). It is clear that the SOFAR hydrophones are increasingly noisy towards the high frequencies. This additional noise impairs the SOFAR hydrophone's ability to detect small P, Po, and So signals that are more easily observed on ocean bottom sensors. The excess noise of hydrophone 10 relative to hydrophone 20 is due to hydrophone 10's proximity to the Wake shoreline, a topic that will be discussed later in this report.

Deep Ocean Seafloor

Extensive measurements of signal and noise on the seafloor in deep ocean using OBS's and using the deep WIA hydrophones have been made at HIG. There are many advantages to taking measurements on the deep ocean floor including: (1) low ambient noise at frequencies above 3 Hz, (2) a high level of separation between transverse and longitudinal components of seismic phases refracted from basement or below, (3) the relative ease of instrument deployment and recovery, and (4) the large number of flat deep-ocean-basin sites (>40% of earth's surface at depths >4000m). Some disadvantages associated with deploying instruments on the deep-ocean floor are: (1) high ambient noise levels around 1 Hz, the dominant frequency of teleseismic P, (2) difficulty in achieving good coupling between the seismometers and the sediments, and (3) ocean-bottom-current induced noise, especially on the horizontal seismometers. A more detailed discussion of these factors follows.

Figure 6 shows some ambient deep ocean noise spectra compared to average continental noise. It can be seen that ambient deep ocean noise is relatively high at frequencies around 1 Hz, but relatively low at frequencies above 3 Hz. For this reason, the deep ocean is a good place to detect seismic signals that are rich in high frequency energy. These include oceanic lithosphere phases Po and So, teleseismic P from deep-focus earthquakes, and teleseismic P from nuclear test explosions. Conversely, the deep ocean bottom is generally a poor place to detect short-period mantle-refracted P from shallow-focus earthquakes, since the dominant frequency of those phases is around 1 Hz where noise levels are high.

Urick (1986) plotted noise levels at several discrete frequencies as a function of depth, and the noise clearly decreases with depth below the axis of the SOFAR channel. But perhaps the most important feature of these curves is that the noise decreases most rapidly below critical depth. Critical depth is the depth of the lower boundary of the SOFAR channel -- the depth at which the acoustic velocity equals the maximum velocity at the upper boundary of the SOFAR channel, near or at the surface. Critical depth can vary from less than 1 km at latitudes greater than 60 degrees, to more than 4 km at mid and low latitudes. Critical depth near Wake is approximately 5 km, and the WIA deep hydrophones at 5.5 km are below this threshold. Although a long-term systematic study of seismic noise levels at different ocean depths has not been made, it would appear from the Urick data that it is important to site the seismic sensors as deep as possible, preferably below critical depth, to minimize noise.

Figure 7, from Duennebier et al. (1987a), shows some ISOBS refraction data recorded in the northwestern Pacific in 1982. The ISOBS was deployed in 5500 m of water, and rested on top of approximately 350 m of pelagic sediment. A striking feature of these data is their high quality, and also the clear separation of compressional and shear arrivals. The hydrophone record mainly shows refractions that are compressional at the instrument, with almost no hint of shear arrivals. Conversely, the horizontal record mainly shows refractions that arrive as shear energy, with almost no hint of compressional arrivals. The vertical record is most like the hydrophone record, but with some low-amplitude shear arrivals. Cross coupling in the sensor package between horizontal and vertical signals that correspond to shear arrivals on this record. These data clearly illustrate one advantage of sensor placement on (or possibly within) ocean bottom sediments. Low sediment velocities force arrivals coming up from basement to propagate nearly vertically, thus compressional motions are in the vertical direction and shear motions in the horizontal direction. This allows for their easy identification on records such as these with vertical and horizontal sensors. The data also illustrate that OBS data can be very high quality, with high signal-to-noise and signal fidelity on all components. Our experience indicates that the primary reason why most OBS data are not of this high quality is that they are usually poorly coupled to the sediments and that they are often noisy due to the instrument vibrations in bottom currents.

Below the Seafloor

Measurements of signal and noise in sediments below the seafloor have been made using the HIG OSS. These measurements show that the primary advantages of deploying seismic sensors below the ocean-sediment interface are higher signal fidelity and lower noise. The data do not show, however, just how deep into the sediments it is necessary to go in order to achieve these gains. The main disadvantage of siting sensors below the seafloor is the greatly increased complication and expense of a drillhole deployment compared to an ocean bottom, OBS-type deployment. This disadvantage might be reduced considerably if it was only necessary to implant the seismometers a short way into the sediment using a method that did not require either a drillship or an old drillhole.

Improved signal fidelity below the ocean-sediment interface is due primarily to improved coupling between the seismometers and the sediments. Figure 8 shows a diagram of the OSS tool deployed down a drillhole and locked into position. The lever arm has been extended by remote control from the drillship, and the sensor package is firmly locked against the side of the drillhole, ensuring that the seismometers are well coupled to the sediments. Figure 9, from Duennebier et al. (1987b), is a 43-minute, rectified, 3-component record of noise produced by a ship passing near OSSIV. This figure illustrates the general fidelity of particle motion that is recorded by the OSS. The signal envelope for the vertical sensor record has one central peak, and the envelopes for the two horizontal sensors are doubly peaked. These features are predicted by a simple computer model of the data envelopes for a ship passing nearby at an oblique angle to the horizontal seismometer axes. Vertical motions are largest when the ship is closest to the array. Horizontal motions are largest when the ship is located in line with the seismometer axis. In the data, maxima for the two horizontals are at different times as expected, and they straddle the maximum for the vertical as expected. Thus, the OSS appears to be yielding accurate particle motion data. Figure 10, from Duennebier (1987), shows arrivals from a magnitude 7.8 earthquake in Japan recorded by OSSIV at 15.7 degrees epicentral distance. The broad range of frequencies recorded (11 octaves) and the unique character of each trace are a strong indication that the OSS sensors are well coupled to their surroundings. What these combined data show is that a seismometer located below the ocean-sediment interface can be effectively coupled to the sediments by a fairly simple mechanical design, and can thus be used for recording accurate particle motions. They also illustrate some of the interesting seismic data that can be studied in the oceans when the sensors are well coupled to the surrounding material.

Reduced noise levels are also observed below the seafloor. Figure 11, from Duennebier et al. (1987b), is a comparison between ambient noise spectra recorded simultaneously on the vertical and horizontal components of OSSIV and a nearby ISOBS. The general similarity in shape of the two vertical spectra indicates that noise sources are essentially the same, however, noise levels down the hole are 20-30 dB lower than those on the ocean bottom. A correction for impedance differences between the sediment down the hole and at the surface can only account for only a few dB of this separation at most. On the horizontal components, the ISOBS spectrum and the two OSS spectra are generally dissimilar in shape, indicating that their respective noise sources are probably not the same. Horizontal noise levels are 30-40 dB less down the hole. More than 15 dB of this separation may be due to the large impedance difference caused by extremely low shear velocities at the top of the sediments (0.05 km/sec are typical). A limited study of nearby, unreported earthquakes recorded by OSSIV and several ISOBS's during different time periods (Cessaro, 1987) appears to indicate an increased sensitivity of approximately one magnitude level on OSSIV compared to the ISOBS's. This implies that OSSIV had a signal-to-noise improvement of around 20 dB over the ISOBS's for signals propagating up from below, approximately the same as the reduction in noise. Thus, it appears that significantly reduced noise levels, and correspondingly increased signal-to-noise levels for signals arriving from below, can be found below the ocean-sediment interface.

REGIONAL SOURCES OF AMBIENT NOISE THAT CAN BE AVOIDED

An important factor to consider in determining future regional sites for the deployment of long-term seismometers in the oceans is excessive or extraneous ambient noise. Just as it is important to choose a quiet site for a land seismic station, away from roads, streams, machinery, and high wind, it is important to choose a quiet site in the ocean. Many sources of ambient ocean noise have been identified, and most of them can be avoided to some extent by careful siting.

High Winds (or High Wind-Driven Ocean Waves)

Ambient deep-ocean noise levels at frequencies from at least 0.5 to 30 Hz are strongly related to windspeeds on the ocean's surface. This relationship is evident from a study of ambient noise spectra on WIA hydrophone 74 and corresponding winds measured at Wake Island (Figure 12 from McCreery, 1988). The data were averaged from three-minute-long ambient noise samples taken once every six hours over a one-year period. From 0.5 to 6 Hz, they show that ocean-bottom noise increases regularly with increasing windspeed until a clearly defined saturation level is reached. However, the noise level variations and saturation are probably the more direct result of corresponding variations and saturation of waves on the ocean surface driven by the wind. The maximum variation in ambient noise with windspeed is 20 dB (a factor of 10) at around 1 Hz. Thus, a significant increase in the detectability of teleseismic P, with its dominant frequency at around 1 Hz, is possible in low-wind (low-wave) conditions at Wake. These data suggest that it should be possible to maximize the sensitivity of future ocean seismometers by choosing sites with low mean windspeeds. Many oceanic regions with low mean windspeeds exist (Figure 13) from McCreery, 1988), primarily at low latitudes, and they may be the quietest sites on earth for the detection of short-period seismic signals.

Coastlines

Ocean waves breaking along coastlines can be a significant source of ambient ocean noise. A comparison between the one-year-mean noise spectra of two WIA SOFAR hydrophones (Fig. 5) illustrates this point. Hydrophone 10, located only 3 km to the south of Wake, is 5-10 dB noisier at all frequencies (0.1-30 Hz) than hydrophone 20, located 200 km to the south of Wake in open ocean. In addition, the variation of the noise with windspeed is much different on hydrophone 10 compared to hydrophone 20 that is more similar to the deep hydrophones 74 and 76. Wake is a very small coral atoll with a radius of only about 2 km. For longer shorelines such as those on the edge of continents, the increase in noise may be significantly more pronounced. Additional measurements need to be made in order to quantify the levels and extent of this extraneous noise. It is clear, however, that sites near coastlines should be avoided whenever possible.

High Bottom Currents

Bottom currents can create at least two separate types of ocean noise: (1) noise caused by direct interaction of ocean-bottom sensor packages with the current, and (2) noise caused by current-induced turbulent pressure fluctuations on the ocean floor. The first type of noise was a problem for many years with the HIG POBS. The POBS package, with its sensors inside, had a small anchor on the bottom and floats on the top. It was an inverted pendulum that rocked back and forth in the current. In addition, the antennas for the recovery radio transmitters vibrated whenever currents swept by them.

These extraneous motions were picked up by the sensors, and often they masked signals of interest. This problem was solved to some extent in the ISOBS design by making a separate low-profile sensor package and isolating it from the recorder and recovery systems. However, current noise of this type is still occasionally seen on the ISOBS. Any type of burial of the sensor package should eliminate this noise. The second type of noise is present most notably on horizontal sensors. When a current-induced cell of turbulence passes over the ocean bottom, it can create tilts due to the pressure differences on opposite sides of the cell. These tilts manifest themselves as horizontal accelerations due to corresponding changes in the gravity component as the sensor is tilted. This type of noise is more carefully described in a paper by Webb (1988). Although this type of noise can be a significant problem on the ocean floor, it can probably be reduced to negligible amounts by burying the sensor package a few meters (the turbulent cell dimension) below the interface. Additional data will be required to verify this solution. In any case, it is probably best to avoid siting ocean seismometers in regions with high bottom currents.

Shipping Lanes

Surface ships are a considerable source of noise in the ocean, even at frequencies from 1-30 Hz. Data from the WIA hydrophones, located far from commercial shipping lanes, are nevertheless contaminated by a significant amount of shipping noise. Over a one-year period, the data from WIA hydrophone 74, at 5.5-km depth, were found to be contaminated by shipping noise in varying amounts that increased regularly with the signal frequency (Figure 14). At 1 Hz, only about 1 percent of the data are contaminated, but at 20-30 Hz nearly 8 percent are contaminated. This contamination would be much worse near any shipping lane. And although the multiple narrow-band noise typically produced by a ship may be removed to some extent by careful filtering, it can prevent many types of modern signal analysis that require exact waveform data or spectral analyses of broadband signals. Thus, the level of shipping traffic should be given careful consideration before choosing a site for an ocean seismic station.

Hurricane and Typhoon Lanes

Although the low latitudes are generally advantageous for reducing noise because of low mean windspeeds, they can sometimes be disadvantageous because of the occurrence of hurricanes and typhoons. These severe storms produce locally high levels of noise at 1 Hz and above due to their high windspeeds and subsequent high waves as described above. In addition, they can produce high noise levels at frequencies of 0.2 Hz and below that propagate to a radius of a thousand kilometers or more (Figure 15). Since hurricanes and typhoons often form and travel within certain known lanes, it is prudent to avoid these lanes if possible.

Whale Migration Routes

Some unusual noises that appear regularly in the data from WIA are high-amplitude pulses, one or two seconds in length, that repeat every 20-30 seconds. They have frequencies ranging from 17-20 Hz, with signal-to-noise ratios often many tens of decibels, and the pulses sometimes continue for many days. They are present most often between February and April, but are seen during the rest of the year as well. These signals are presumed to be from whales, and they have been observed in many regions of the Pacific. This noise is above the frequency range of teleseismic P, but not above the range of oceanic seismic phases Po, So, or T, and not above the range of explosive seismic data. It may therefore interfere with those types of signals to some extent. Migration routes for these whales should be avoided if signals in the 17-20 Hz band are to be recorded and studied.

SUMMARY

Many factors should be considered in the design and siting of long-term, broadband seismic instrumentation in the oceans. Sensors located in the SOFAR channel are ideal for detecting energetic waterbourne signals at long distances. Sensors located on the ocean bottom are more sensitive to refracted signals such as P because of reduced noise, especially if the bottom is below the SOFAR channel. But noise on the ocean deep bottom may still be high if the sensor package is exposed to currents, and poor coupling of the sensors to the sediments may severely degrade the recorded particle motions. Careful design of the sensor package, and possibly a shallow burial of the sensor package should greatly reduce or eliminate these problems. Drillhole sensors located far below the water-sediment interface have clear advantages in terms of signal-to-noise ratios and coupling, but the cost of deployment is very high. Sensors should be located in regions that minimize the sources of extraneous ambient noise. These sources include: (1) high winds, (2) coastlines, (3) high bottom currents, (4) shipping, (5) hurricanes, and (6) whales.

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FIGURE CAPTIONS

Figure 1. The ISOBS isolated sensor package deployment. (A) The ISOBS at the end of its free-fall to the ocean bottom. (B) After a few hours, a magnesium-wire release activates, and the sensor package is pushed by an elastic cord away from the rest of the ISOBS. (C) The sensor package free falls away from the main package, pivoting on a hinge bar. At an angle of about 60 degrees, the hinge bar is restrained by a small wire and the sensor package tubles free. (D) The ISOBS in its fully deployed configuration.

Figure 2. The various components of a deployed OSS system include: (A) the sensor package locked in the drillhole, (B) an electromechanical data transmission cable, (C) recorder package, (D) positively buoyant recovery rope, (E) recovery buoy, (F) anchor, and (G) recovery ship.

Figure 3. Location map for hydrophones of the Wake Island Array.

Figure 4. A typical section of helicorder record from a SOFAR channel hydrophone. This particular record was made on June 26, 1988 from Wake hydrophone 10. The time between tic marks is one minute, and the time between adjacent lines if 30 minutes.

Figure 5. The one-year mean noise spectra of Wake hydrophones 76, 10, and 20 plotted relative to the one-year mean noise spectrum of hydrophone 74 (zero dB).

Figure 6. Ambient deep-ocean noise spectra for the Wake Island Array (WIA) hydrophones 74 and 20, for a hydrophone bottomed off Eleuthera Island at 1200 m depth, and for a hydrophone bottomed off Bermuda at 4300 m depth (Bermuda and Eleuthera data from Nichols, 1981). Also shown is average continental seismic noise (Brune and Oliver, 1959), and a very quiet continental noise measurement at Lajitas, Texas (Herrin, 1982).

Figure 7. ISOBS data from an airgun line off of the Kuril Islands. Data from the three different sensors of the ISOBS -- hydrophone (a), vertical (b), and horizontal (c) -- are shown. Arrivals "C" and "G" are respectively the direct and first-multiple water-wave arrivals. Arrivals "A" and "B" are crustal P and S phases, respectively, that arrive at the ISOBS as a compressional energy through the sediment. Arrivals "D" and "E" are crustal P and S phases, respectively, that arrive at the ISOBS as shear energy. Arrival "F" is a basement reflection that has converted from compressional to shear at the sediment/basement interface.

Figure 8. The OSS borehole sensor package (tool).

Figure 9. Signal from a ship passing nearby to OSSIV. Smoothed and unsmoothed rectified traces of the signal at 20 Hz are shown for the E-W, N-S, and vertical components. The smooth curve shows the theoretical amplitude function for a ship on a course of 120 degrees at a speed of 30 km/hr.

Figure 10. The 26 May 1983 Japan earthquake recorded by the OSSIV geophones. The spiky arrivals before the earthquake are from small explosive charges.

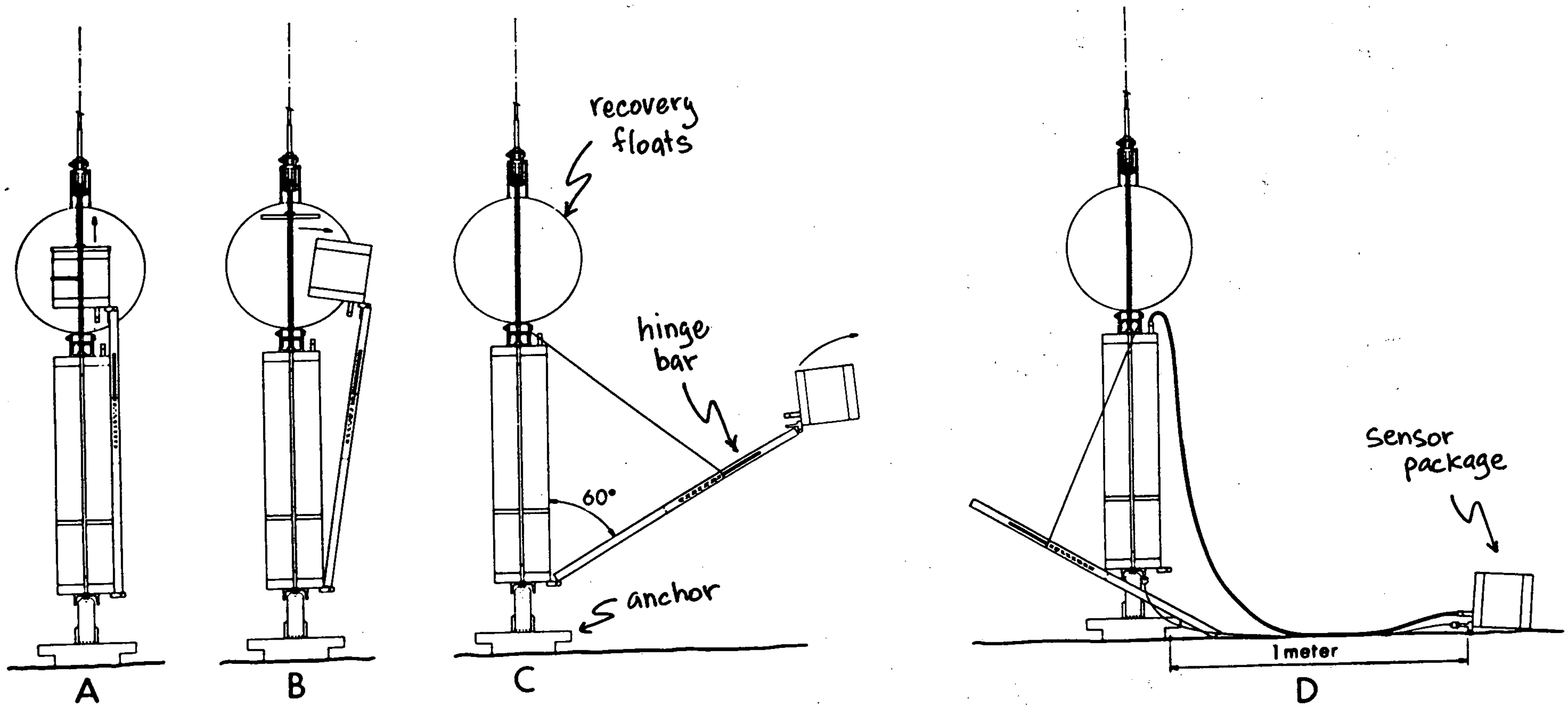
Figure 11. A comparison between simultaneous noise levels recorded by OSSIV, by a nearby ISOBS, and by an Oregon State University (OSU) ocean bottom seismometer. On the left (a) are the vertical components and on the right (b) are the horizontal components.

Figure 12. The average noise spectra of Wake hydrophone 74 for eight windspeed ranges (from McCreery, 1988). Noise levels increase regularly with windspeed at all frequencies, and two types of noise are observed. The first noise type, between about 0.5 and 6 Hz, increases with windspeed until a saturation level (arrowhead) is reached. The second noise type, from about 4 Hz to more than 30 Hz, increases with windspeed for winds above 12-16 mph.

Figure 13. Eighty-year-mean ocean-surface windspeed compiled by the University of Hawaii's Department of Meteorology (from McCreery, 1988). The contour interval is 1 m/s (2.24 mph), and shaded regions have mean windspeeds less than 4 m/s (8.95 mph). The one-year-mean windspeed measured at Wake is 6.26 m/s (14 mph).

Figure 14. The percent of noise samples over a one-year period on Wake hydrophone 74 that were contaminated by shipping noise, plotted as a function of frequency (from McCreery, 1988). The feature at 20 Hz is an artifact due to a 60 Hz aliased signal in the original data.

Figure 15. Spectra from Wake hydrophone 74 during the passage of Typhoon Owen. Noise levels at 0.13 Hz were highest on 21 October, when the typhoon had sustained winds of 100 kts but was still more than 1000 km from Wake. Noise levels at frequencies above 0.2 Hz were highest on 24 October, when Owen made its closest approach to Wake (about 500 km) with windspeeds of 50 kts. Velocities noted by each spectrum are the daily-mean windspeeds at Wake.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1 meter
A
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D

Figure 1

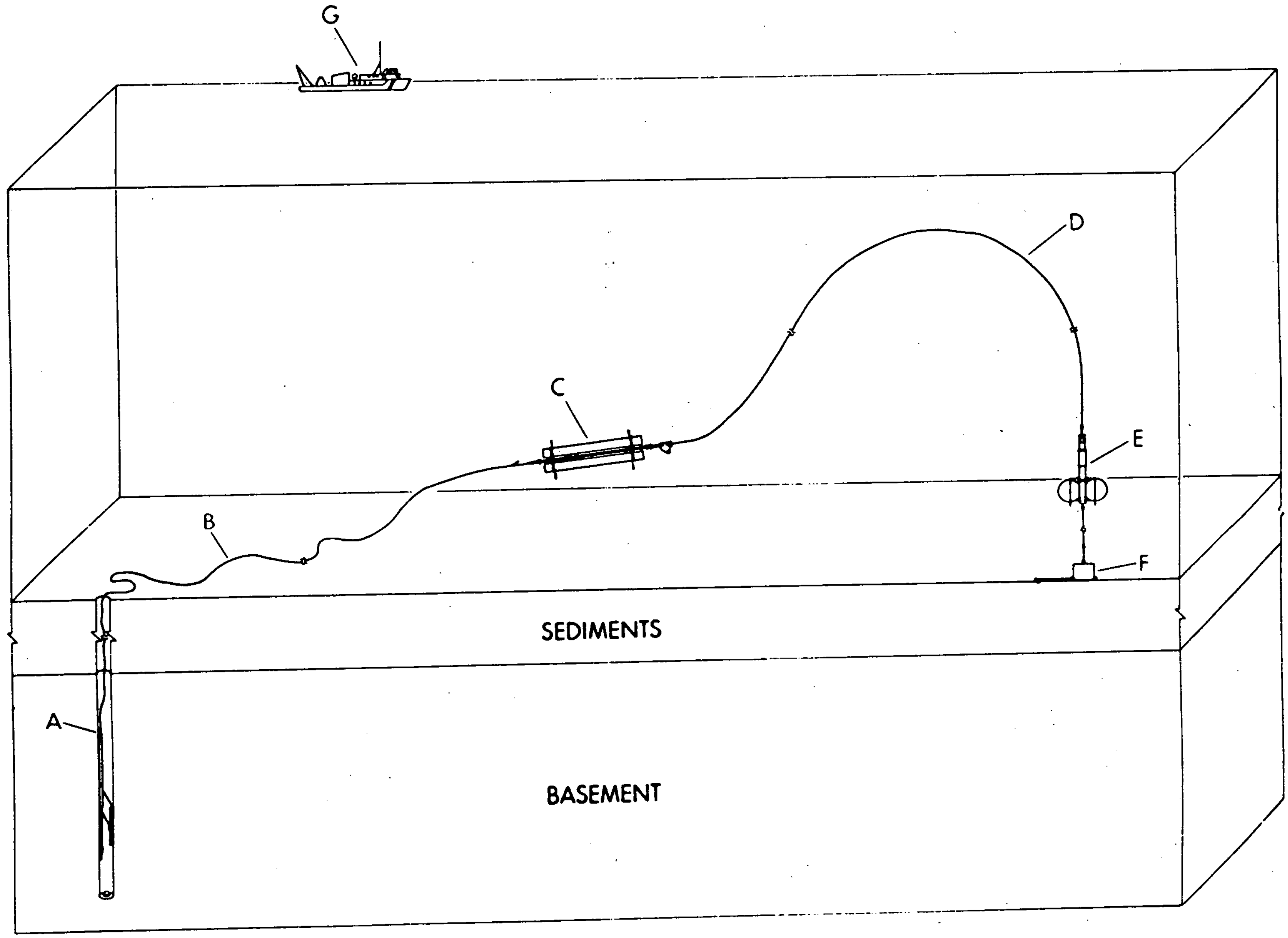


Figure 2

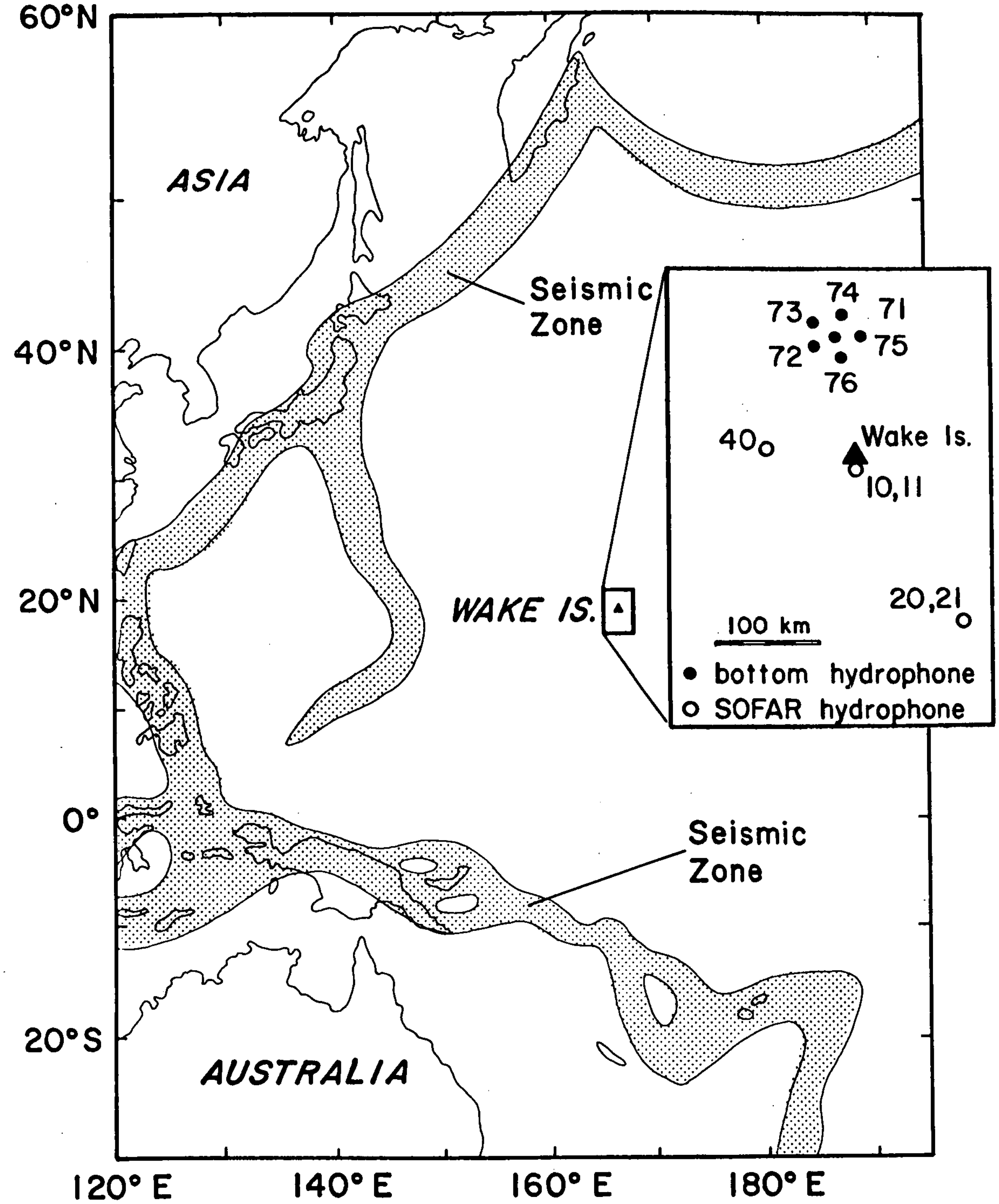

60°N
ASIA
Seismic Zone
40°N
20°N
WAKE IS.
0°
Seismic Zone
20°S
AUSTRALIA
120° E
140° E
160° E
180° E
74
73
71
72
75
76
40
Wake Is.
10,11
20,21
100 km
bottom hydrophone
SOFAR hydrophone

Figure 3

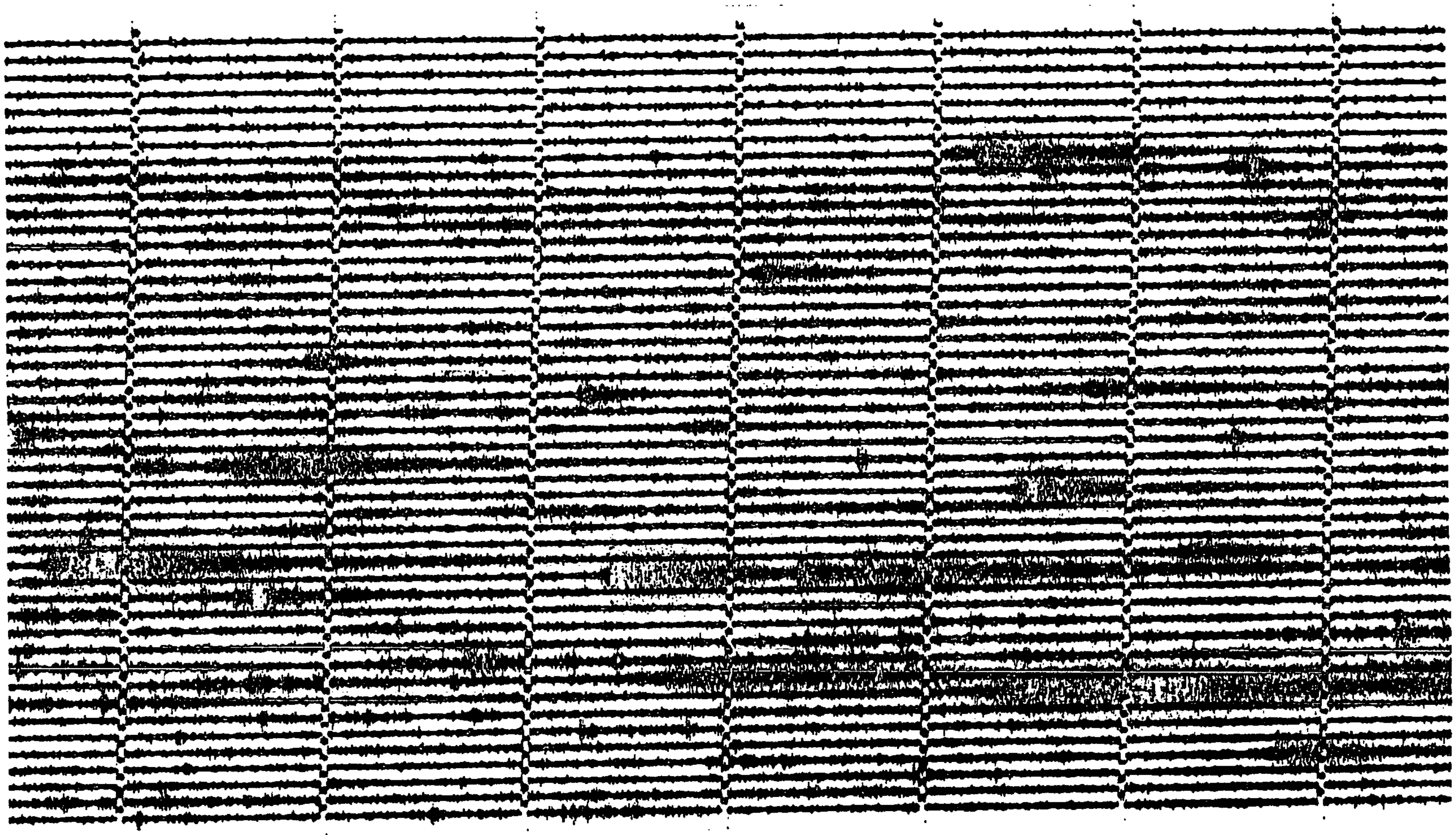


Figure 4

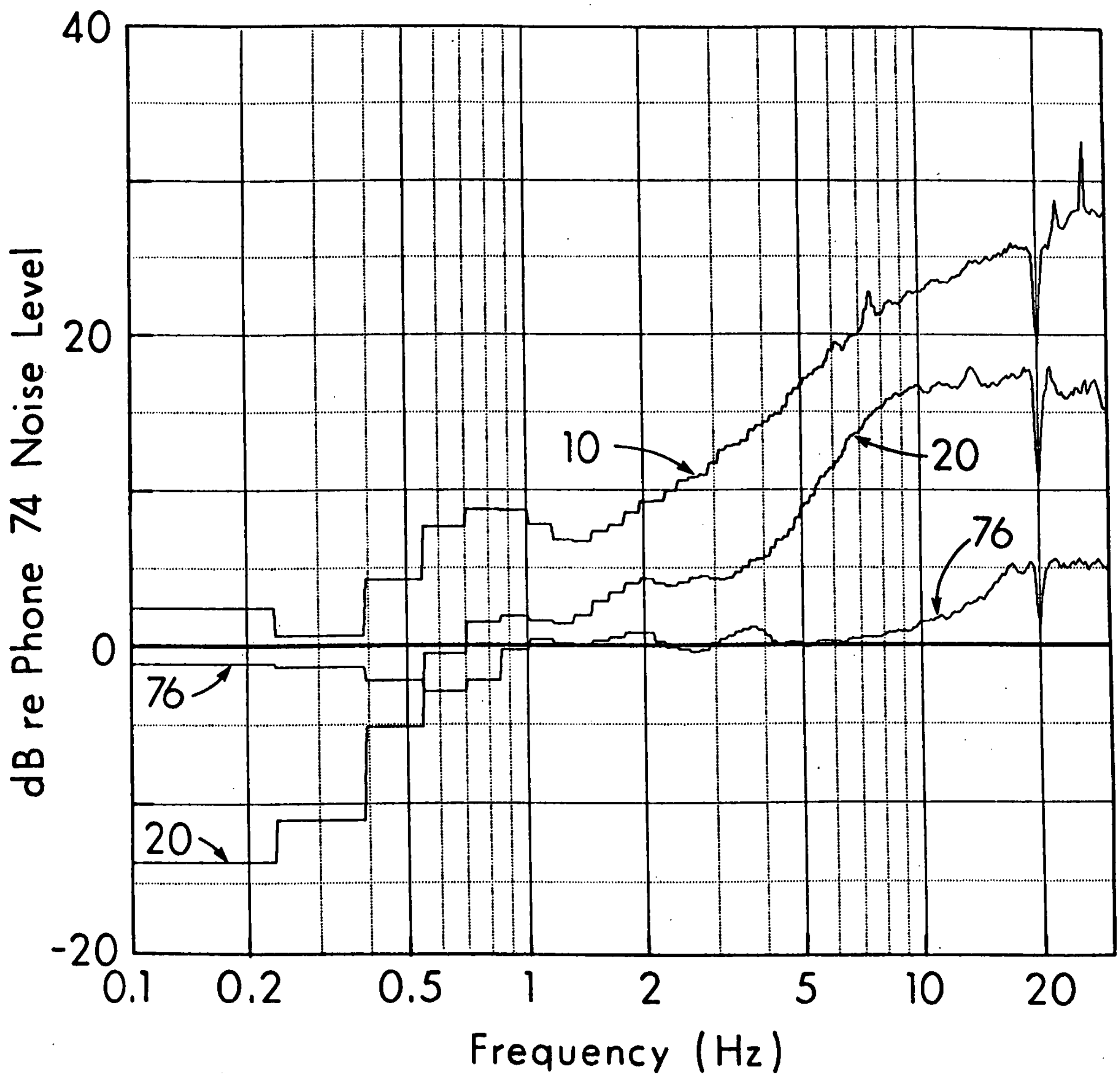
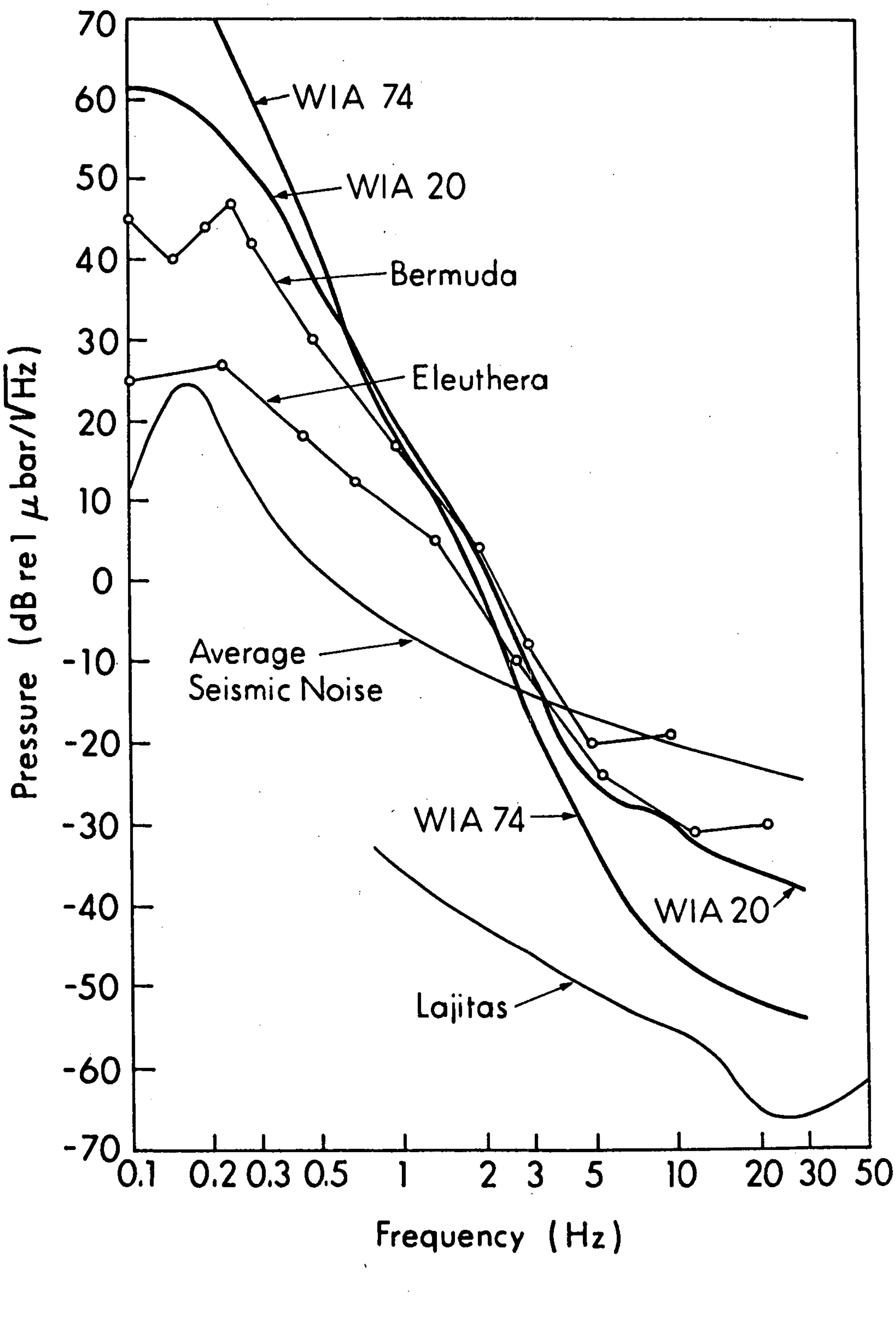

40
20
0
-20
dB re Phone 74 Noise Level
10
20
76
76
20
0.1
0.2
0.5
1
2
5
10
20
Frequency (Hz)

Figure 5



Pressure (dB re 1 μbar/√Hz)
70
60
50
40
30
20
10
0
-10
-20
-30
-40
-50
-60
-70
0.1
0.2
0.3
0.5
1
2
3
5
10
20
30
50
Frequency (Hz)
WIA 74
WIA 20
Bermuda
Eleuthera
Average Seismic Noise
WIA 74
WIA 20
Lajitas

Figure 6

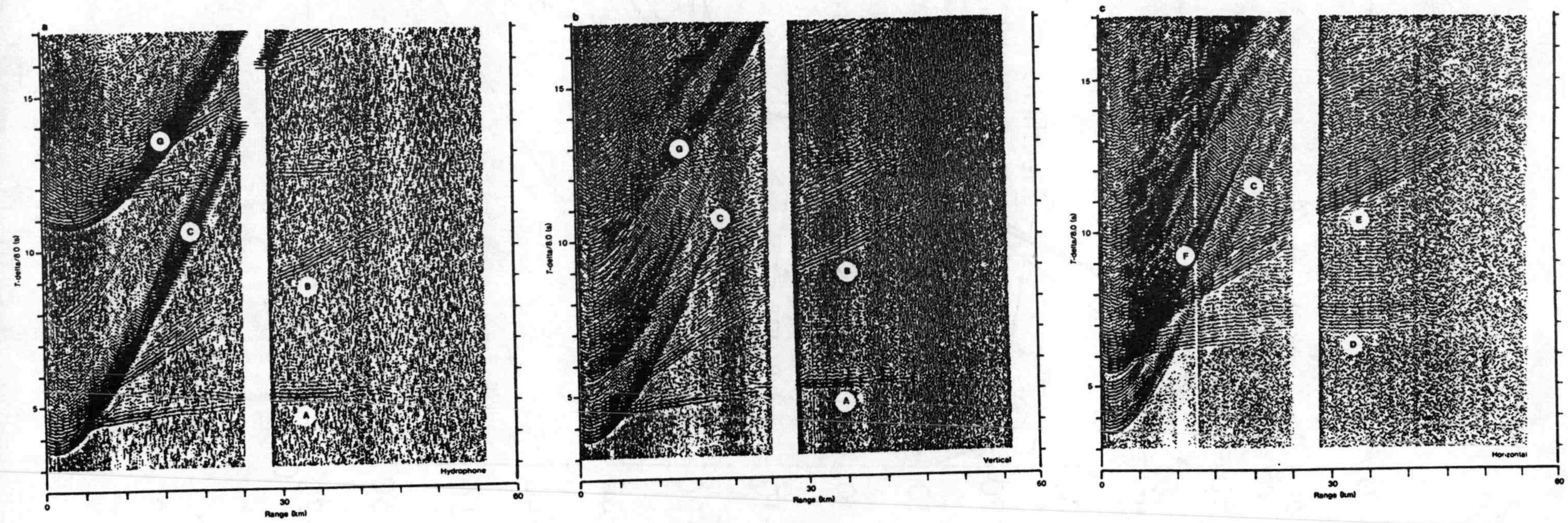
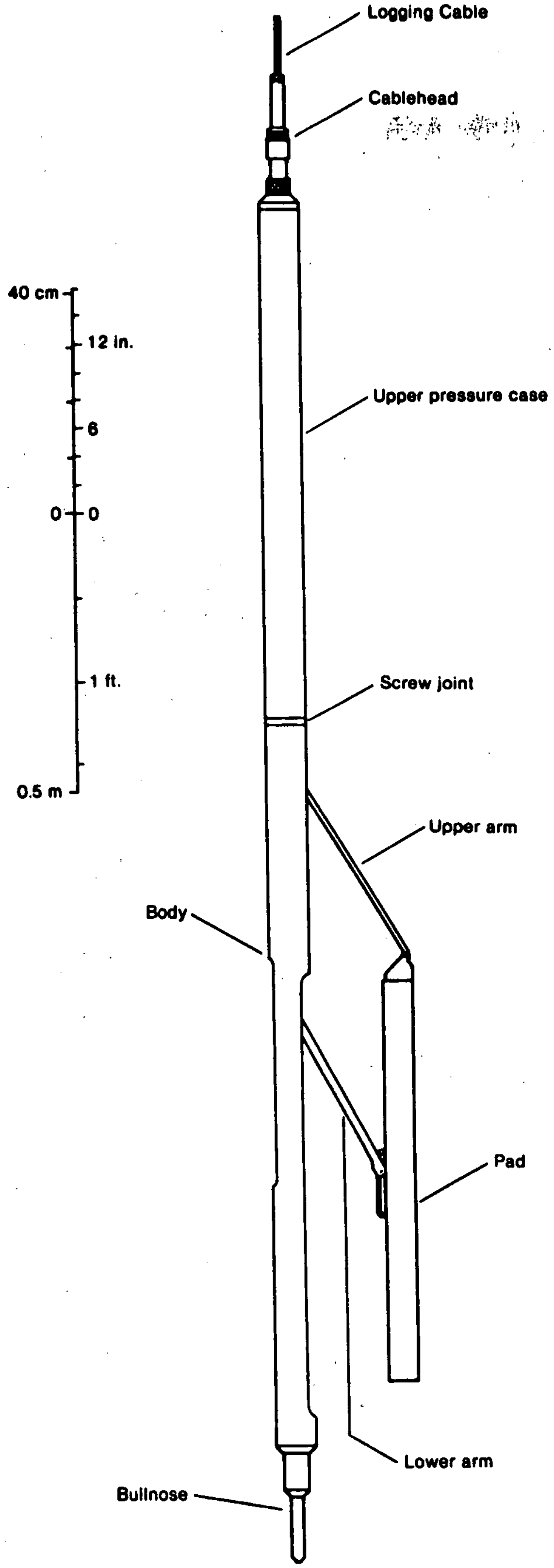


Figure 7



Borehole package (tool) as installed on OSS IV.

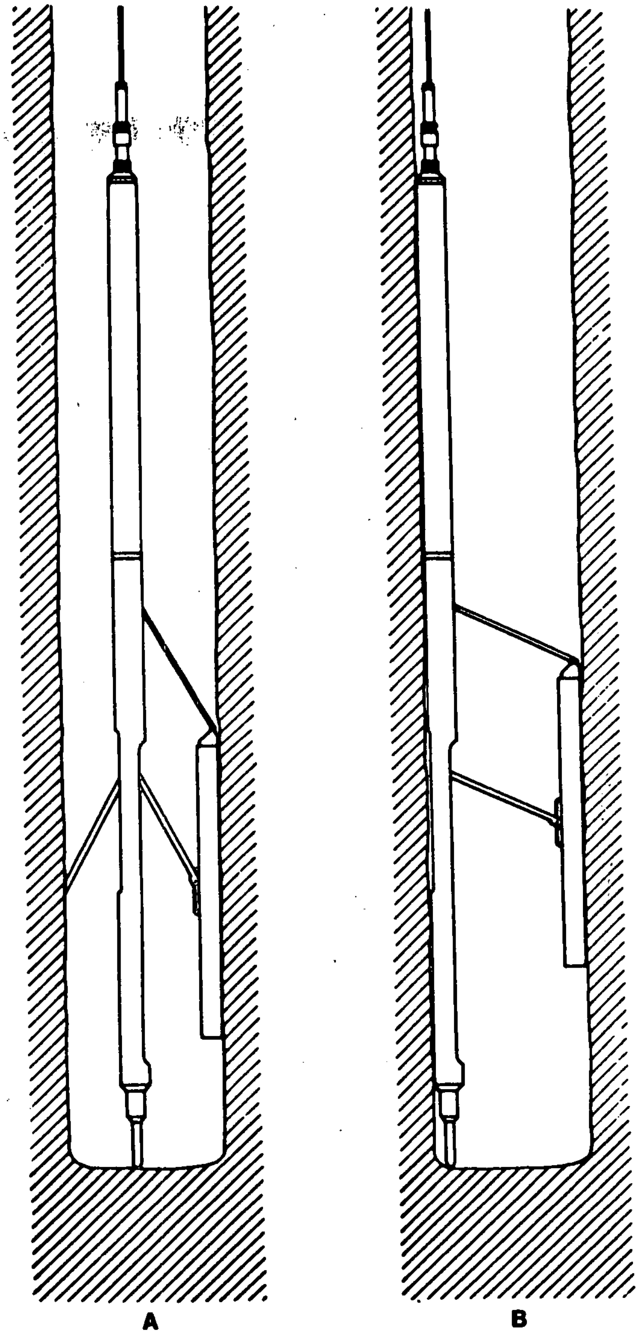


Figure 4 Modification of borehole package. A. Original package as supplied by Gearhart Industries. B. HIG modification to improve seismic coupling to borehole.

Figure 8

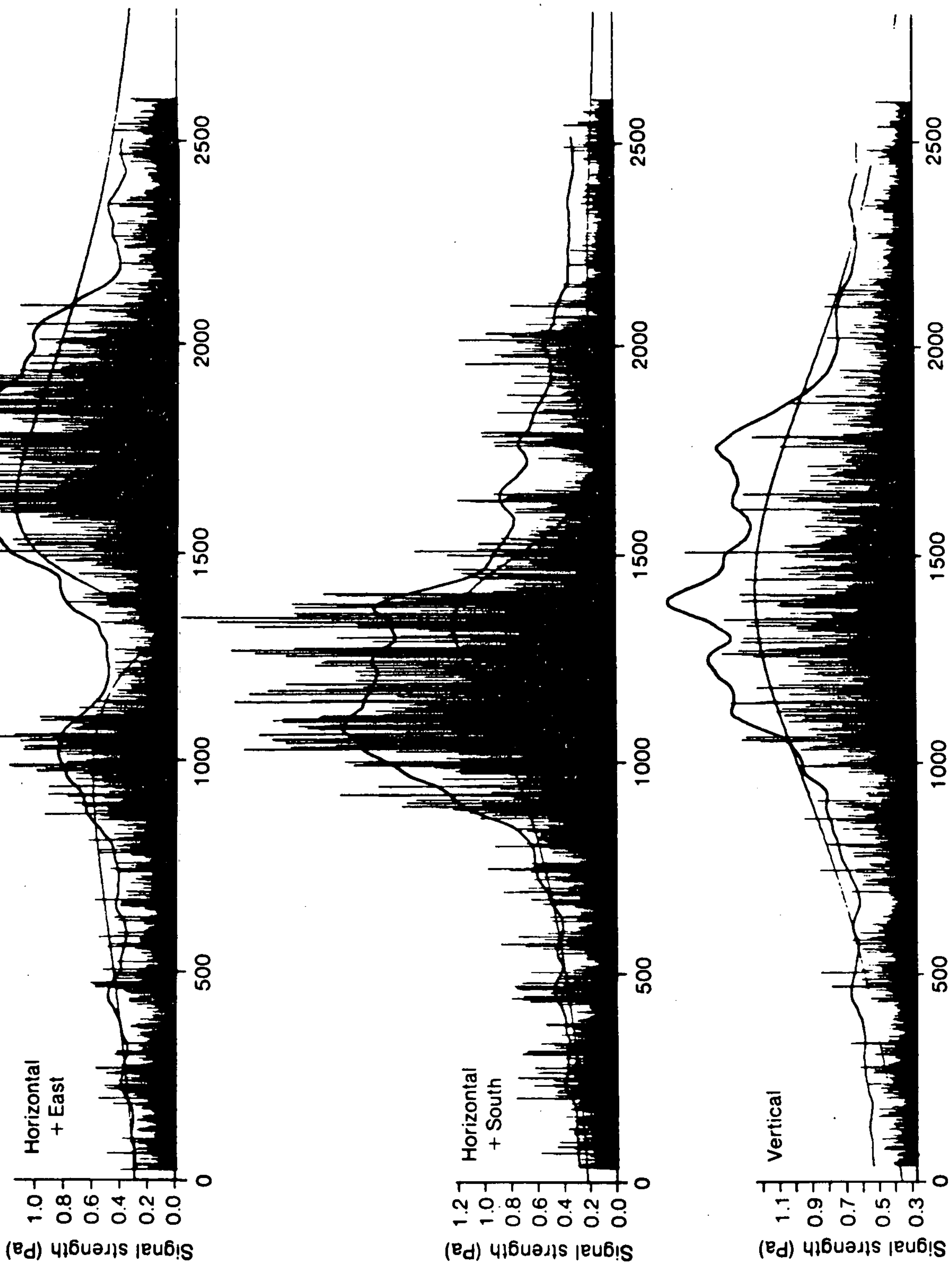

Horizontal
+ East
Horizontal
+ South
Vertical
Signal strength (Pa)
Signal strength (Pa)
Signal strength (Pa)
Time (s)
0
500
1000
1500
2000
2500
1.0
0.8
0.6
0.4
0.2
0.0
1.2
1.0
0.8
0.6
0.4
0.2
0.0
1.1
0.9
0.7
0.5
0.3

Figure 9

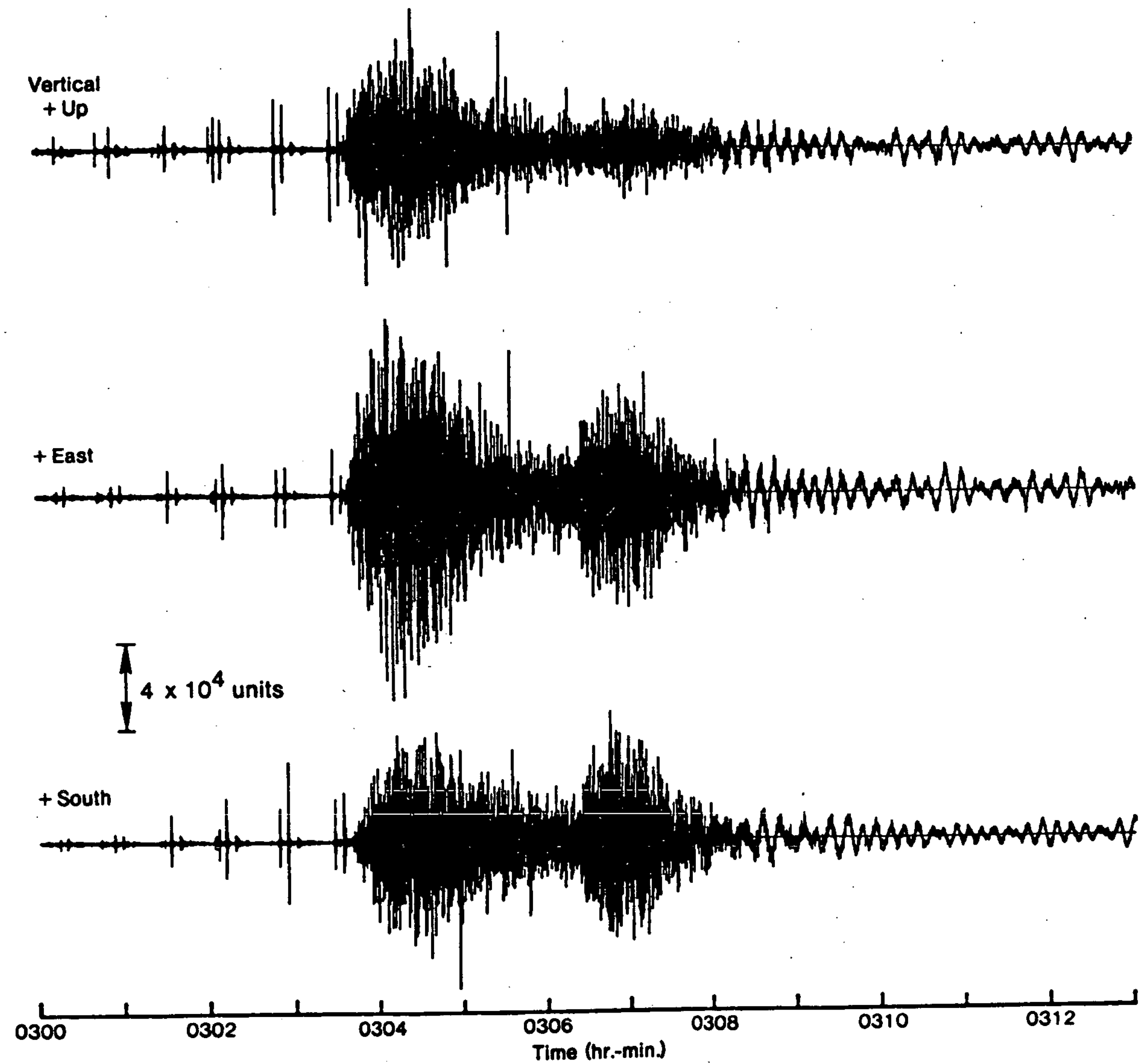
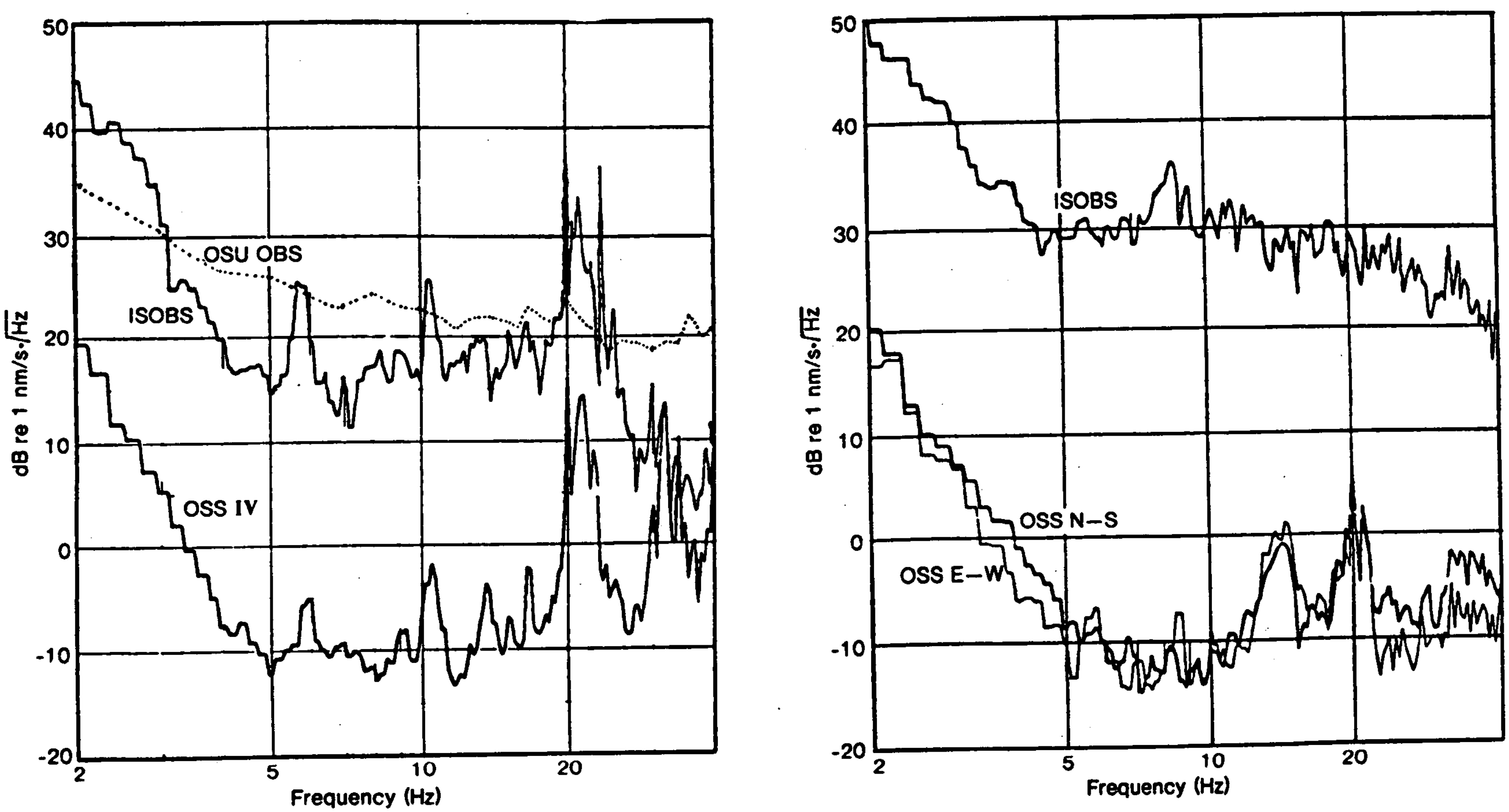

Vertical
+ Up
+ East
4 x 10^4 units
+ South
0300
0302
0304
0306
0308
0310
0312
Time (hr.-min.)

Figure 10



50
40
30
20
10
0
-10
-20
dB re 1 nm/s·$\sqrt{Hz}$
OSU OBS
ISOBS
OSS IV
2
5
10
20
Frequency (Hz)
ISOBS
OSS N—S
OSS E—W

Figure 11

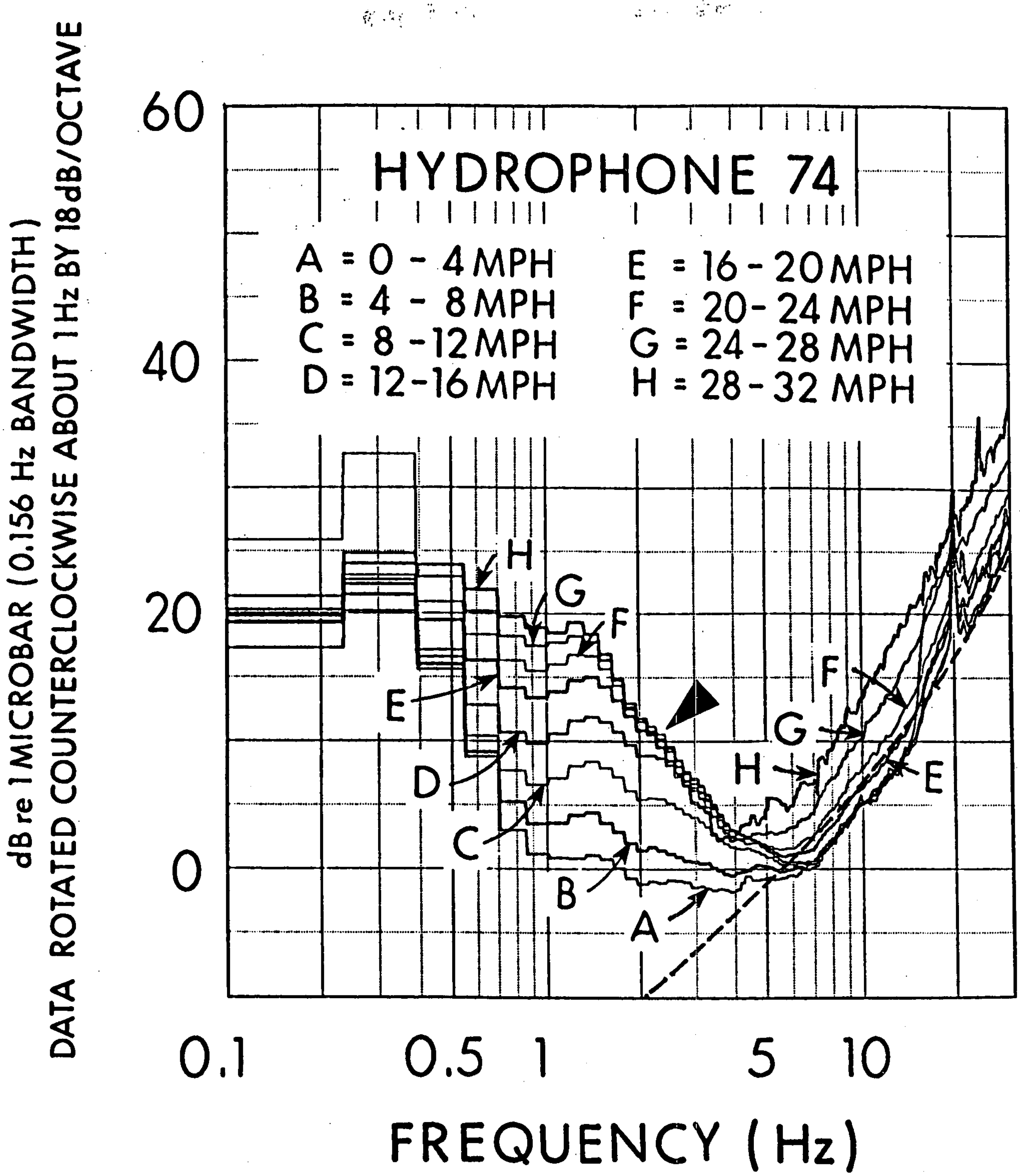

HYDROPHONE 74
A = 0 - 4 MPH
B = 4 - 8 MPH
C = 8 -12 MPH
D = 12-16 MPH
E = 16 - 20 MPH
F = 20- 24 MPH
G = 24 - 28 MPH
H = 28 - 32 MPH
dB re 1 MICROBAR (0.156 Hz BANDWIDTH)
DATA ROTATED COUNTERCLOCKWISE ABOUT 1 Hz BY 18 dB/OCTAVE
60
40
20
0
0.1
0.5
1
5
10
FREQUENCY (Hz)
H
G
F
E
D
C
B
A
F
G
H
E

Figure 12

80-YEAR MEAN OCEAN SURFACE WIND SPEED

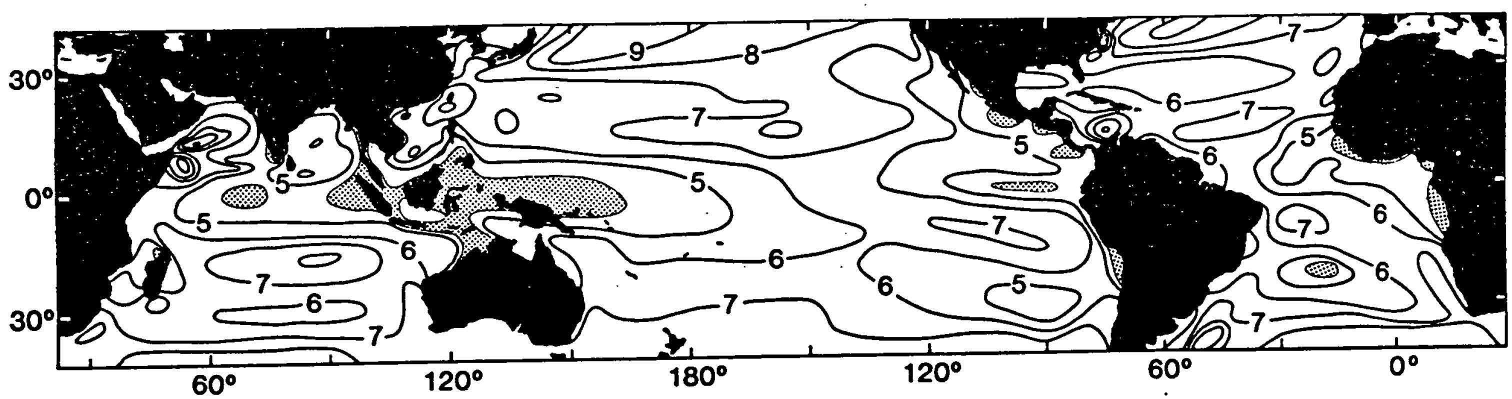


Figure 13

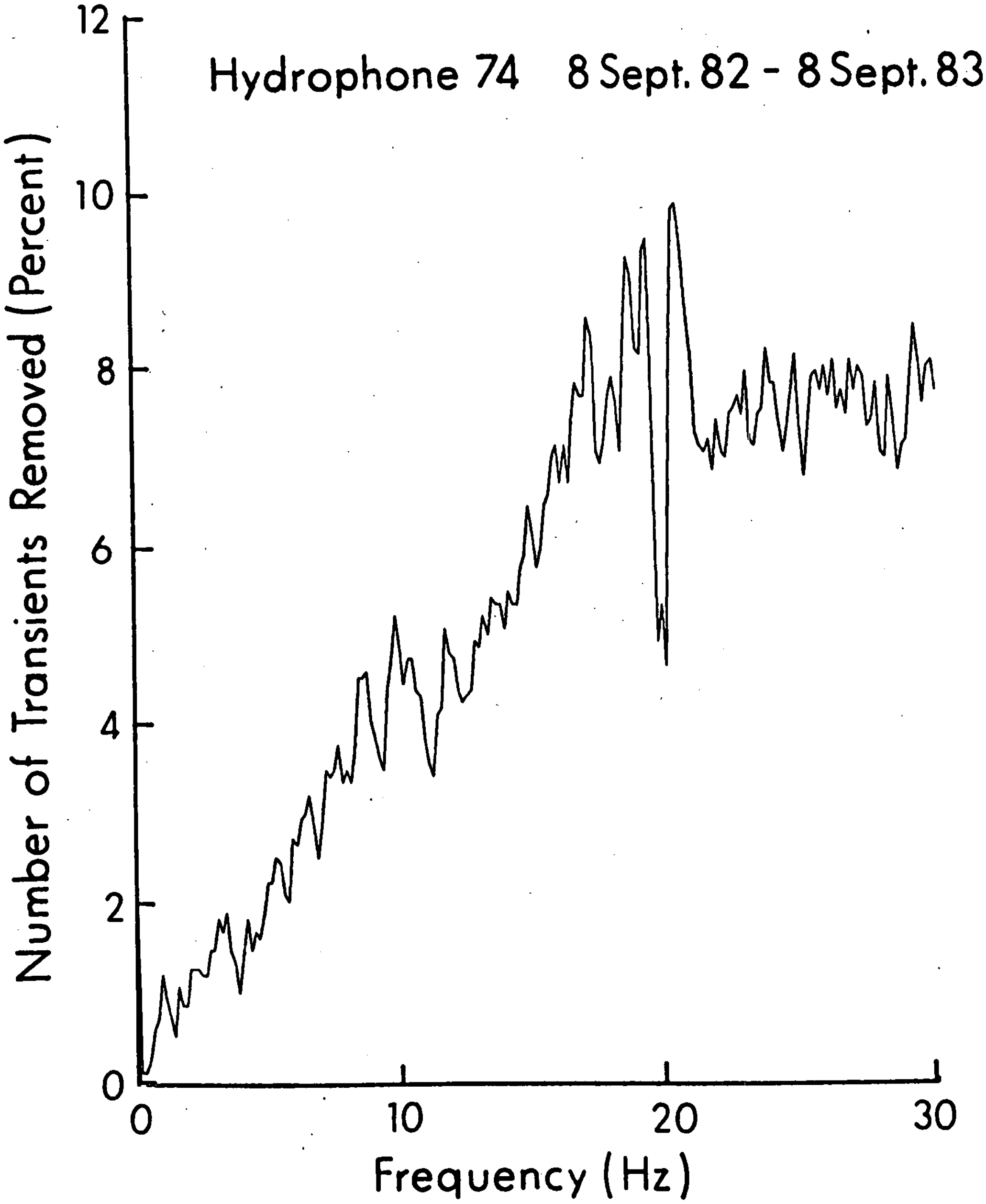

Hydrophone 74 8 Sept. 82 - 8 Sept. 83
Number of Transients Removed (Percent)
Frequency (Hz)
0
2
4
6
8
10
12
0
10
20
30

Figure 14

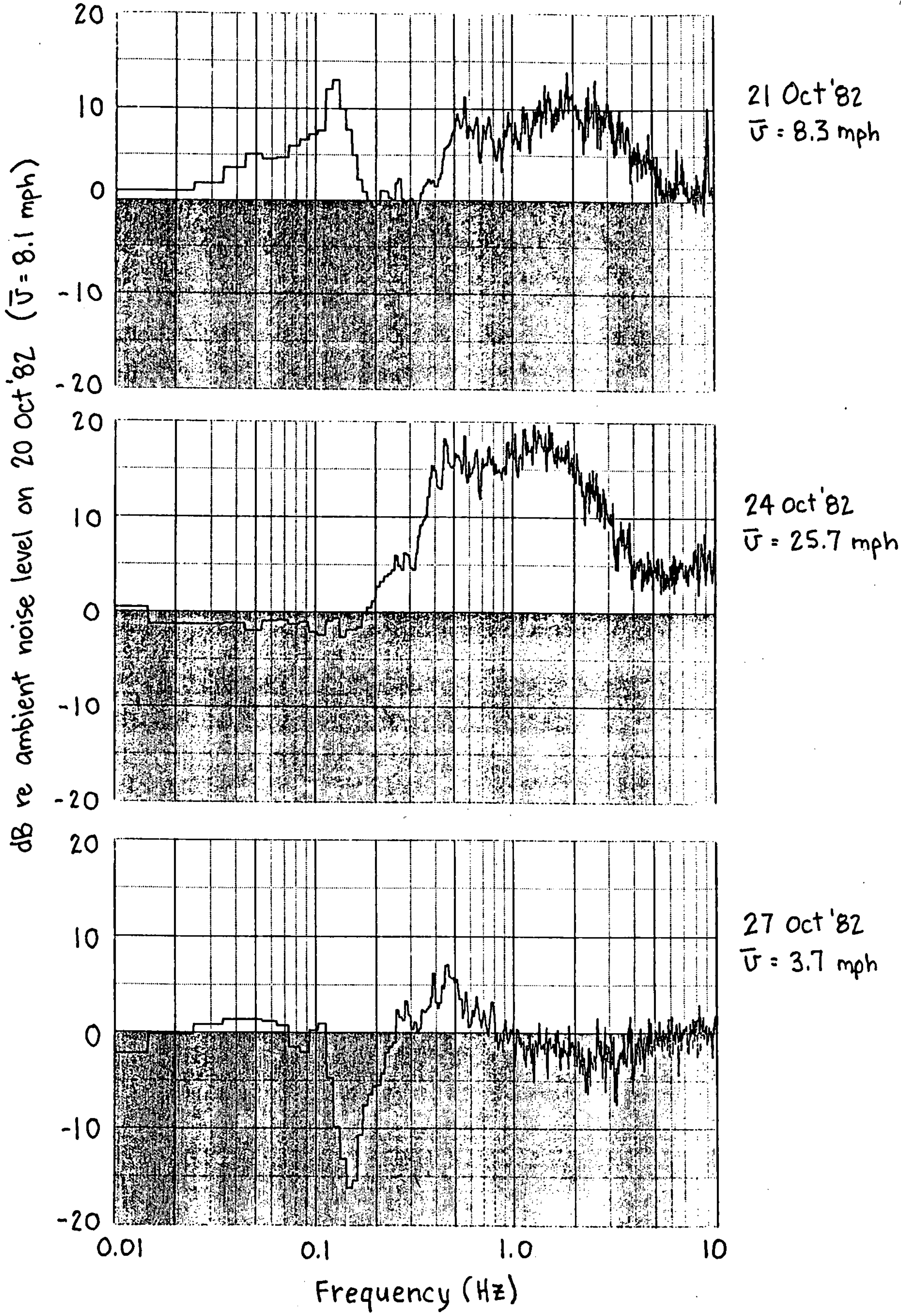

dB re ambient noise level on 20 Oct '82 ($\overline{U}$ = 8.1 mph)
20
10
0
-10
-20
21 Oct '82
$\overline{U}$ = 8.3 mph
24 oct '82
$\overline{U}$ = 25.7 mph
27 oct '82
$\overline{U}$ = 3.7 mph
0.01
0.1
1.0
10
Frequency (Hz)

Figure 15

Section B5: Instrumentation

1. Broad-band Seismometers for Seabottom Operation: E. Wielandt and J.M. Steim
2. Some Comments on Deep Ocean Downhole Seismometer Emplacement: F.N. Spiess
3. Pilot Borehole Seismic Experiments: J.D. Phillips
4. Trans-Oceanic Telecommunications Cables: Re-Use for Ocean Bottom Geoscience Observatories: S. Nagumo and D. A. Walker
5. Satellite Telemetry in Oceanography: D.E. Frye and M.G. Briscoe
6. Wireline Re-Entry of Boreholes in the Deep Ocean: J. LeGrand
7. PASSCAL - A New Generation of Portable Seismic Instrumentation: R.A. Phinney and J.C. Fowler
8. Long-Term Recording of Seismic Data on the Ocean Floor: G.M. Purdy
9. The Global Seismic Network Data Acquisition System: A. Dziewonski

BROAD-BAND SEISMOMETERS FOR SEABOTTOM OPERATION

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**Quanterra, Inc., Shirley, MA 01464

Broad-band force-balance seismographic systems (sensor, A/D converter and recorder) for surface installation have achieved an operating range from the level of minimum expected teleseismic signal levels, in a frequency band covering the whole teleseismic spectrum (approximately from 0.3 mHz to 10 Hz). No comparable seismographic system exists for the marine environment.

For installation at or in the seabottom, a seismometer must have a safe, remotely operated mass-locking mechanism since rough handling may occur during deployment. None of the presently available broad-band seismometers has such a mechanism. A deep-sea borehole seismometer must also have a remotely operated levelling mechanism and a holelock. The design of these mechanisms is as difficult as the design of the sensor itself. Designing a completely new sensor for deep-sea borehole deployment is probably easier than adding the required mechanisms to an existing sensor.

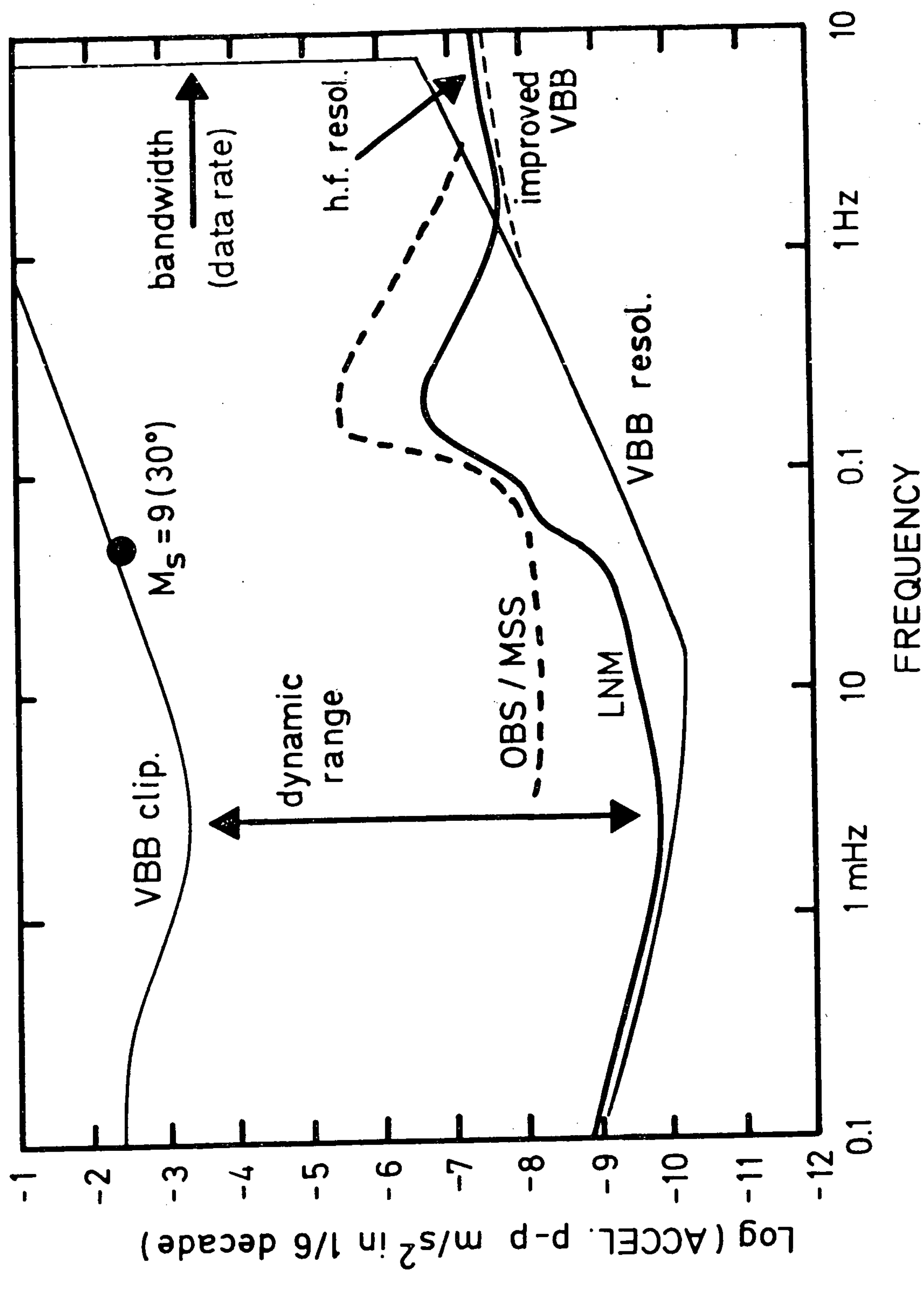
External power will only under exceptional circumstances be available at the seabottom. Low power consumption is therefore a primary design goal. It will impose limitations to the performance in three areas: instrumental noise level, maximum on-scale signal, and bandwidth. Present broad-band sensors consume in the order of one watt per component. For long-term, deep-sea deployment, the power consumption should be at least ten times lower. This will preclude the use of integrated operational amplifiers optimized for low noise. However, since the seismic noise level even in boreholes seems to be considerably higher at the seabottom than at quiet land sites, a somewhat elevated instrumental noise may be acceptable. We believe that the requirement to minimize the power consumption will not seriously compromise the performance of a deep-sea broad-band seismograph, and that basically the same electronic circuits (although not with the same semiconductor components) could be used as in land-based systems.

The situation is comparable for the digitizer. It would at present be difficult to maintain 24-bit resolution at a power consumption of 0.1 watt, but a resolution of 20+ bits at a 20 Hz sampling rate could be achieved with presently available low-power components. The overall bandwidth of the system will probably be more restricted by limitations in data storage and retrieval than by the available electric power.

In summary, we believe that the construction of a deep-sea broad-band seismograph with a low power consumption will not pose problems of a fundamental nature. It is however a major engineering task that should be planned in time.

Figure 1

Design criteria for land- and sea-based broad-band seismographs. A broad-band seismograph should resolve seismic ground noise in the entire (tele)seismic frequency band. Minimum ground noise at land sites is given by the solid line "LNM" (Low Noise Model by Peterson and Tilgner, 1985). Reliable information on minimum seismic noise in boreholes at the seabottom is not available at this time; the broken line "OBS/MSS" is a tentative compilation of the partly indirect evidence presented by other authors at this meeting. As the maximum signal of interest for stations in a global network one usually considers the surface waves from a magnitude 9 earthquake at 30 distance (dot). State-of-the-art VBB (very-broad-band) equipment has a dynamic range just sufficient to record minimum and maximum signals of interest over the whole band; some improvements in high-frequency response and resolution remain desirable. The requirement to use low-power components would affect the dynamic range, the bandwidth, and the resolution at high frequencies (arrows). The diagram suggests that ocean-bottom seismographs could be designed for about 10 times less resolution and a 10 times smaller dynamic range compared to the VBB system while their operating range could be the same. Data storage and transmission problems will probably impose more serious limitations on the overall bandwidth than sensor design.



Log (ACCEL. p-p m/s^2 in 1/6 decade)
-1
-2
-3
-4
-5
-6
-7
-8
-9
-10
-11
-12
0.1
1 mHz
10
0.1
1 Hz
10
FREQUENCY
VBB clip.
M_S = 9 (30°)
bandwidth
(data rate)
dynamic
range
OBS / MSS
LNM
VBB resol.
h.f. resol.
improved
VBB

Figure 1

SOME COMMENTS ON DEEP OCEAN DOWNHOLE SEISMOMETER EMPLACEMENT

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La Jolla, CA 92093

My interests in a program such as might emerge from this workshop are primarily in the operational support aspects, with the added possibility of geodetic research that could benefit from the fact that one might repeatedly visit some sites in order to replace power supplies or repair or replace sensors, communication nodes, etc.

The first order involvement would have to do with the capability to reconnoiter existing holes to determine their status, and to install equipment in those deemed appropriate for use. A closely related capability would have to do with operations on the seafloor to connect bottom laid communication links. These possibilities can best be discussed in terms of three devices, thruster, re-entry probe, RUM III all operable from our larger conventional research ships, particularly (but not solely) those having good position keeping ability (KNORR, MELVILLE, AGOR 23). All three of these devices are under development and, judging by the history of our Deep Tow system, will undoubtedly evolve continuously throughout their useful lives. They all depend on use of 0.68" electromechanical cable.

The Thruster (Fig. 1) is the farthest along, having been used for precision placement of OBS units in April, 1987, and as the deploying unit for our seafloor gravity meter in March, 1988. This is a unique type of ROV, relying on the cable and winch for vertical control and on an active thruster/deflector combination to provide horizontal positioning. Since it is supported by the 0.68" cable, it is capable of handling loads of the order of a few thousand pounds in deep water. It is powered from the ship, using 2400 V, 60 Hz, up to 6 A, transmitted down the coax core of the electromechanical cable. The cable has about 1 MHz of useful bandwidth available for up or down-link telemetry, some of which is used for control and status signals, transponder navigation, up and down echo sounders, scanning sonar, and compass. The remainder of the bandwidth is available for data telemetry and operation of research payload either on or suspended below the vehicle.

At the vehicle the 2400 V power is transformed to 220 V for propulsion and various DC voltages to supply electronic packages. Propulsion is provided by a 5 HP, 1800 rpm, constant speed electric motor directly coupled to an Innerspace Co., ducted propeller having variable pitch blades to provide thrust magnitude control. A pair of flaps just outboard of the discharge end of the duct provide a deflection force, and thus a turning moment that allows us to direct the primary thrust in whatever direction may be required.

Typical on-board payloads have been a CTD, strobe light, cameras and snapshot (slow scan) television. Telemetry modifications are underway to provide a 32 k baud link to work with a television equipped borehole re-entry probe and and a set of downhole geophones.

Construction has just started on the re-entry probe, designed to operate with the thruster. It will be a slim instrument (less than 15 cm diameter) and about 3 m long. It will include two 3 arm calipers with 60 cm maximum spread, a receiving hydrophone for transponder navigation coordinated with the thruster, a slow scan television and surrounding lights, a tiltmeter and a pressure gauge to warn if the unit is being held up by some obstruction in the hole. This unit will be satisfactory for operation with the thruster to log the condition of any holes prior to inserting long term measurement seismometers, strain meters, etc. If a systematic program of hole reconnaissance were undertaken it would be desirable to add other logging devices (e.g., televiewer, gravimeter) that might not have been used during drill-ship operations. The February 1987, wireline re-entry workshop contains recommendations relevant to this aspect of the program. At this state of development it is anticipated that the probe would be operated by hanging it with an appropriate length of coax well logging cable below the thruster rather than using a winch. This is a simple matter of minimizing cost and complexity until we have more experience. It would of course be feasible to have a logging package including a winch (as in the proposed Nadia system) and place that in the cone with the thruster. Electrical connection would be maintained through a soft tether deployed between the thruster and the logging package, much as we do now, using a 200 m tether, with our seafloor gravity meter.

Once the condition of the hole and its exact location in the transponder net have been determined, a simpler device can be used as a re-entry guide, to be left in the hole indefinitely. This would need only the navigation receiving hydrophone for precise location (better than 1 m uncertainty) during hole entry, and that function might be simply incorporated in the lead instrument package (seismometer, etc). In this context the thruster would provide the load handling and positioning functions, releasing the entire system eventually, in the same manner as in OBS placement.

Some desired manipulations (e.g., making electrical connections) might best be carried out with a work vehicle that can be placed in the seafloor and thus carry out fine work completely isolated from sea surface induced motions. V. Anderson's RUM III (Remote Underwater Manipulator) would fill this requirement very well. This vehicle (Fig. 2) is being adapted for operation on 6,000 m of 0.68" electromechanical cable, with the goal of being used from ships such as MELVILLE. Its power and telemetry requirements and capabilities are similar to those for the Thruster, except that during manipulative operations it utilizes nearly all of the available bandwidth for its two slow scan TV channels. This vehicle emphasizes very localized maneuverability, relying on being lifted off bottom to carry out any significant translation from one work site to another.

In the timeframe under consideration, the expected advent of optical fibers in the context of the 0.68" electromechanical cables will expand telemetry capabilities substantially. Given the tasks involved this will have more impact on RUM operation than on the hole reconnaissance and entry guide functions.

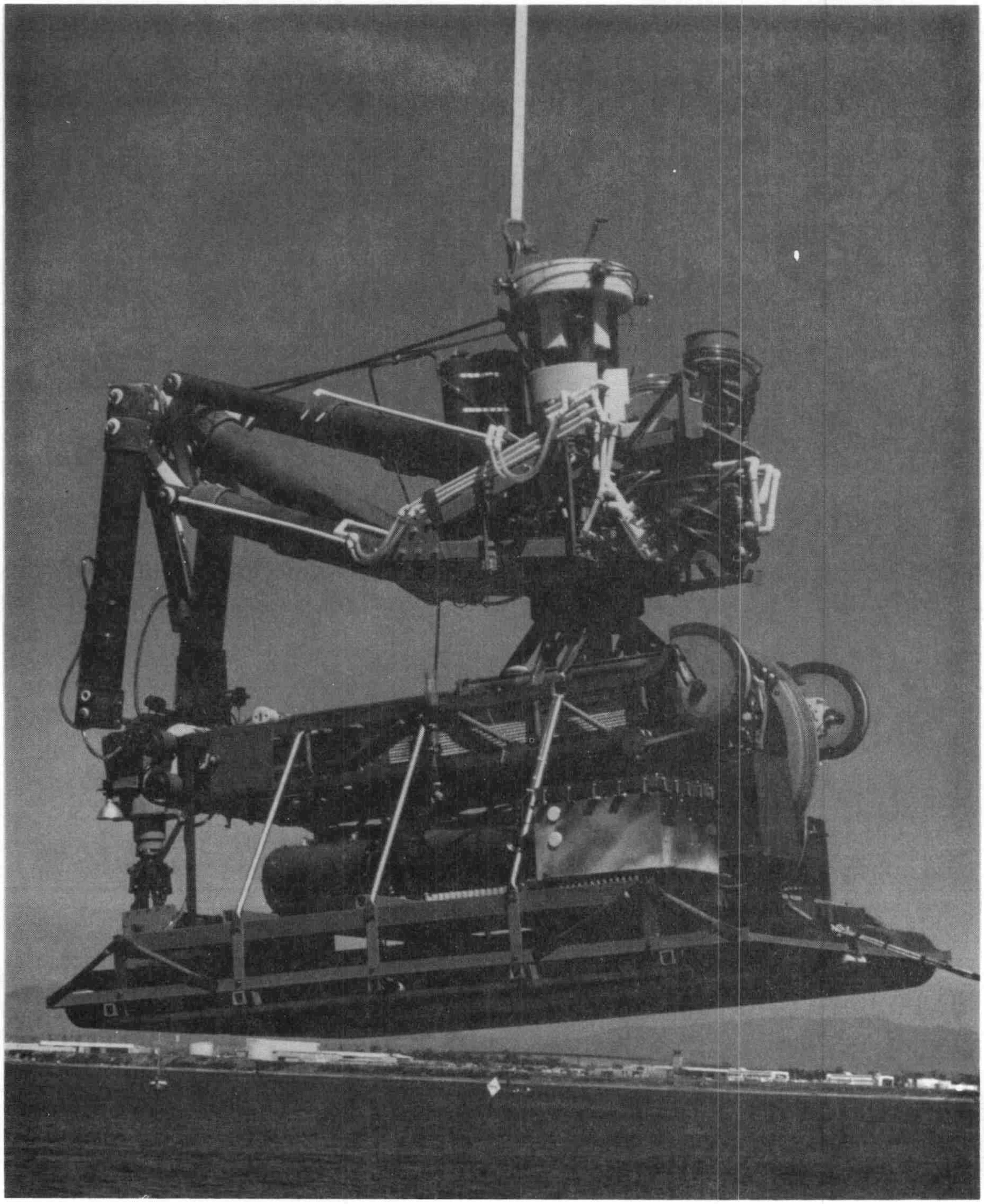

Fig. 2 V. Anderson's RVM III (Remote Underwater Manipulator)

Fig. 1 The Scripps Institution of Oceanography Thruster unit operable on a conventional 0.68" coaxial cable.

PILOT BOREHOLE SEISMIC EXPERIMENTS

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Recent comparisons of short period and ocean bottom seismometer (OBS) observations (Leg 83/41b and Leg 91/DARPA) suggest that considerable signal to noise improvement may be achieved by emplacing the sensors beneath the ocean floor to isolating them from waterborne noise sources propagating along the water/bottom interface. Unfortunately, virtually no long period observation data is available to estimate the S/N performance of broadband borehole or ocean-bottom seismometers. Accordingly, to facilitate the overall objective of comparing the performance level of island observatories, ocean bottom (OBS) and borehole seismometers, the primary objective of the pilot phase BSE program is to determine unequivocally the broadband (+100 seconds to 50 Hz) S/N ratio of borehole seismometers as a function of observation depth beneath the ocean floor. Observations should be made at several tectonically-contrasting sites (ridges, trenches, oceanic plateaus, etc.) in the deep ocean.

Measurements should be made in the unconsolidated oozes and lithifical sediments as well as the volcanic and intrusive rock typical of the upper layers of the oceanic crust. Geophysical site survey information should be available to characterize the velocity structure beneath the borehole site.

Although the borehole location should be selected to allow direct comparison with near island seismic observatories and OBS's, the site selection criteria should also consider the pilot scientific objectives for locating ocean floor sites on normal and anomalous deep ocean crust/upper mantle near ridge crest/trench earthquake source area and to extend the global coverage for long period observation in regions where there are no islands. Ideally, simultaneous seismic measurement will be made in the borehole and at the island and OBS sites along the environmental observations of barometric pressure, wind, velocity, bottom current velocity and pressure, surface wave height and tides. This will provide correlation of the environmental noise with the seismic observation to allow prediction of likely S/N ratio performance for siting future ocean floor seismic observatories.

In order to minimize the drilling cost for emplacing borehole seismometers, special care should be taken to acquire closely spaced observation depths down each borehole, especially in the sediment above the hard oceanic basement rock to determine the minimum depth required to significantly increase the broadband S/N ratio for various sub-bottom structures and environmental conditions. In fact, recent studies suggest it may be possible to achieve marked S/N improvement for OBS type instrument by simply implanting the sensor a few meters beneath seawater/bottom interface.

Three modes of seismometer deployment are envisioned to attain our proposed pilot program objectives:

1) Thru pipe lowerings using a 'slim tool' version (3.5" diameter) of conventional broadband borehole seismometers such as the Teledyne-Geotech model 36000. Attached to the well logging cable aboard the JOIDES Resolution. These experiments would probably be of short duration with recording limited to the period while the drill ship is overhead. However, use of 'pipe stripping' technique, the logging cable can be attached to a tethered seafloor recording package which can be left on the seafloor for longer duration

observations (months) as was done in the HIG short period borehole experiments off Kamchata aboard the GLOMAR Challenger (DSDP Leg 83). A further justification for a 'slim tool' seismometer will be for emplacement in the small diameter mining type borehole planned for deep penetration into igneous rock especially in ridge crests.

2) Strap-on lowerings on the outside of the ODP drill pipe as was done in the previous DARPA borehole seismometer experiment aboard the GLOMAR Challenger (DSDP Leg 91). These experiments could use existing large diameter, broadband borehole seismometers. Also, it may be more easily accomplished aboard the ODP drillship JOIDES RESOLUTION since a television camera shuttle system with co-axial cable is now routinely lowered to locate and insert the drill pipe into the seafloor re-entry cone. This system could be adapted to insert the larger diameter borehole seismometers (~10"). Then development of a smaller diameter broad-band seismometer would not be required.

Strap-on type deployments can also utilize both retrievable mechanical arm and cement bond type hole locking techniques by inserting the drill pipe string into the re-entry cone and pumping cement down the borehole hole before/after emplacing the seismometer.

3) Wireline Re-entry into existing ODP/DSDP borehole using conventional research ships rather than a drillship's pipe may also be available in the near future. For these experiments existing large diameter broadband seismometers could also be emplaced in present and future rotary drilled ODP/DSDP boreholes for both short and long term recording as in the thru-pipe and strap-on. However, hole locking may be limited to mechanical damping arm technique. This may limit the usefulness of the wireline technique for borehole deployments if convective seawater circulation is a serious source of long period noise.

TRANS-OCEANIC TELECOMMUNICATIONS CABLES: RE-USE FOR OCEAN BOTTOM GEOSCIENCE OBSERVATORIES

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GLOBAL BROADBAND SEISMOLOGICAL NETWORKS: THE OCEAN DILEMMA

One of the most promising areas of international co-operative research in geophysics is broadband seismology. At a recent symposium on this topic (International Symposium on Global Seismology and the POSEIDON Project; University of Tokyo; August, 1988) papers were presented by representatives of the United States, Japan, Russia, the Republic of China, Italy, Germany, Canada, the Netherlands, and Australia. National and international associations of earth scientists committed to the establishment of a global network of broadband instruments include: the USGS; POSEIDON (Pacific Orient Seismic Digital Observation Network) of Japan; IRIS (Incorporated Research Institutions for Seismology) of the U.S.; GEOSCOPE of France; MEDNET (Mediterranean Network) of Rome and Geneva, but comprising many Mediterranean countries; ORFEUS (Observatories and Research Facilities for European Seismology), a working group of the European Geophysical Society; and CANDIS (the Canadian Digital Seismic Network). An umbrella organization for all of these associations is the international FDSN (Federation of Digital Seismographic Networks).

Broadband instruments might best be defined as those capable of recording all of the seismic waves generated by an earthquake or underground explosion. Typical specifications would include a dynamic range of about 140 dB and a frequency band from about 20 Hz to 10^{-5} Hz. Although many broadband stations are planned (e.g., the USGS plans a network of 150 for the U.S., and POSEIDON plans 50 for East Asia and the Western Pacific), only a handful are now in operation.

Major barriers to the realization of an effective global network of broadband seismographs are the world's oceans which cover an estimated 70% of the earth's surface. For comprehensive studies of source mechanisms and earth structure, broadband instruments will have to be widely distributed throughout the world's oceans. Such a requirement poses formidable technological and financial difficulties. Systems under consideration by the POSEIDON group include: (a) modified pop-up ocean bottom seismometer (OBS) systems; (2) OBS systems connected to a buoy with satellite transmission of the data; and (3) OBS systems connected to trans-oceanic, fiber optic cables. At present, prohibitive costs as well as substantial technical difficulties are associated with all of these systems. Islands, of course, could be instrumented with little difficulty, but their distribution throughout the world's oceans is insufficient to meet the requirements of an effective global network.

RE-USE OF RETIRED CABLES AS GEOSCIENCE STATIONS

Late in 1988, AT&T and KDD (Japan's international telecommunications company) will complete the laying of a new fiber optic cable from Japan to Guam to the U.S. by way of Hawaii (Kobayashi, 1987). The section connecting Japan, Guam, and Hawaii is called TPC-3 (Trans-Pacific Cable No. 3), and the section connecting Hawaii to the west coast is called HAW-4 (Hawaiian Cable No. 4). Since the present cable (TPC-1) connecting Japan and the U.S. has only 138 channels and TPC-3 has 7560 channels, TPC-1 is to be retired from commercial use. Many other new fiber optic trans-oceanic cables are either under construction or are planned. These include: GPT between Guam, the Philippines, and Taiwan to be completed in 1989; HJK between Hong Kong, Japan, and Korea (1990); TASMAN-2 between Australia and New Zealand (1991); Pac Rim East between Hawaii and New Zealand (1993); and Pac Rim West between Guam and Australia (1996). A cable connecting Japan directly to the U.S. mainland (TPC-4) is also being planned. In the Atlantic, TAT-8 (Trans-Atlantic Cable No. 8) should be completed this year, and the completion of TAT-9 is planned for 1991.

As more of the new trans-oceanic telecommunications cables are put into service, many cables now in use will be retired (Fig. 1). In the near future, the retirements of TPC-1, GP-1, HAW-1, and COMPAC (Fig. 2) could provide a unique opportunity for the establishment of a trans-Pacific geoscience cable system. Instruments could also be connected to retired cables in the Atlantic. In addition to broadband seismometers, other instruments could include meters to measure gravity, tsunamis, earth tides, deep ocean currents, and electromagnetic fields.

FEASIBILITY TESTS USING THE JAPAN-GUAM SEGMENT OF TPC-1

For a number of years, Japanese seismologists have been discussing the use of oceanic telecommunications cables with officials of KDD. Recent discussions also have been undertaken by U.S. and Japanese seismologists with officials of AT&T. As a result of these efforts, the re-use of the Japan-Guam segment of TPC-1 has been positively discussed by KDD and AT&T. Thus, the first step in fulfilling the dream of a trans-Pacific geoscience cable system may soon become a reality. Instrumenting the Japan-Guam segment will provide a unique opportunity to test the feasibility of such a system. In the discussions that follow, some special problems associated with re-use are presented.

Residual Life

Although the idea of re-use is easily understood and widely accepted, some serious questions concern the estimated residual life of the retired cables. Major factors affecting the life of the system are failures of electronic tubes used in the repeaters and failures of the cable line due to natural as well as artificial disasters. These types of failures have been investigated with KDD and NOEL (NEC Ocean Engineering, Ltd. of Japan's Nippon Electric Company) by use of papers and reports of the original design and manufacture of the cables, as well as data and records of actual performance of the system for more than 20 years.

Two types of tube failures could be expected: the breakdown of various tube components and the eventual "wear-out" of thermionic emissions. Regarding "breakdown", the tubes were manufactured so as to satisfy the requirement of a mean time between failures of 885 years (Holdaway, et al., 1964). The estimated number of failures along the Japan-Guam segment in a thirty-five year period due to "breakdown" is 1.0. Design and manufacturing data for thermionic emissions indicate that the theoretical life is more than 50 years (Kern, H., 1960).

Adjustments of Power and Data Transmission

The installation of sensors into the present cable will require an adjustment of the power feed. The current throughout the cable is a constant 370 mA, and all repeaters are connected in series. The supply voltage will have to be increased according to the demands of the added sensors so as to maintain the required 370 mA current. The voltage drop across each repeater is 45 V, and the present supply voltage to the Japan-Guam cable is 4500 V.

To meet the wide dynamic range requirements of broadband seismic data, the existing cable transmission format will have to be modified from analog to digital. More specifically, the existing FDM-AM (Frequency Division Multiplex - Amplitude Modulation) system will have to be changed to an FDM-PCM (Pulse Code Modulation) system. These analog to digital conversions will take place at the sensor unit. The transmission capacity of one voice channel will be 7200 Baud, which is appropriate for one sensor unit containing three-component seismometers with sampling rates of 100 Hz.

Sensor Units

Two types of sensor packages are planned: Unit A and Unit B. Unit A will be a three-component seismometer with a shape similar to that of the repeaters used in the existing system. The unit will be installed in the same way that cable defects are repaired, and will be installed mid-way between repeaters so as to minimize the possibility of disturbing the repeaters (Fig. 3). Given the present state of OBS development, the requirements of Unit A for small size and ruggedness prevent it from fulfilling all of the requirements of a broadband system. The instruments now being considered for Unit A are small-size, high-sensitivity accelerometers with flat displacement responses up to 30 seconds. The accelerometers resolving power of about 10^{-7} g corresponds to 1 μm of displacement at a period of about 6.3 seconds. Such sensitivity should be adequate for body wave tomography. In addition, these instruments will, of course, be capable of recording all types of body phases and many surface waves.

Unit B will serve a variety of purposes (Fig. 4). It could contain a gravimeter, magnetometer, electro-potentiometer, pressure sensor, and current meter. It also could be the desired broadband seismometer, or possibly, and ODP (Ocean Drilling Program) downhole seismometer. Because of the large size and weight of this sensor, it could not be installed in the same manner as Unit A. Instead, the deployment of Unit B must utilize a "branching technique" which will require the unit to be attached to one end of a special branching cable which would then be connected to the main cable by a special device called a "branching unit". Since Unit B could contain several sensor elements (some exposed to seawater), special safety devices will be installed to prevent malfunctions of any sensor component from affecting the performance of other components, or the whole system.

FINAL REMARKS

The anticipated retirement of trans-oceanic telecommunications cables in the Atlantic and Pacific will provide unique opportunities for the community of earth scientists to establish much-needed deep-ocean science observatories. An especially important re-use of these cable would be the establishment of deep-ocean broadband seismographs. Such instruments are especially useful for regional and worldwide studies of earth structure using seismic tomography. Some of the topics which can be investigated include mantle flow patterns, mantle slab penetrations, and the topography of the mantle-core interface. Broadband instruments are also especially useful for comprehensive studies of earthquake source mechanisms and seismic risk assessment.

Furthermore, the possible re-use of telecommunications cables has added significance at this time because of: (1) the small number of broadband instruments now in existence; (2) the many national and international associations of earth scientists dedicated to the establishment of a worldwide network of broadband seismographs; (3) the special difficulties posed by the world's oceans to the establishment of a useful global network, and (4) the recent discovery of significant unreported earthquakes in the interior of the Northwestern Pacific Basin and the implication that ocean basins may be more seismically active than is generally believed (Walker and McCreery, 1988).

Most importantly, we hope to rigorously test the feasibility of utilizing trans-oceanic telecommunications cables beginning in 1989 with the establishment of geophysical observatories along the Japan to Guam segment of TPC-1.

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Kobayashi, K., The first optical fiber submarine cables system in the Pacific Ocean, *AEU, J. Asia Electronics Un., 4,* 78, 1987.

Walker, D., and C. McCreery, Deep-ocean seismology: Seismicity of the Northwestern Pacific Basin interior, *EOS Trans. AGU, 69*, 737,1988.

FIGURE CAPTIONS

Fig 1. The map shows the existing worldwide network of coaxial trans-oceanic telecommunications cables which are being complemented by a new generation of fiber optic cables. Because of the greater channel capacity of fiber optic cable, most of the existing coaxial cables will be retired from commercial use, thereby providing the community of earth scientists with a unique opportunity for deploying a variety of geophysical instruments throughout the world's oceans. Actual testing of the re-use of trans-oceanic cables for geophysical observatories could begin as early as 1989. This diagram of cable routes is after "Global Network System" (plate, Kokusai Denshin Denwa, Ltd./1987 (Ann. Rept.), Kokusai Denshin Denwa, Ltd., Tokyo, p. 22-23 and after "Coaxial Submarine Cables and Cable Ships in the Pacific Area", (plate), The Cable Ship KDD Maru, Kokusai Cable Ship Co., Ltd., Tokyo (inside back cover).

Figure 2. Map showing the location of coaxial trans-Pacific telecommunications cables which may be retired from commercial use in the near future (after Kobayashi, 1987).

Figure 3. Schematic diagram showing sensor unit A (a small three component seismometer) installed between two repeaters.

Figure 4. (a) Schematic diagram showing sensor unit B (broadband seismometer, gravimeter, magnetometer, electro-potentiometer, pressure sensor, and/or current meter) connected to the end of a special branching cable which is connected to the main cable with a special branching unit located midway between repeaters. (b) Unit B could also be an Ocean Drilling Program downhole seismometer.

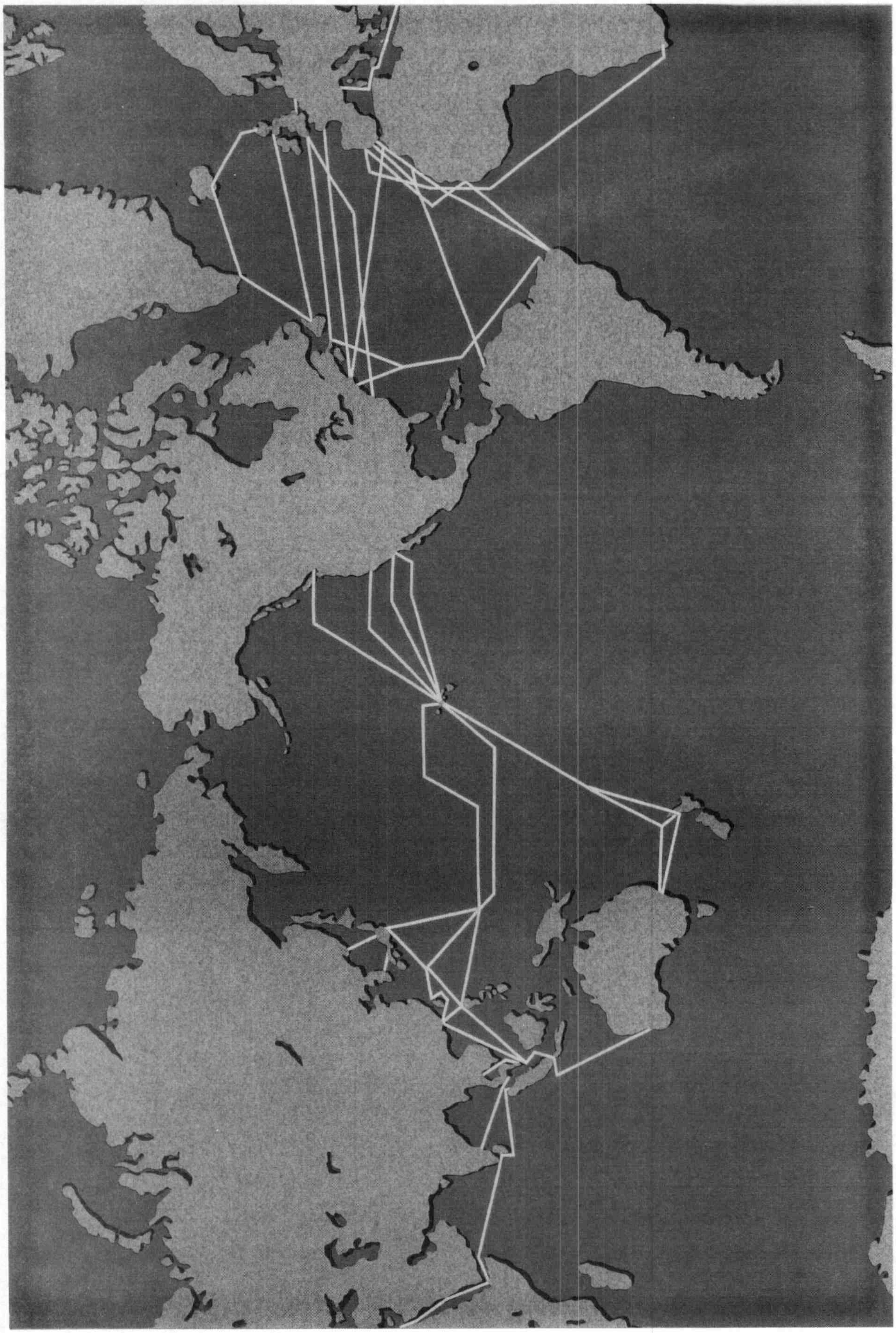

Figure 1

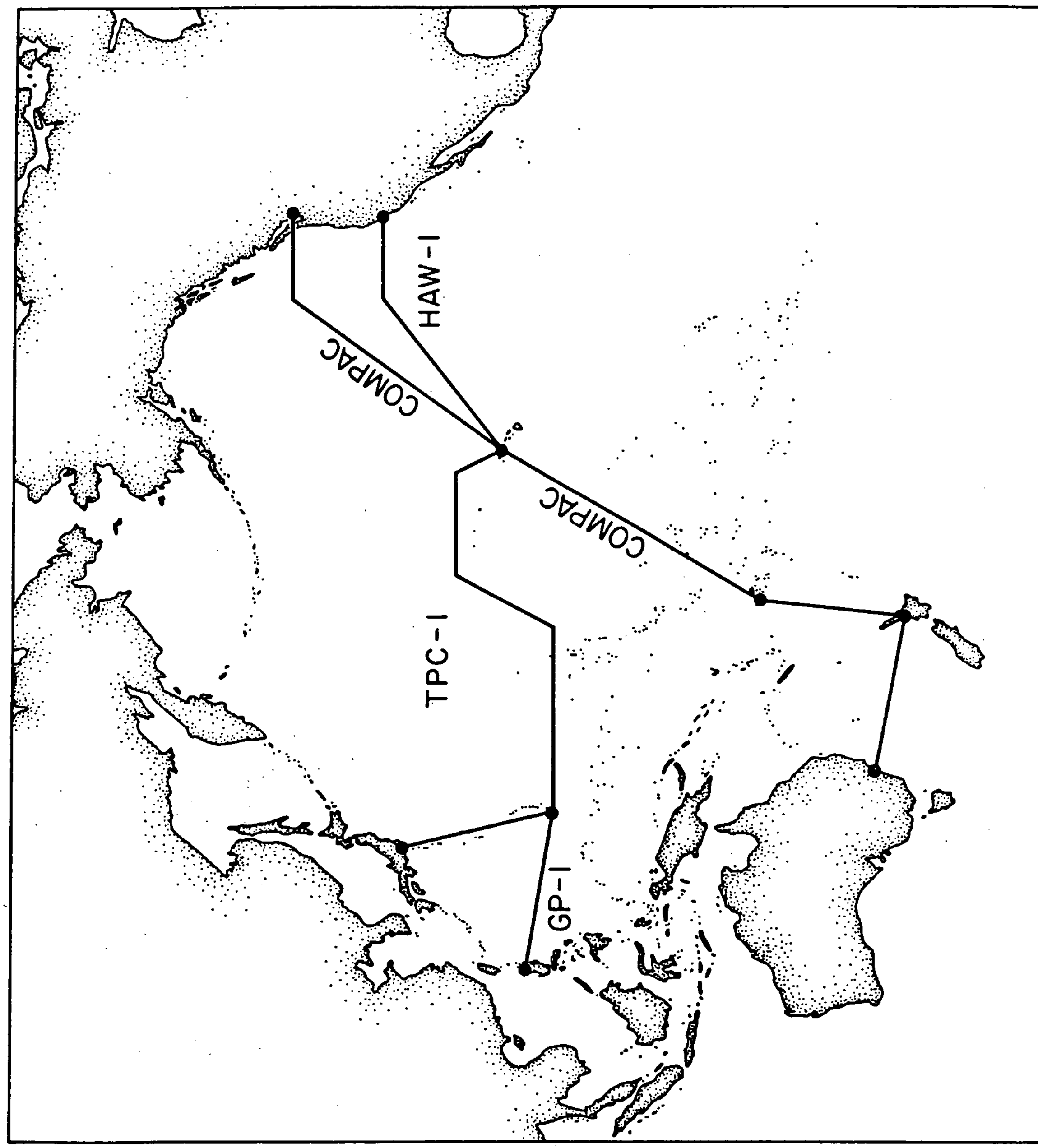

Figure 2

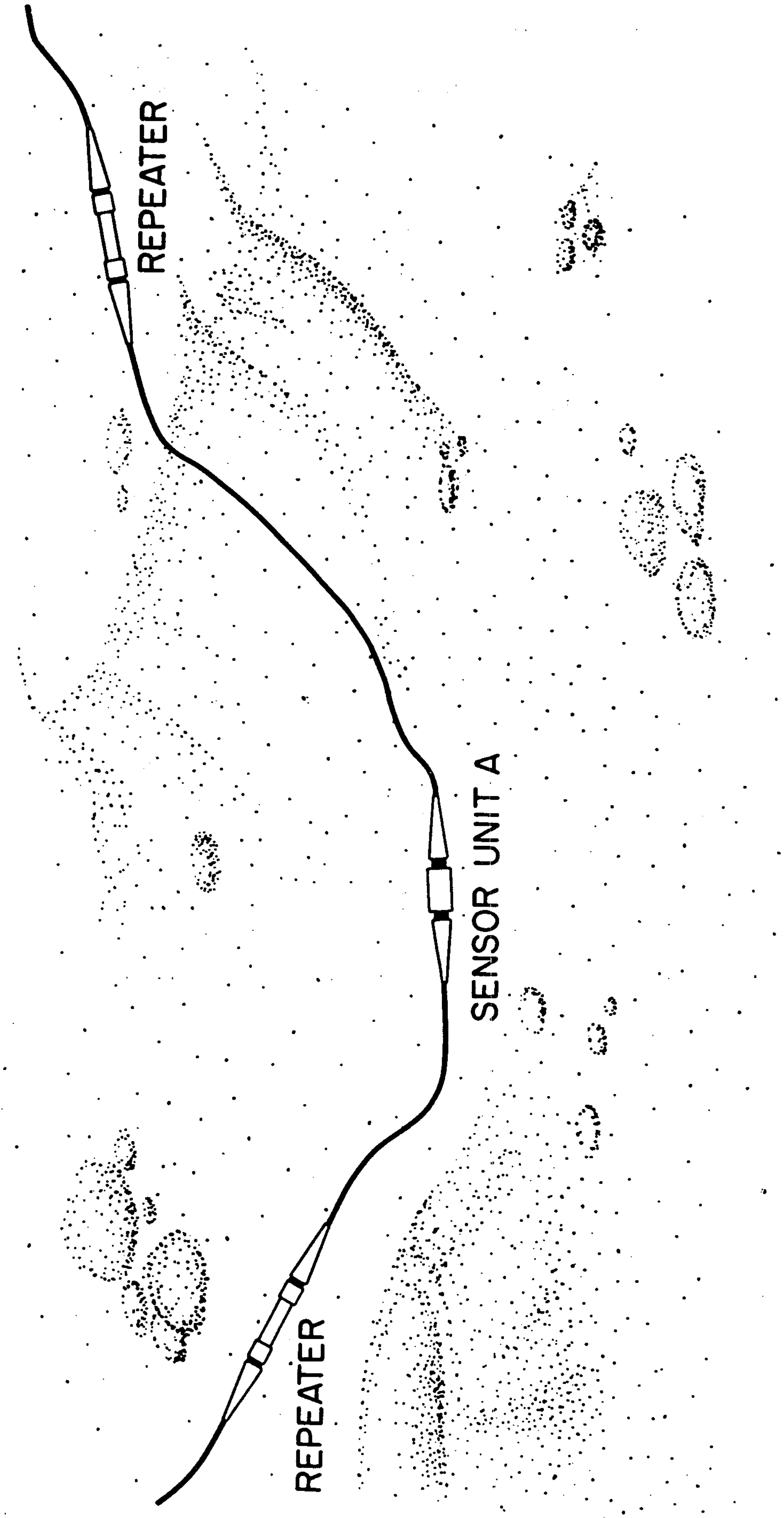

Figure 3

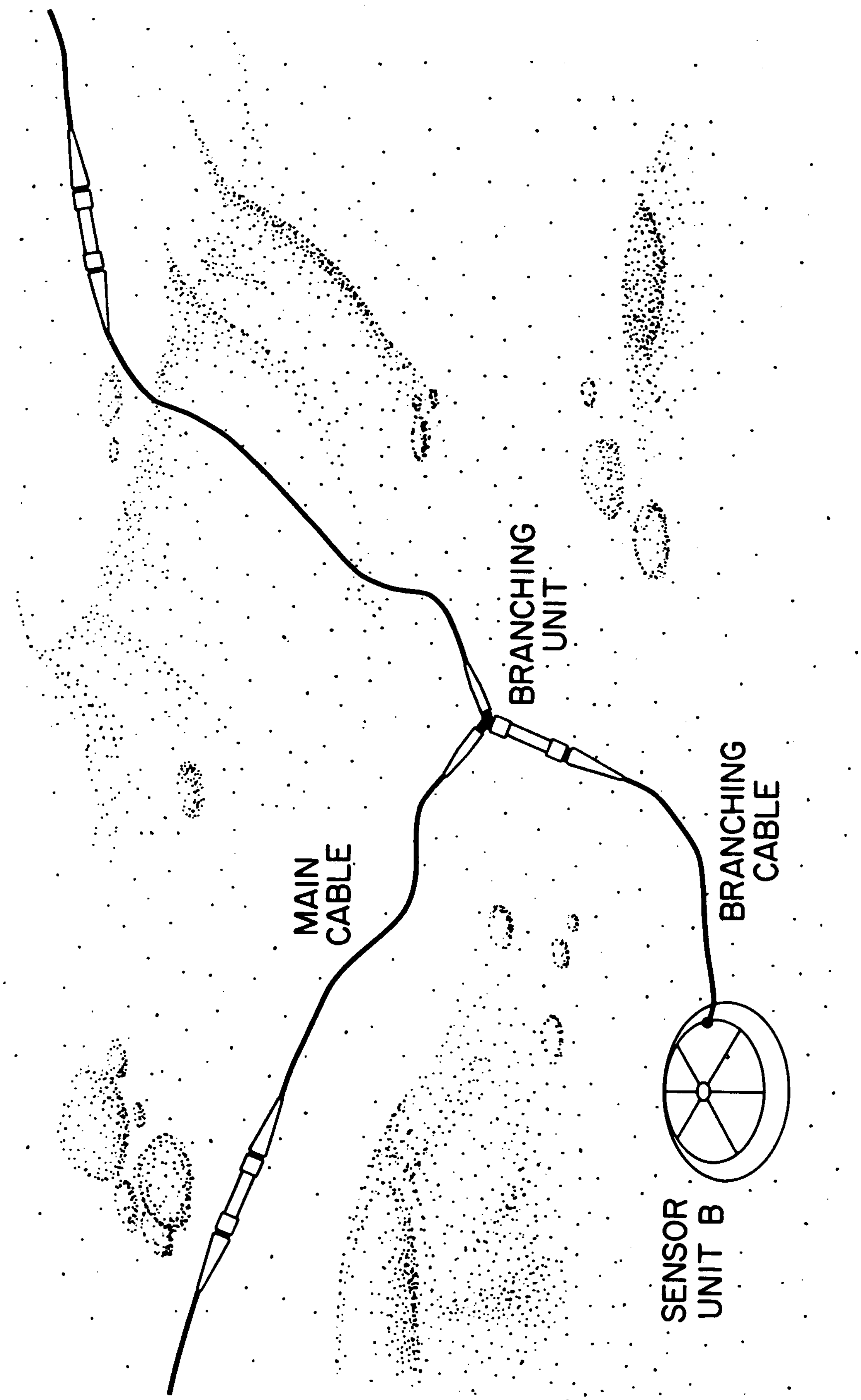

Figure 4a

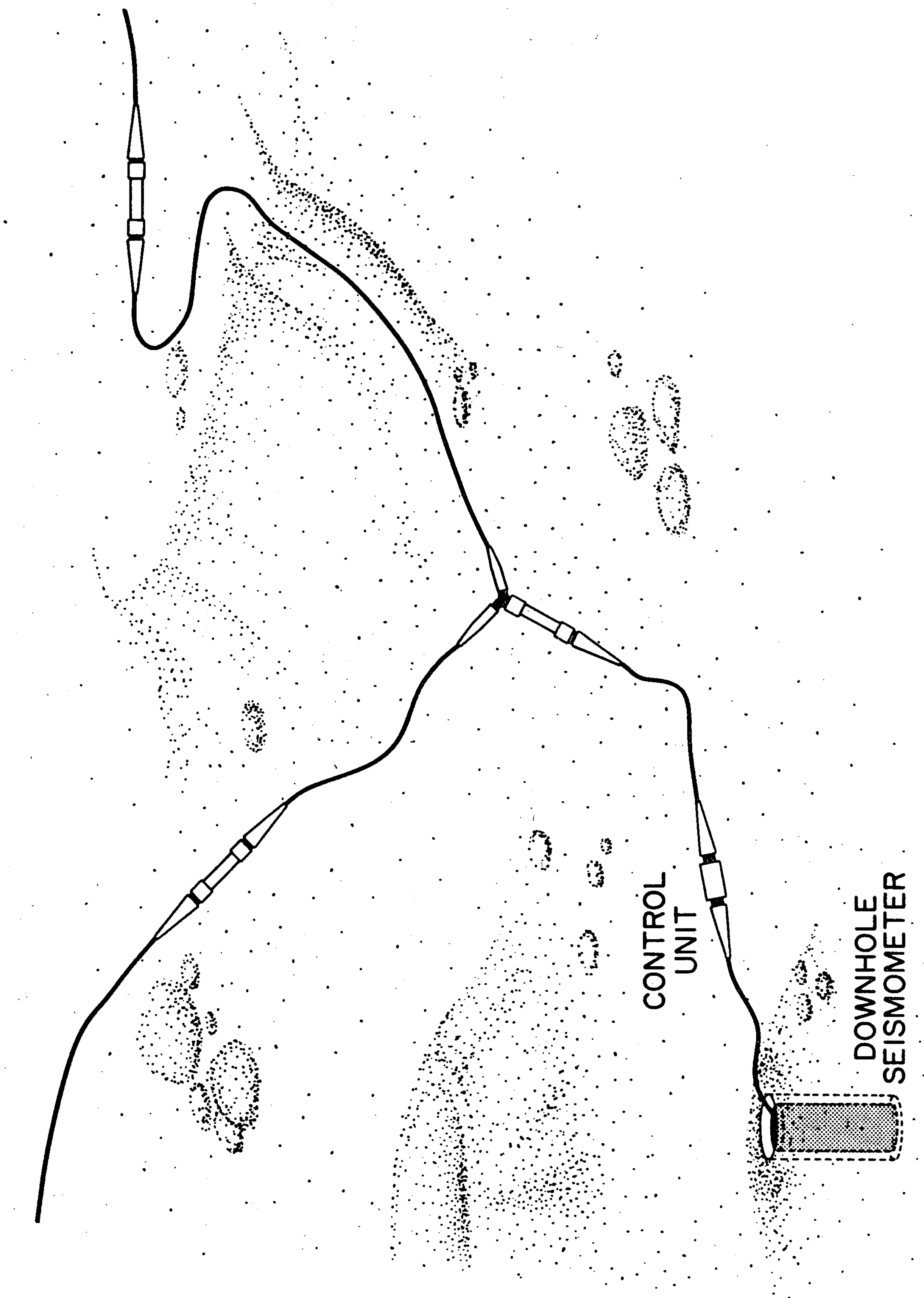

Figure 4b

SATELLITE TELEMETRY IN OCEANOGRAPHY

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Telemetry of data from oceanographic sensors is not a new idea, but one which has recently become more feasible with the advent of low power satellite communication techniques which have come into use in the last decade. While none of the existing satellite telemetry links can provide the high data rates and low power/bit efficiencies required to telemeter unprocessed, rapidly sampled seismic data from buoys to shore, there are several systems under development which may provide partial solutions to this problem.

The satellite data links which are available for use by oceanographers include both geostationary satellites and polar orbiting satellites. The difference between the two is that the geostationary satellites, which maintain their position relative to a point of longitude, usually over the equator, provide continuous coverage of a large part of the hemisphere beneath them, exclusive of the high polar regions. The polar orbiting satellites, which fly lower and require less power to reach, provide occasional coverage of the entire globe. Geostationary satellites systems include:

ATS

ATS-3 was launched in 1967 and made available to the oceanographic community for experimental communications from research ships in the early 1970's. It is presently operated by the University of Miami under joint funding from NSF and NASA. It operates on 149 MHz (uplink) and 136 MHz (downlink) and requires a stabilized antenna trained to point at the satellite. Typical use of the ATS is for voice communications from oceanographic research vessels or remote experimental stations to their home laboratories plus some data transmissions. Use of ATS for offshore buoy applications is impractical due to the age of the satellite (it was designed for an 18-month life and has now been flying for more than 20-years) and the type of antenna normally used.

INMARSAT

The International Maritime Satellite Organization (INMARSAT) employs an array of geostationary communications satellites which provide global (except the poles) coverage. It is a commercial system which was designed primarily for the commercial shipping industry, providing voice communications as well as several classes of data communication service. It can support data telemetry at speeds up to 56 kbaud. Its two principal disadvantages for oceanographic use are the cost and power required by the shipboard transceiver and the requirement for large gyro-stabilized antennas. INMARSAT charges a fee for its services based on the type of service and the time of use.

A new service to be offered by INMARSAT, which has some promise for oceanographic use, is now being developed and tested. This system is referred to as Standard-C. It uses the INMARSAT satellite constellation to provide low data rate (600 baud), two-way data communication. It is aimed at small boat owners and is designed to operate with low cost transceivers and small, omni-directional antennas of a type that would be suitable in buoy applications. There is presently a written system specification and several manufacturers are testing prototype radio equipment for use with this system. Costs for use of the system have not yet been defined, but may be significant for a user who wants to make full use of the available data throughput. The equipment presently under development is meant for shipboard use and as a result is not particularly low power, but it may be feasible to modify these designs to better meet the power requirements of buoy operation.

GOES

The Geostationary Operational Environmental Satellite (GOES) system is a constellation of three satellites which provide continuous coverage of the non-polar regions of the Western Hemisphere. The GOES Data Collection System is operated by NOAA expressly for the purpose of collecting environmental data. This system is used most heavily for hydrological monitoring in the US and Canada, but a number of oceanographic buoys have carried GOES transmitters, most notably the NDBC surface buoys used to monitor meteorological conditions around the US coastline. Analogous geostationary satellites are maintained by Japan (GMS) and the European Space Agency (METEOSAT) which cover most areas of the Eastern Hemisphere and allow the use of identical equipment at the remote sites, thus effectively providing worldwide non-polar coverage. Use of this system is free of charge. Its major limitations for oceanographic use are low data throughput and a relatively high power/bit ratio. The GOES system does have the potential for limited two way communication, although equipment to implement this feature is not commercially available, at present.

GEOSTAR

Geostar is a new commercial venture which uses geostationary satellites (only one is operational at present) to provide two-way data telemetry and position determination from small remote transmitters. Its goal is to offer low cost service to large scale users, such as the commercial trucking industry, the National Weather Service, and others who need data and position from large numbers of remote locations or platforms. Its primary coverage area is determined by the overlapping footprints of multiple satellites and since the principal markets for its services are land-based, so are the footprints. Geostar plans to provide coverage for the US, Gulf of Mexico, much of Central and South America, and eventually Europe, Australia, China and India (and the Indian Ocean). The exact timetable for this expansion and the ultimate coverage of the oceans is unclear at this time. A pilot oceanographic program, however, is under discussion between Geostar and WHOI researchers to use the system on a surface mooring to be deployed off Bermuda late this year. The advantages of Geostar for oceanographic use are relative high data rate, small size, low power, and relatively low cost. The disadvantage is obviously, the lack of coverage over the open ocean.

ARGOS

The polar orbiting satellite system which is available for data telemetry from ocean platforms is the Argos system. The hardware for this system flies on two satellites which orbit the globe every 105 minutes and provide discontinuous global coverage. The system is operated by Service Argos, a French organization with a North American component located outside Washington, D.C. This system allows the user to telemeter small quantities of data from anywhere on the globe for a reasonable power and cost per bit. It also provides position information for each remote unit on each pass of the satellite. Its major limitations for oceanographic use are the low data throughput available and the lag time between data collection and user reception, which may be of the order of several hours or more in some situations.

Other telemetry techniques which have application in oceanography include VHF radio for line-of-sight applications, HF radio for over the horizon applications, meteorburst radio communications which offer limited over the horizon capability with more reliability than HF, and acoustic telemetry using the SOFAR channel, which can be used for low data rate telemetry over basin wide distances. In addition there are various schemes for hybrid satellite techniques which utilize a combination of polar orbiting and geostationary satellites. None of these other techniques, however, offer a general purpose, reliable method for telemetering data from ocean buoys, but they do have promise in special applications where existing satellite links cannot meet the needs for data throughput, cost, or security. Typical specifications for some of these non-satellite telemetry methods have been included in the following tables for comparison purposes.

Additional information on these and related topics is available from the following sources:

1) Briscoe, M.G. and D.E. Frye, Motivations and Methods for Ocean Data Telemetry, MTS Journal, Vol. 21. No. 2, 42-57.

2) MacCallum, D.H. and M.J. Nestlebush, The Geostationary Operational Satellite Data Collection System, NOAA Technical Memorandum NESDIS 2, June 1983.

3) INMARSAT Standard-C System Definition Manual, INMARSAT Technical and Operations Division- Draft Issue No: 2.3, August 1986.

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5) Richards, R.T., GEOSTAR Radio Determination Satellite System - Principle of Operation and Positioning Accuracy.

Telemetry Options for Ocean Buoys

OPTION	SATELLITE	COVERAGE	LOCATION	HARDWARE	NOTES
ATS	Geostnry	W. Hem.	N/A	$6,000	a, d
Standard-C	Geostnry	Worldwide (ex. Poles)	N/A	$5,000	b, c, e
GOES/MET-EOSAT/GMS	Geostnry	Worldwide (ex. Poles)	N/A	$3,000	
Geostar	Geostnry	US Cont.	+ 10m	$3,000	b, e
ARGOS	Polar	Worldwide	+1Km	$1,000	
VHF Radio	N/A	10-50 Km	N/A	$1,000	e, f
HF Radio	N/A	Variable	N/A	$2,000	c, e, f
Meteorburst	N/A	100-1200 Km	N/A	$5,000	e, f

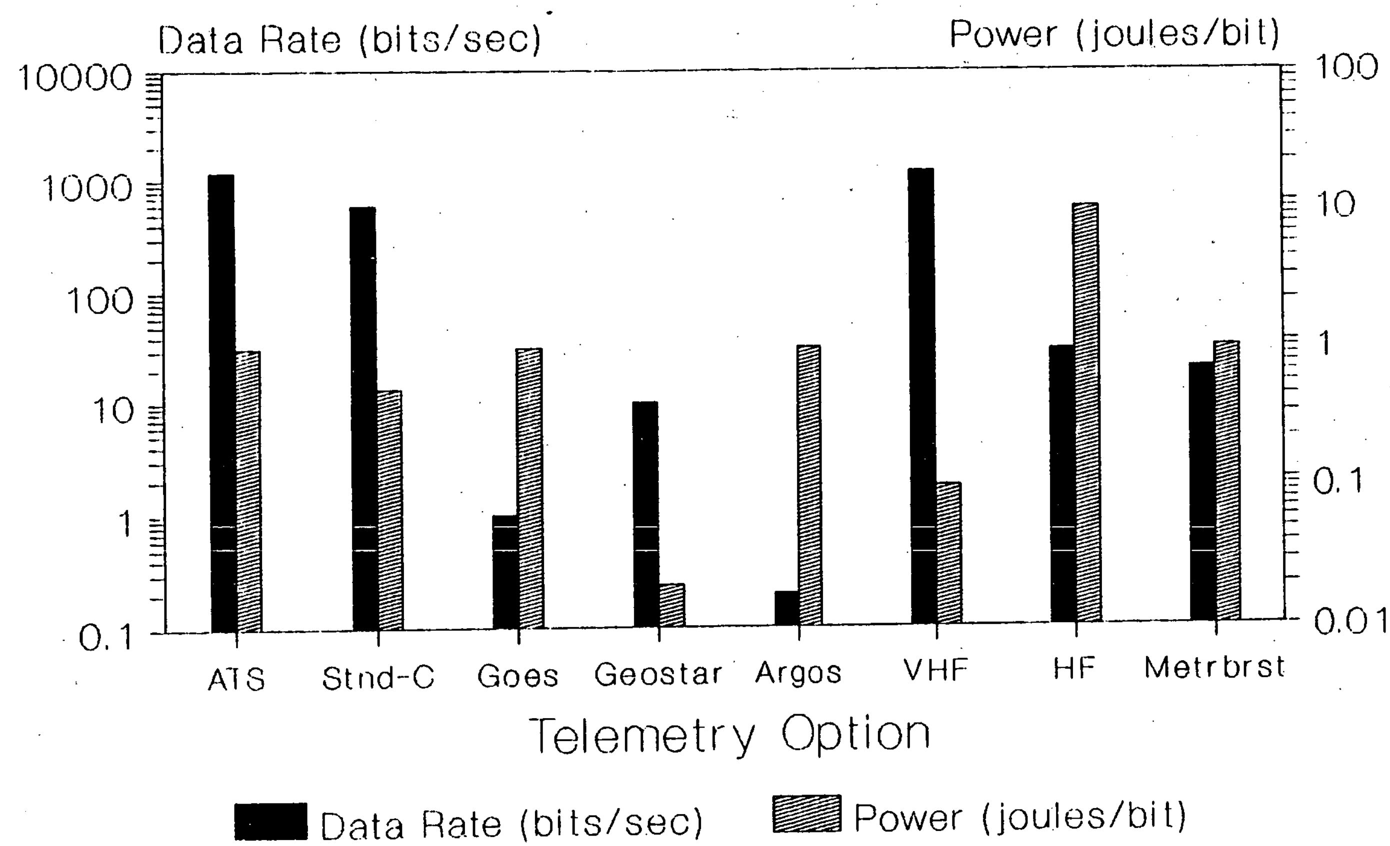
NOTES:
a) may not be available in the future
b) not yet operational
c) hardware not optimized for low power applications
d) requires directed antenna
e) two-way telemetry system
f) receive station required

Telemetry Options for Ocean Buoys

OPTIONS	DATA RATE	POWER	COST/KBITS	NOTES
ATS	300 - 1200 bps	.5 - 1 j/b	0	d
Standard-C	150 - 600 bps	.5 j/b	$12.50	a, b, d
GOES/MET-EOSAT/GMS	.4 - 1.2 bps	1 j/b	0	
Geostar	.2 - 10 bps	.02 j/b	$.08	b
ARGOS	.02 - .2 bps	.5 - 5 j/b	$.50 - 5.00	
VHF Radio	300 - >1200 bps	.02 - .1 j/b	0	c
HF Radio	1 - 30 bps	1 - 10 j/b	0	c
Meteorburst	1 - 20 bps	.2 - j/b	0	c

NOTES:
a) tariff not yet determined
b) estimated values
c) requires receive station
d) data rate values are typical for continuous telemetry applications

Telemetry Options
Data Rate and Power



WIRELINE RE-ENTRY OF BOREHOLES IN THE DEEP OCEAN

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Wireline re-entry is in the process of becoming a state-of-the-art technique. At the present time, at least three initiatives are underway. One constructed by the French Oceanographic Institution IFREMER is known as the project NADIA; the other conducted by a group of 4 American institutions (WHOI, SIO, NORDA and Johns Hopkins University) is named LFASE; and the USSAC Wireline Re-entry Program. The French and USSAC projects were described in the proceedings of the workshop: "Science Opportunities Created by Wireline Re-entry of Deep Sea Boreholes" held at Scripps Institution of Oceanography in February 1987.

NADIA

The concept is based on the use of a manned submersible as the vector for rendezvous with the re-entry cone and control and power supply to the logging module NADIA. A re-entry operation has been done this summer 1988 in DSDP Hole 396B. A temperature probe (self-recording, 200mm in diameter) and a water sampler have been lowered into the hole to the depth of 300m. Despite a re-entry cone found buried into the sediment by approximately 1m the re-entry operations have been conducted very smoothly.

Following plans are to transform NADIA into a real logging unit, i.e., installation of a new winch holding 1500m of 7-conductors cable, a data recording unit and a power unit. Tool length will be 4.5m maximum.

LFASE (Low Frequency Acoustic Seismic Experiment)

The concept in this case is based on the use of the dynamic positioning capability of the surface ship, and a motorized frame suspended from the ship by an electromechanical coax cable, to hold position for a certain time. The objective of the project is to place a string of borehole seismometers in Hole 418A. Data will be recorded at the ship for a 7-day period, via the EM cable. Then, the seismometer string with a data recording unit will be detached from the main cable and will record autonomously for a 30-day period. Data will be retrieved either after the recovery of the borehole equipment by means of the same system used for installation, or by acoustic release of a data cartridge separated from the data recording unit.

Plans are to test the system at sea in April 1989 and to conduct the re-entry operation during the following summer.

USSAC Wireline Re-Entry Program

The U.S. Science Advisory Committee (USSAC) has recommended that a lead organization be identified within the U.S. to develop a wireline re-entry system, carry out acceptance tests and operate the wireline re-entry facility for the U.S. scientific community. The organization would also be the focal point for U.S. cooperation with international efforts on wireline re-entry.

The RFP for the Wireline Re-entry Project Director should be released by July 1988.

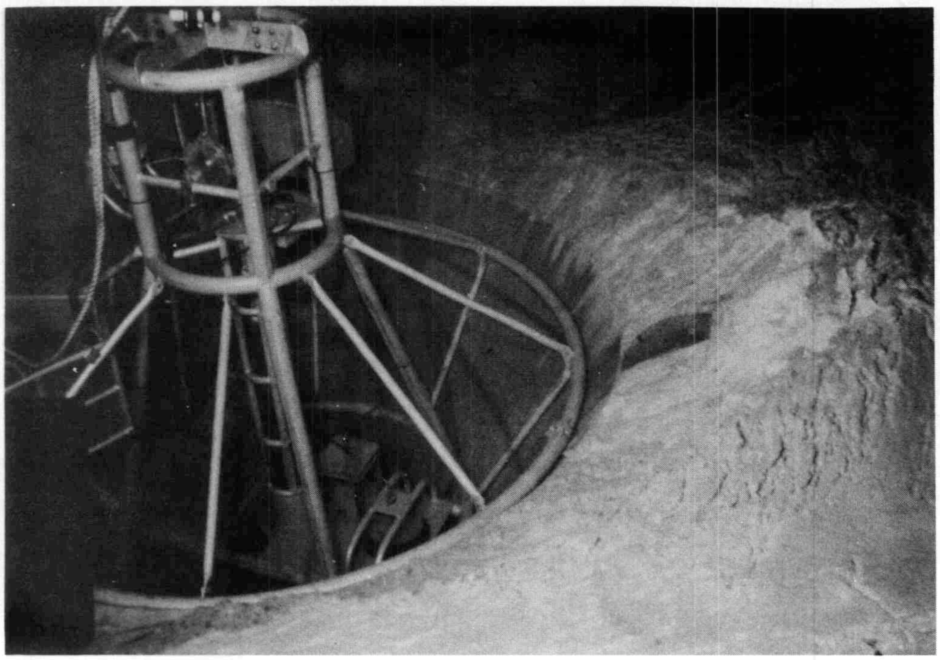

Fig. 1 The logging module NADIA emplaced in the reentry cone of DSDP Hole 396B in the summer of 1988. Note how the reentry cone has sunk into the sediment.

PASSCAL -- A NEW GENERATION OF PORTABLE SEISMIC INSTRUMENTATION

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†Incorporated Research Institutions for Seismology

Summary

A portable 6-channel seismic recording system with state-of-the-art digital technology has been developed by the Incorporated Research Institutions for Seismology (IRIS), funded under cooperative agreement with the National Science Foundation. These instruments are a response to a long established need for hundreds of channels of matched data for studies of the continental lithosphere by extension of the conventional reflection and refraction methods. They were also developed with event triggering and timing capabilities which would make them suitable for a wide variety of earthquake recording applications involving large numbers of channels and the use of temporary sites. Development of these instruments was coordinated by the Program for Array Seismic Studies of the Continental Lithosphere (PASSCAL), founded in 1984 under the IRIS corporate umbrella. Specifications were drawn up by a joint Committee of university and industry participants, and the prototype systems were designed and implemented by Refraction Technology Inc. as a result of competitive bidding. Standardization of data formats, field playback facilities, exchange media, and basic processing software greatly simplifies the use of these instruments by many investigators and the exchange and processing of data. With an anticipated complement of 85 instruments (510 channels) by mid-1989 and 500 instruments (3000 channels) by 1993, IRIS will be able to supply instruments for a wide variety of multi-channel experimental needs in the on-land seismological community.

History of the PASSCAL Instruments

By the end of the 1970's portable seismic instruments for university use in refraction studies were becoming technologically obsolescent, and individual investigators were unable to obtain funding to acquire newer instrumentation through ordinary channels. At the same time, the great power of densely sampled portable arrays using digital technology had been impressively demonstrated by industry engaged in seismic reflection work. At the global scale, the need for a broad-band digital seismic network was being felt. It became evident that a major investment in technology and field programs would be necessary if the U.S. seismology community were to retain a position at the cutting edge of the science. A series of special studies by the National Academy of Sciences/National Research Council, formulated recommendations for new large-scale cooperative national programs to organize and carry through on these new opportunities [1,2,3], with funding to be focused in the NSF. Then ensued in 1983-84 a series of organizational meetings of seismologists which led to the formation of IRIS and the adoption of the PASSCAL program; the technical requirements for the new generation of portable instruments were brought into focus concurrently and published in two workshop reports and in the PASSCAL Program Plan itself [4,5,6]. A more complete organizational history of IRIS and PASSCAL was given by Smith[7].

Initial technical discussions focused strongly on the requirement that the new instruments be built around a public bus architecture to insure that future technological advances could be implemented at the module level and that the program not be captive to a single vendor. Out of this developed an innovative project at UCLA to develop a new type of communications bus which would put the new instrumentation at the cutting edge in system design, by maximizing the modularity in a system which would be modeled after a local area network.

Startup of PASSCAL under the initial funding (1986) involved two efforts. An Instrumentation Committee completed work on the technical requirements for the new instruments, and an evaluation was made of the likely status of the UCLA bus effort in terms of the PASSCAL schedule. An RFP was issued in fall, 1986, setting forth the requirements, which included that of a public bus architecture; although the UCLA bus met the requirements, its migration to "off the shelf" status would have taken another two years, and it could not be further supported. Award of a contract was made to Refraction Technology, Inc. of Dallas, in March, 1987, for construction and delivery of 5 prototype systems, by March 1988. These systems have been delivered (June, 1988) and are undergoing bench and field testing, preliminary to acceptance and the planned ordering of the first production run.

Description of the PASSCAL systems

The systems are composed of four physically distinct units (Figure 1). The 6-channel *Data Acquisition Subsystem* is the heart of the system, containing most of the functionality, shown in the system architecture (Figure 2). Each DAS is equipped with a *Time Keeping Subsystem*, which can be selected from among several options. 12v battery power is packaged in the base of the DAS. A handheld *Field Setup Terminal* is used to monitor DAS status and to download operating instructions.

With 4mbyte of onboard RAM, the DAS can provide full onboard data recording service for most field programs, and an *Auxiliary Recording Subsystem* is provided for the downloading of data from the DAS onto a 2.3gbyte 8mm video cartridge. For such applications, only one ARS is needed for every N (~20-40) DAS units. Alternatively, for remote earthquake recording arrays where external power is available and very large data quantities are involved, one ARS may be installed for each DAS unit.

Specifications: *Table 1* gives the most important technical parameters of the instrument.

Field playback systems:

A prime requirement for any application involving portable seismic recorders is the availability of high performance playback facilities in reasonable proximity to the field area. The availability of the high performance workstation technology makes this a natural path for PASSCAL. Such systems can easily be broken down into two or three small shipping containers and set up at a local site, such as a motel or school. Among the functions performed by these field computers are:

- Quality control checks
- Formation of event catalogs
- Plotting record sections
- Preliminary data assessment and processing

Since these chores often take many months when done back in the research lab, a substantial speedup in total interpretation time is achieved by moving the housekeeping efforts into the field, were they overlap the acquisition activities.

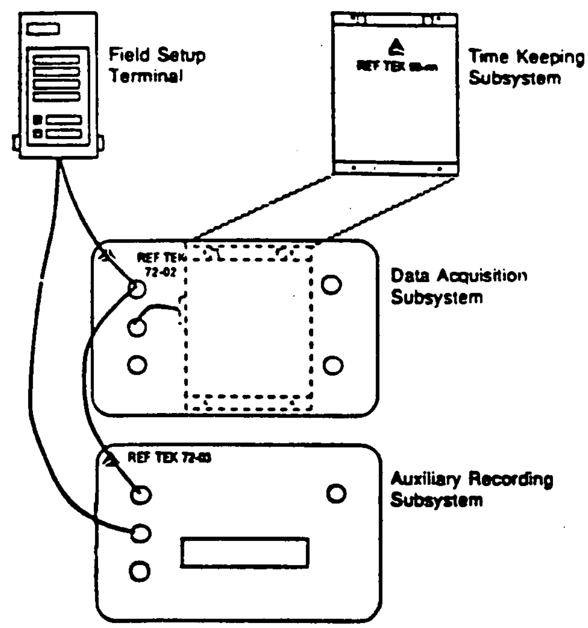

Figure 1: System Components

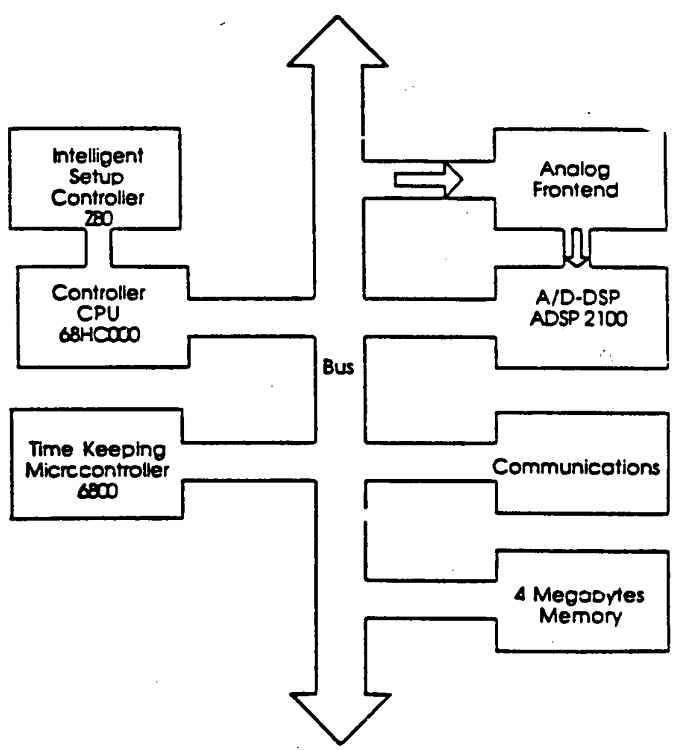

Figure 2: System Architecture

Table 1
PASSCAL Instrument Summary

Physical
- 32#, including batteries
- 12v Lithium batteries, isolated from electronics compartment
- 11" x 18" x 9"
- specified water, dust, shock resistance
- 1.2W operating power, 800 mw on trigger, 5mw sleeping

Architecture (Figure 2)
- Multibus II, with ...
- Nonproprietary hardware and software
- 68HC000 Controller
- intelligent front end A-D module with ADSP 2100 digital signal processor chip
- 6 channels of low level seismic voltage input using AMP01 op amp.
- 16 bit A-D converter
- self-test and calibration
- internal stabilized time keeping microcontroller with connection to external Time Keeping Subsystem
- 4 megabytes static RAM
- RS 232C and SCSI data ports

Triggering
- 16 preset time programs
- event triggering, user selectable, including LTA/STA
- external trigger
- 2048 s/channel pre-event buffer

Timing
- 10ms absolute accuracy, 1ms relative accuracy across array
- External clip-on modules: choice of Omega, WWVB, and GOES time, and UHF radio telemetry
- Time tagging of data packets

Data Flow
- Time and parameter labelled data packets are demultiplexed
- Storage in onboard 4 mbyte static RAM, with separate event catalog
- Download over SCSI port or RS232 port

Bulk recording medium
- 2.3 gbyte 8mm helical scan video tape cartridges (EXOBYTE)
- may be installed in battery compartment, if power is external
- used in Auxiliary Recording Unit, for download from DAS

Any common middle performance workstation could serve as a field computer provided the following hardware is interfaced:

- Cartridge drive for the 8mm helical scan tapes
- Plotter for seismic sections and maps

The prototype PASSCAL field computer is a Sun 3/280, with an 11" Benson thermal plotter, a monochrome large screen monitor, a 9 track 3-density tape drive, and a cartridge drive. In addition, a basic seismic processing package is being acquired as a software substrate to sit atop the Unix operating system. The choice of a widely available workstation, operating system, and processing package makes the field computers a simple extension of many university seismic computing facilities. At sites where national networks such as bitnet, nsfnet, or arpanet are available, the field computers can be connected nationally, thus providing access to the data by investigators not on the field site.

Data management

In the 1984 Program Plan (pg. 103) are given estimates of the gross data quantities involved in different deployment modes. Without justification of the parameters here, the overall numbers (*Table 2*) are instructive:

Table 2

Data quantities

Type of experiment	Typical traces per day	Duration, days	Total bytes
Refraction/reflection	600	10	72 mbyte
3-D explosion study	36,000	30	13 gbyte
Local earthquake study	1,500	300	5.4 gbyte
Surface wave teleseism study	3,000	300	10.8 gbyte

We infer:

1. Almost all programs using artificial sources will generate less than 2 gbyte of data (1 cartridge), except for programs involving particularly large scale effort and expense, such as a 3-D study.

2. Almost all programs recording natural events will generate more than 2 gbyte of data, owing to the long duration of the experiment. On the other hand, it will still be difficult to generate more than 10 gbyte of data per experiment.

3. Consequently, media capacities are not a problem. Moreover, performance improvements in the workstation field mean that throughout capability for data preprocessing will be a non-problem in the very near future.

4. The problem lies in providing the software capability for investigators to catalog, sort and edit their data. When earthquake data are being recorded, an investigator must manage as much event data as passes into the Global Seismic Network.

Consequently, data management is considered to be the responsibility of each team of investigators which has undertaken an experiment. The development of the techniques and software for such experiments will be undertaken incrementally over several years, as different teams undertake a series of increasingly ambitious studies. No centralized archive or data service can be considered until the issues concerned with management of individual experiments are sorted out.

Cost estimates:

Table 3 gives some current cost figures for the PASSCAL system elements, along with some rough cost estimates for some idealized field configurations, exclusive of costs per deployment or spares and replacements. These estimates show that the total cost for large scale portable applications runs from $23,000 to $40,000 per basic 6-channel DAS, when sensors, playback systems, and bulk recording needs are allowed for.

Table 3
Sample cost figures (1988)

System components	
Data Acquisition Subsystem	$ 15,000
Time Keeping Subsystem	$ 1,500
Auxiliary Recording Subsystem	$ 10,000
Full featured field computer	$ 80,000
Basic field computer	$ 35,000
100 km long string of vertical 4.5 hz geophones at 167m:	$2,280,000
100 DAS units	$1,500,000
100 timing units	$ 150,000
200 sets of 3 channel cable	$ 100,000
600 4.5 hz vertical geophones	$ 50,000
3 Field Computer	$ 240,000
4.7 km long, 48 channel reflection spread	$ 232,500
8 DAS units	$ 120,000
8 timing units	$ 12,000
16 sets of 3 channel cable	$ 8,000
48 8 hz geophones	$ 2,500
1 Auxiliary Recording Unit	$ 10,000
1 Field Computer	$ 80,000
Portable broad-band earthquake recording network of 50 instruments	$ 992,500
25 DAS units	$ 375,000
25 timing units	$ 37,500
25 broad-band 3-component sensors	$ 250,000
25 Auxiliary Recording Units	$ 250,000
1 Field Computer	$ 80,000

note: about $200,000 reduction would be possible if power availability permitted packaging the tape cartridges in the DAS battery compartment, and omitting the ARU's.

The PASSCÂL instrumentation is designed for use by the research community. That leads to a one-time capital cost, followed by costs of deployment. Consequently, the system design has been directed to providing both large-scale users the means of taking many different types of economies in managing field operations. For example, a 3-person university group, assisted by a local blaster, could easily conduct a 72 channel (12 unit) low fold, high resolution reflection survey on a 20 km line. On the other hand, the same data acquisition equipment lends itself to regional reflection profiling by conventional roll-along, simply by increasing the size of the crew and the number of utility vehicles. For earthquake recording, the tradeoffs involve the number of sites desired versus the degree to which low noise, ultra broad band requirements lead to large sensor and site preparation costs.

Representative Applications

The PASSCAL dataloggers are usable for a wide spectrum of geophysical field measurements in which portability or movability must be combined with high performance and flexible configurability. While they also meet the requirements for many applications with fixed sites and functions, the quantity and the irrelevance of weight and power considerations may make lower cost achievable through use of more specialized instruments. The 1984 PASSCAL program plan gave detailed examples of different types of field studies which needed this kind of hardware. Broadly, these include:

1. Conventional reflection or refraction studies: All-offset combined reflection/refraction studies with arrays of 100km aperture or more. The capability exists for vibroseis processing on board the PASSCAL dataloggers.

2. Unconventional refraction or reflection geometries: use of marine airguns fired into onshore arrays; use of quarry blasts and local earthquakes as sources fired into linear or more complex arrays.

3. Regional and local network seismic stations, or institutional seismic station with a teaching emphasis.

4. Broadband deployments for deep earth structure: regional/continental scale arrays to record teleseisms, regional events.

5. Portable arrays for studies of the propagation and generation of regional phases...in support of nuclear verification research.

6. Tomographic deployments for studies of crust/mantle structure.

7. Portable arrays for aftershock chasing or for volcano monitoring.

8. Dataloggers for EM or MT field studies.

Current Status and plans for 1989:

The 1988-89 period will involve a number of implementation activities which are needed to go from the prototype stage to an all-up operational program. These consist of:

1. Bench testing of the first prototype PASSCAL instruments.

2. Field testing in different experimental situations.

3. Determination of final required software and hardware modifications.

4. Procurement of approximately 85 systems.

5. Continued iterative development of field computer software.

6. Selection and procurement of short period sensors.

7. Testing and evaluation of candidate portable broad-band sensors.

8. RFP for and selection of the first field maintenance center.

In addition, IRIS is going through a phase of evaluating commercially available portable instruments to serve as "simple instruments". Where a large number of closely spaced channels is the maximum priority, as in certain reflection and refraction studies, it is desirable to seek portable instruments which do not conform to many of the functional specifications of the basic PASSCAL design, but which provide a maximum number of channels at a minimum cost per channel. Several manufacturers have delivered candidate dataloggers to IRIS, for consideration as "simple instruments". These are now being field tested onsite next to the five prototype PASSCAL instruments.

The goal of efforts in this period is to have available by mid-1989 a functional complement of PASSCAL hardware and software. With a target of about 85 DAS's (510 channels) and another 100 simple instruments (300 channels) IRIS plans to be ready to serve the research community as a supplier of portable seismic instrument services. Under a recent reorganization of the NSF Continental Lithosphere Program, all proposals for use of the PASSCAL facilities, whether large-scale or not, would be reviewed by a special NSF panel, and IRIS would serve as a facilities manager to make available the needed hardware and software support. With the number of channels targeted for mid-1989, it should be possible to make PASSCAL arrays available for several different groups right at the start.

During this period, technical advances are continuing to move the PASSCAL instrument toward even better characteristics than anticipated six months ago. The availability of the 2.3 gbyte 8mm helical scan video cartridge as a reliable, low cost technology, was established only early this year. This in turn opened up the possibility of encapsulating a cartridge drive in the battery compartment when external power is available. Advances in board packaging since last year have made it possible to squeeze the DAS components onto smaller boards, thus permitting a substantial decrease in the dimensions, and consequently, the weight of the instrument which will be purchased in quantity.

Changes in methods of conducting field seismic studies

The enormous increase in capability represented by the new instruments brings with it some changes in the way research groups acquire and process data. The field setup terminal has so many options that the field operator must be adequately trained in its use if errors are to be avoided. The large number of instruments out in each deployment imposes a requirement that the organization of the field team activities and instrument setup be carefully structured to insure that the standardization designed in the instruments is actually achieved in the field. Except for the smallest projects, the use of these instruments can never be undertaken casually, and learning be done "on the job". The data playback phase must also be planned and carried out by agreed protocol, if the quality control and rapid processing capabilities of the field computer are to be realized. The enhancement of channel densities by factors of 10 for most types of seismic application leads to a need in most cases for thoroughly redesigned processing and interpretation methods. These remarks will perhaps give a slight flavor of the kind of team organization and discipline which we recognize as part of "big science", and which seismologists will need to sample if they are to enjoy the fruits of the new technology.

Plans for 1989-1993

By 1993 it is planned for the PASSCAL national facility to have a complement of 500 instruments (3000 channels), along with perhaps another 1000 channels in simple instruments. About 100 of the instruments are to be outfitted with 3-component broad-band portable sensors and 24 bit encoders, to permit their specialization as a portable array for detailed studies of deep earth structure using teleseisms and regional earthquakes. Probably ten field computers and five regional maintenance centers will come into existence. The complement of different timing systems and short period sensors will evolve from year to year, based on the kinds of projects which are approved for support by the NSF.

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LONG-TERM RECORDING OF SEISMIC DATA ON THE OCEAN FLOOR

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Introduction

A continuously recording broad-band station will collect data at a rate of approximately 30 MB per day. With effective event triggering this can be reduced by a factor of ten, resulting in a requirement to store or telemeter about 1.1 GB of data per year. Given the remoteness of the sites required for a useful global network (see cover of this report), the determination of how these data can be reliably recorded, either in situ or at some land-based site following transmission by cable or satellite, is a fundamental component of any plan to install and operate a set of permanent ocean floor observatories. Our experience with making very long-term seismic observations on the ocean-floor is extremely limited: a unique case is the Columbia-Point Arena Ocean Bottom Seismic Station that operated for over six years and is discussed in a paper by George Sutton in Section B3 of this report. But this is an exception. The vast majority of experience within the community lies in the operation of internally recording ocean bottom seismometer (OBS) instruments that are rarely deployed (at least intentionally!) for durations in excess of 1-2 months. There exist 10-20 groups worldwide that operate their own design of OBS: it is not the intention of this article to attempt a review of these widely varying instruments. Suffice it to say that none have been built for permanent observatory-type operations and almost all have negligible response below 1-2 Hz. The notable exception to this is the Scripps Institution of Oceanography OBS, that using 1 Hz geophones, has useful response down to 10-20 secs period.

Satellite Telemetry or Cable Telemetry or Internal Recording?

Given both the permanence of the planned observatories and the large volumes of data to be handled, it is natural to consider modes of operation other than conventional internal recording. The article by Frye and Briscoe in this same section of the workshop report makes it clear that existing technology in data telemetry by satellite falls far short of meeting our needs, except perhaps for special applications (real-time warning of the detection of special types of events, for example). There is no doubt that the preferred solution to data acquisition is the use of permanent ocean floor cables. Data is obtained in real-time, data compression and event triggering processes can be carried out in a shorebased laboratory where substantial computing resources are available and human intervention is simple, and power can be supplied to the sensor modules, thus removing (or at least reducing) the need for the development of new micropower broad-band seismometers. The clear disadvantage with the use of cable telemetry is the cost of cable emplacement. The accompanying article by Nagumo and Walker makes a case for overcoming this difficulty by using the existing global telephone cable network. However, a comparison of their Fig. 1 with the cover of this report shows that even if it should prove feasible and economic to do this, there would remain many important portions of the world's oceans without coverage, primarily in the southern oceans. If the cost of installing dedicated cables to these southern ocean sites is judged prohibitive, then the option must be investigated of recording the data in situ for periodic collection by a research vessel.

The New ONR Ocean Bottom Seismometer

The U.S. Office of Naval Research is supporting the development of a new OBS for use in studies of ultra low frequency inertial and acoustic noise on the deep ocean floor.

This project is a cooperative effort between Woods Hole Oceanographic Institution, Scripps Institution of Oceanography, University of Washington and Massachusetts Institute of Technology. It is planned to have approximately 30 of these new instruments available for deployment in 1990. Many of the design criteria placed on this instrument would apply to a recording system needed for a permanent global network.

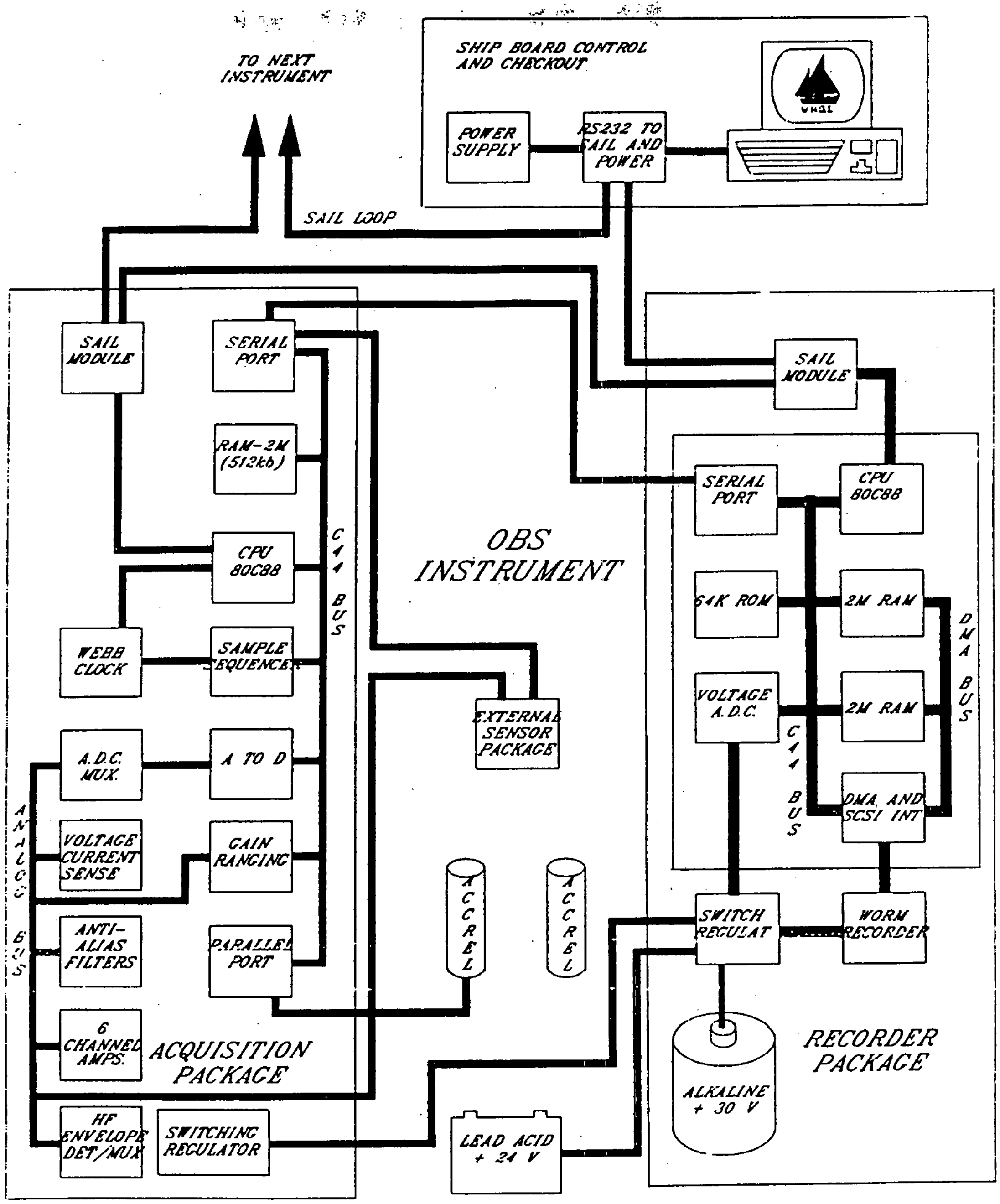
It will be capable of one year long deployments and with the use of the Webb Research Corp. BVA clock will retain timing accuracy of a few milliseconds over this period. It has a six channel primary A to D system and two additional channels for low rate auxiliary/environmental data acquisition. The primary system uses 14 bit digitization (effectively) and active gain-ranging to give a total dynamic range of approximately 120 dB. The recording medium is optical disc with a total capacity of 400-800 MB depending upon the outcome of tests ongoing at the time of writing. The overall architecture of this instrument has been influenced greatly by the need for flexibility in future usage (Fig. 1). It is modularized into five independent systems: sensor, which currently is planned to be triaxial 1 Hz geophones with a Cox-Webb long period pressure sensor and a conventional OAS hydrophone; acquisition package which is a separate 7" ID pressure case containing the analog amplifiers, the BVA clock and the ONSET 80C88 that controls all acquisition functions; recording package which again is a separate 7" ID cylindrical pressure case that receives the data down a serial data link from the acquisition package and contains the optical disc drive(s), a second ONSET 80C88 to control the recording and the batteries; and lastly two completely independent acoustic releases, each one of which can bring about instrument recovery, and which provide ranging information via their transponders for instrument location. Present plans call for arranging these modules on a single fibreglass frame with conventional glass ball buoyancy (Fig. 2). But it would be simple to rearrange the system such that the sensor and acquisition package remained permanently on the ocean floor, and were connected by several kilometers of low-cost coax cable to a recoverable unit containing the batteries and recording unit. Every year or so, a vessel could recover the recording package, replace the batteries and recording medium, check proper operation of the acquisition system and redeploy without having to disturb the sensor.

Operation of a Global Network

The data capacities that we plan today (~800 MB) are already tantalizingly close to the 1.1 GB needed for a one year operation of a global network station, and new deployments in high density magnetic cartridge tape drives and the continuing development of optical disc drives will undoubtedly close this gap in the next 2-3 years. The accurate timing problem is solved with the advent of the BVA clock. The power requirements for recording are manageable using conventional batteries in reasonable sized pressure cases. The primary disadvantage is the need for servicing at regular intervals (~ 1 year) for battery and recording medium replacement. Given the harshness of the deep ocean environment it may prove that servicing at comparable intervals may be required, anyway, just to overcome the inevitable system failures. This will undoubtedly be true in the early years. With a station distribution like that depicted on the front cover of this report, the servicing of 20 observatories would require approximately 200 days of ship time. If advantage were taken of ships of opportunity this could be reduced substantially. Maintenance and operation of a full global network using remote internally recording systems could then be achieved at a cost (including ship time) of approximately $3M per year.

Conclusions

The most feasible and economical approach to the operation of a global ocean-floor network is to use high-capacity recording devices which are refurbished, and from which the data is recovered, every 1-2 years. In special cases, during instrument development,



TO NEXT INSTRUMENT
SHIP BOARD CONTROL AND CHECKOUT
POWER SUPPLY
RS232 TO SAIL AND POWER
SAIL LOOP
SAIL MODULE
SERIAL PORT
RAM-2M (512kb)
CPU 80C88
WEBB CLOCK
SAMPLE SEQUENCER
A.D.C. MUX.
A TO D
VOLTAGE CURRENT SENSE
GAIN RANGING
ANTI-ALIAS FILTERS
PARALLEL PORT
6 CHANNEL AMPS.
HF ENVELOPE DET/MUX
SWITCHING REGULATOR
C 4 4 BUS
ANALOG BUS
ACQUISITION PACKAGE
OBS INSTRUMENT
EXTERNAL SENSOR PACKAGE
ACCREL
ACCREL
LEAD ACID + 24 V
SAIL MODULE
SERIAL PORT
CPU 80C88
64K ROM
2M RAM
VOLTAGE A.D.C.
2M RAM
DMA AND SCSI INT
DMA BUS
C 4 4 BUS
SWITCH REGULAT
WORM RECORDER
ALKALINE + 30 V
RECORDER PACKAGE

Figure 1

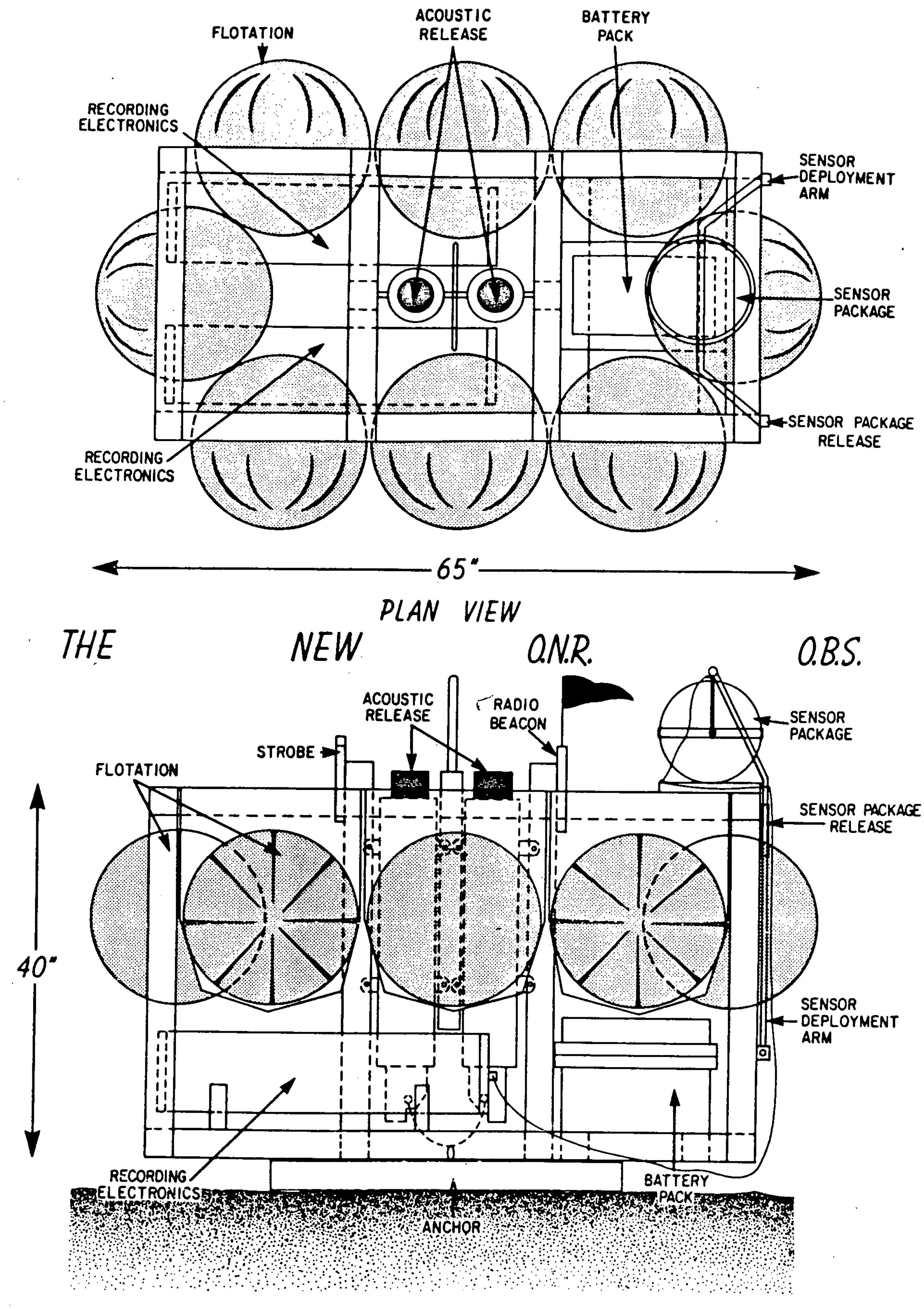

FLOTATION
ACOUSTIC RELEASE
BATTERY PACK
RECORDING ELECTRONICS
SENSOR DEPLOYMENT ARM
SENSOR PACKAGE
SENSOR PACKAGE RELEASE
RECORDING ELECTRONICS
65"
PLAN VIEW
THE NEW O.N.R. O.B.S.
ACOUSTIC RELEASE
RADIO BEACON
STROBE
SENSOR PACKAGE
FLOTATION
SENSOR PACKAGE RELEASE
40"
SENSOR DEPLOYMENT ARM
RECORDING ELECTRONICS
BATTERY PACK
ANCHOR

Figure 2

when immediate knowledge of failure is important, or for the mission of tsunami warning, real-time data telemetry is necessary and is only practical at this time using ocean-floor cables. Although little experience of such high-capacity, long-term instrument deployments exists in the marine seismology community at this time, and some instrumentation development would be required before it was practical, it is clearly a realizable goal that is not dependent for its success upon new technological breakthroughs. Without question the primary technological challenge that faces us at this time is the development of a reliable, rugged, micropower broadband sensor of suitable dimensions for emplacement downhole.

THE DESIGN OF A GLOBAL SEISMOGRAPHIC NETWORK STATION

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The Science Plan for New Global Seismographic Network (IRIS, 1984) documented the need for a broad band station with certain characteristics. This issue was followed up in greater detail in the "Design Goals for GSN" (IRIS, 1985), which lists several basic requirements for the system:

1. The sensitivity must be sufficient to resolve seismic signals at the level of the lowest ambient noise within the band from 0.3 mHz to 5 or 10 Hz.

2. The system should record on scale the largest teleseismic signals - say, an earthquake with M_w - 9 at 30°.

3. The linearity of the system should be such that signals near the ground noise minimum can be resolved in the presence of the maximum ground noise at other frequencies (microseismic storms).

Wielandt and Steim (1986) presented evidence that it is possible to meet these requirements with a single data stream generated by a feedback system with a response flat to velocity between 5 Hz and 360 sec period; they call it a very-broad-band response: VBB.

The reason for presenting this relatively detailed description of a GSN Station in the context of this workshop is the hope that some of its features such as the VBB response could be implemented in the ocean bottom version. The division of the roles between the Data Acquisition and Data Processing units may be of help in designing a system with real time telemetry.

In the spring of 1986 the "Design Goals" were translated into design specifications, and submitted to IRIS's administration. In April 1987, following appropriate procurement procedures, IRIS entered into a contractual agreement with Gould, Inc., for construction of a prototype station processor. The choice of the architecture of the system was guided by several considerations:

1. meeting of the scientific objectives;

2. reliability and ease of maintenance;

3. modularity and minimal dependence on unique products;

4. feasibility of future upgrades and modifications of the chosen elements of the system;

5. scientific and operational benefits for the host organization.

Figure 1 is a block diagram of the system. The two principal components are the data acquisition unit (DA) and the data processing unit (DP). They are connected by a serial link or by any form of telemetry, from a telephone line to a satellite. The following is a brief description of the major features of a GSN station.

Data Acquisition (DA) Module

The DA is the heart of the system. With only minor software modifications, with respect to the agreed upon specifications, and the addition of a tape recorder, this unit would be, in principle, capable of fulfilling the principal objective of the entire system, namely, the acquisition and recording of the data.

The major components of the DA system are:

1. A very broad-band seismic sensor system. The STS-1V(H)/VBB, built by Streckeisen & Co., is a set of sensors that satisfy the design goals of IRIS for installation in a vault and has a sufficient linearity and dynamic range to accommodate in a single data stream frequencies from D.C. to 10 Hz.

2. A high resolution digitizer/calibrator unit (HRDCU). This is a three-channel 24-bit formatted digitizer/calibrator for use with the VBB sensors. There is also a fourth channel for recording the calibration current. Its data rate is twenty samples/second (sps) per channel.

3. Kinemetrics OMEGA UTC time receiver.

4. Auxiliary digitizer unit (AUXDU). This is for monitoring various state-of-health parameters of the VBB system and, optionally, other parameters such as atmospheric pressure or geomagnetic field. The data rates are programmable.

5. Low-resolution digitizer-calibrator unit (LRDCU, optional). This is for digitization of the output of additional sensors.

6. Very short period (VSP) sensor system (optional). "Design Goals" suggested a sampling rate of 100 sps. Consideration is being given to increasing it up to 200 sps.

7. Low Gain (LG) sensor subsystem. This was originally thought of as a means to extend the dynamic range of the VBB system, and it could serve as a strong motion monitor. The sampling rate is the same as that of the VSP system. There could be a limitation on the aggregate data rate of the VSP and LG channels (about 600 sps total - all sensors and all components).

8. Data Acquisition unit (DA). This is based on a 68000 or 68020 CPU with a VME-bus, operating under an OS-9 operating system and special application software. The DA processor interfaces with the three GSN digitizers (HRDCU, LRDCU, and AUXDU) and with the data processor unit (DP). All interfaces are bi-directional. The major data manipulation functions of the DA processor are to collect data from the three digitizers, time tag data blocks, compress VBB data, perform VSP and LG event detection, and transmit data to the DP.

Data Processor (DP) Module

The DP module provides the means for on-site and remote monitoring of the system performance, communications with remote users, digital and analog recording of data, maintenance of data buffers, display of data for a selected time window on a graphic terminal, and many other functions. Its principal role is the enhancement of the functions of the DA unit, on-site quality control, and providing the station personnel and remote users with information on seismic events.

The station DP consists of commercially available hardware modules and is based on a standard VME bus. Figure 2 gives some details of the DP configuration. The computing power of the 68020 chip is such that only a small fraction of its capacity will be used for the planned tasks, leaving much room for future expansion. Even now it is a fairly complex, sophisticated system.

Figure 3 is a block diagram of the DP unit's data manipulation functions. The VBB data are decompressed, so that they can be processed through a series of digital filters which produce short period (SP), long period (LP; 1 sps), and very long period (VLP; 0.1 sps) data. The LP and VLP data streams are obtained using FIR filters with a very sharp corner near the edge of the pass-band. This assures retention of maximum information in the pass-band and may modify the way in which seismologists evaluate the usefulness of various data streams.

Figure 4 is taken from Steim (1986). It shows the original VBB channel and five channels derived from it. There is a twenty second delay caused by a FIR filter. The trace labeled LP is obtained from the VBB stream by using a 201 points FIR filter. It is clear that it contains most of the information present in the VBB stream, but with twenty times fewer samples. It very well may be that this will be the most frequently used data stream, particularly because of its relatively light data transmission or archival burden. The other data streams shown contain less high frequency energy, with the simulated SRO-LP channel representing an extreme. The VBB response is particularly convenient for a stable recovery of the ground displacement function (bottom trace).

The command, control, and algorithmic functions executed by DP are shown in Figure 5. Rather than discuss them in detail, let us list functions that the operator can execute from the system console without interrupting acquisition of data:

1. adjust the scale and select active analog monitor channels, including the selection of simulated WWSS and SRO response function;

2. view a continuously updated full screen status display that shows a snap-shot of all data channels, the internal and received UTC time, and several other system parameters;

3. view the system event log;

4. change the tape cartridges without loss of data;

5. examine the status of active processes;

6. view selected data waveforms from buffers or in real-time;

7. set, change or display event detection parameters;

8. exchange message text over the real-time and dial-up ports;

9. control and set up a calibration cycle, or program the onset time of a calibration sequence to be recorded with the station data;

10. run a calibration analysis as a low-priority background job;

11. log messages and the results of calibration to the system's mass storage device.

Many of the functions described above can be performed through a dial-up port, which allows for frequent checks of station performance from the network maintenance center.

The prototype of the GSN system will be delivered in the fall of 1988, and following the tests, production units will be ordered and the first ten, budget permitting, deployed in 1989.

Figure Captions

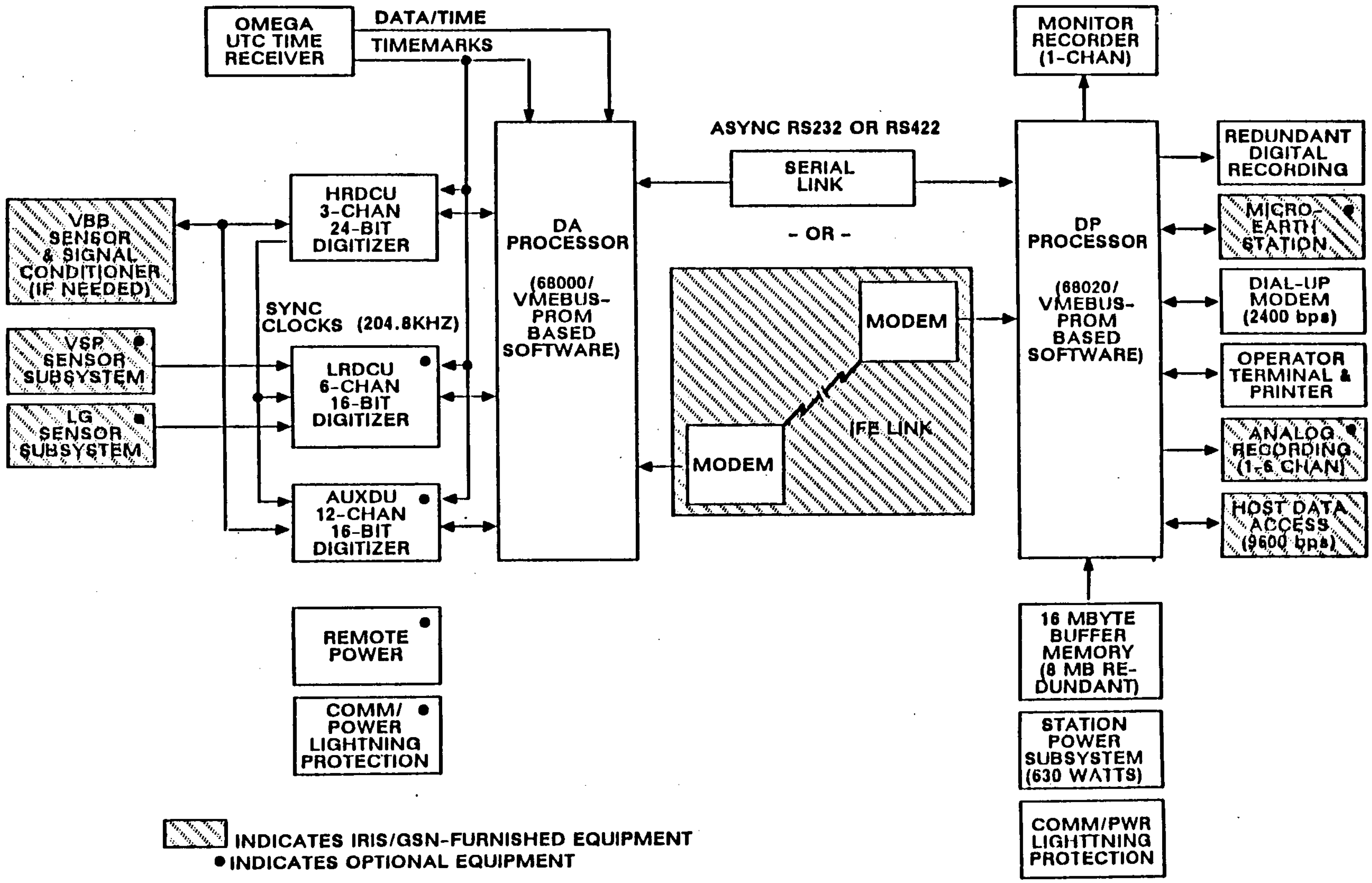
Figure 1. Block diagram of the prototype GSN station. From report U87-260 by Gould Inc.

Figure 2. Configuration of the GSN station data processor (DP). From report U87-260 by Gould Inc.

Figure 3. Data processing functions carried out by the DP unit. From report U87-260 by Gould Inc.

Figure 4. The information content of the "broad band" 1 Hz LP data stream, obtained by processing VBB data with a FIR filter with a very sharp roll-off near the edge of the pass-band. The four lower traces, showing alternative responses, are derived from the LP data stream. The 20 sec delay is due to the finite length of the FIR filter. From Steim (1986).

Figure 5. Command, control and diagnostic function structure within the data processor (DP) unit. From Gould Inc. report U87-260.



OMEGA UTC TIME RECEIVER
DATA/TIME
TIMEMARKS
VBB SENSOR & SIGNAL CONDITIONER (IF NEEDED)
VSP SENSOR SUBSYSTEM
LG SENSOR SUBSYSTEM
HRDCU 3-CHAN 24-BIT DIGITIZER
SYNC CLOCKS (204.8KHZ)
LRDCU 6-CHAN 16-BIT DIGITIZER
AUXDU 12-CHAN 16-BIT DIGITIZER
DA PROCESSOR (68000/ VMEBUS-PROM BASED SOFTWARE)
ASYNC RS232 OR RS422
SERIAL LINK
- OR -
MODEM
IFE LINK
MODEM
DP PROCESSOR (68020/ VMEBUS-PROM BASED SOFTWARE)
MONITOR RECORDER (1-CHAN)
REDUNDANT DIGITAL RECORDING
MICRO-EARTH STATION
DIAL-UP MODEM (2400 bps)
OPERATOR TERMINAL & PRINTER
ANALOG RECORDING (1-6 CHAN)
HOST DATA ACCESS (9600 bps)
16 MBYTE BUFFER MEMORY (8 MB RE-DUNDANT)
STATION POWER SUBSYSTEM (630 WATTS)
COMM/PWR LIGHTTNING PROTECTION
REMOTE POWER
COMM/ POWER LIGHTNING PROTECTION
INDICATES IRIS/GSN-FURNISHED EQUIPMENT
• INDICATES OPTIONAL EQUIPMENT

Figure 1

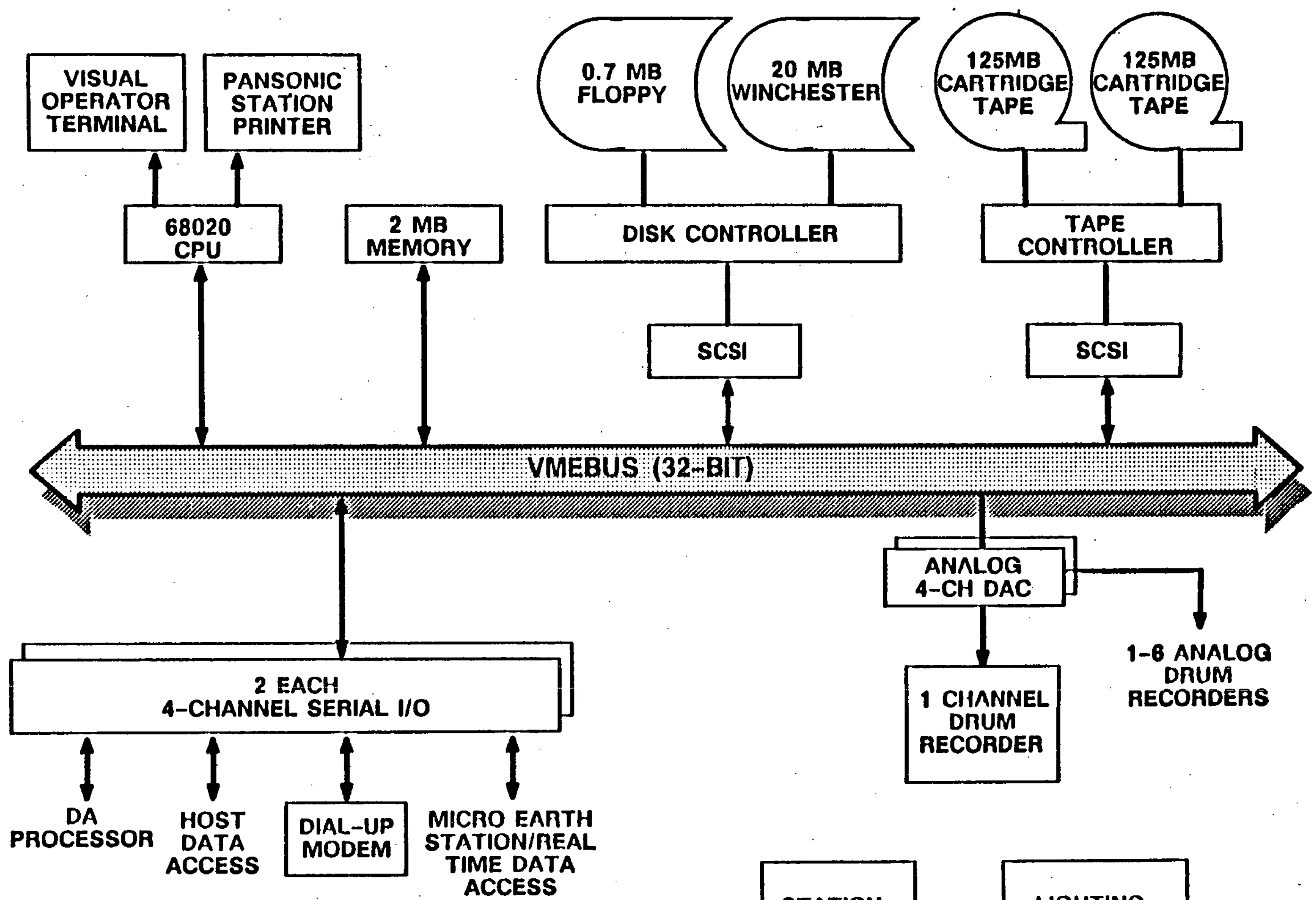

VISUAL OPERATOR TERMINAL
PANSONIC STATION PRINTER
68020 CPU
2 MB MEMORY
0.7 MB FLOPPY
20 MB WINCHESTER
125MB CARTRIDGE TAPE
125MB CARTRIDGE TAPE
DISK CONTROLLER
TAPE CONTROLLER
SCSI
SCSI
VMEBUS (32-BIT)
ANALOG 4-CH DAC
1-6 ANALOG DRUM RECORDERS
1 CHANNEL DRUM RECORDER
2 EACH 4-CHANNEL SERIAL I/O
DA PROCESSOR
HOST DATA ACCESS
DIAL-UP MODEM
MICRO EARTH STATION/REAL TIME DATA ACCESS
STATION POWER
LIGHTING PROTECTION

Figure 2

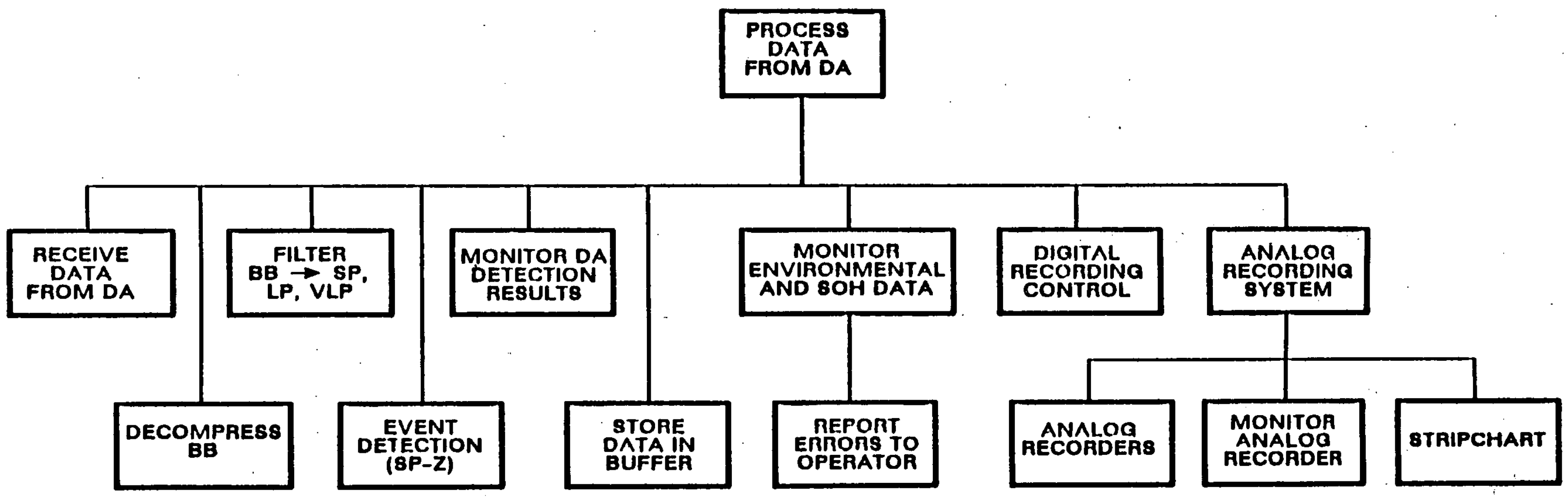

PROCESS DATA FROM DA
RECEIVE DATA FROM DA
DECOMPRESS BB
FILTER BB → SP, LP, VLP
EVENT DETECTION (SP-Z)
MONITOR DA DETECTION RESULTS
STORE DATA IN BUFFER
MONITOR ENVIRONMENTAL AND SOH DATA
REPORT ERRORS TO OPERATOR
DIGITAL RECORDING CONTROL
ANALOG RECORDING SYSTEM
ANALOG RECORDERS
MONITOR ANALOG RECORDER
STRIPCHART

Figure 3

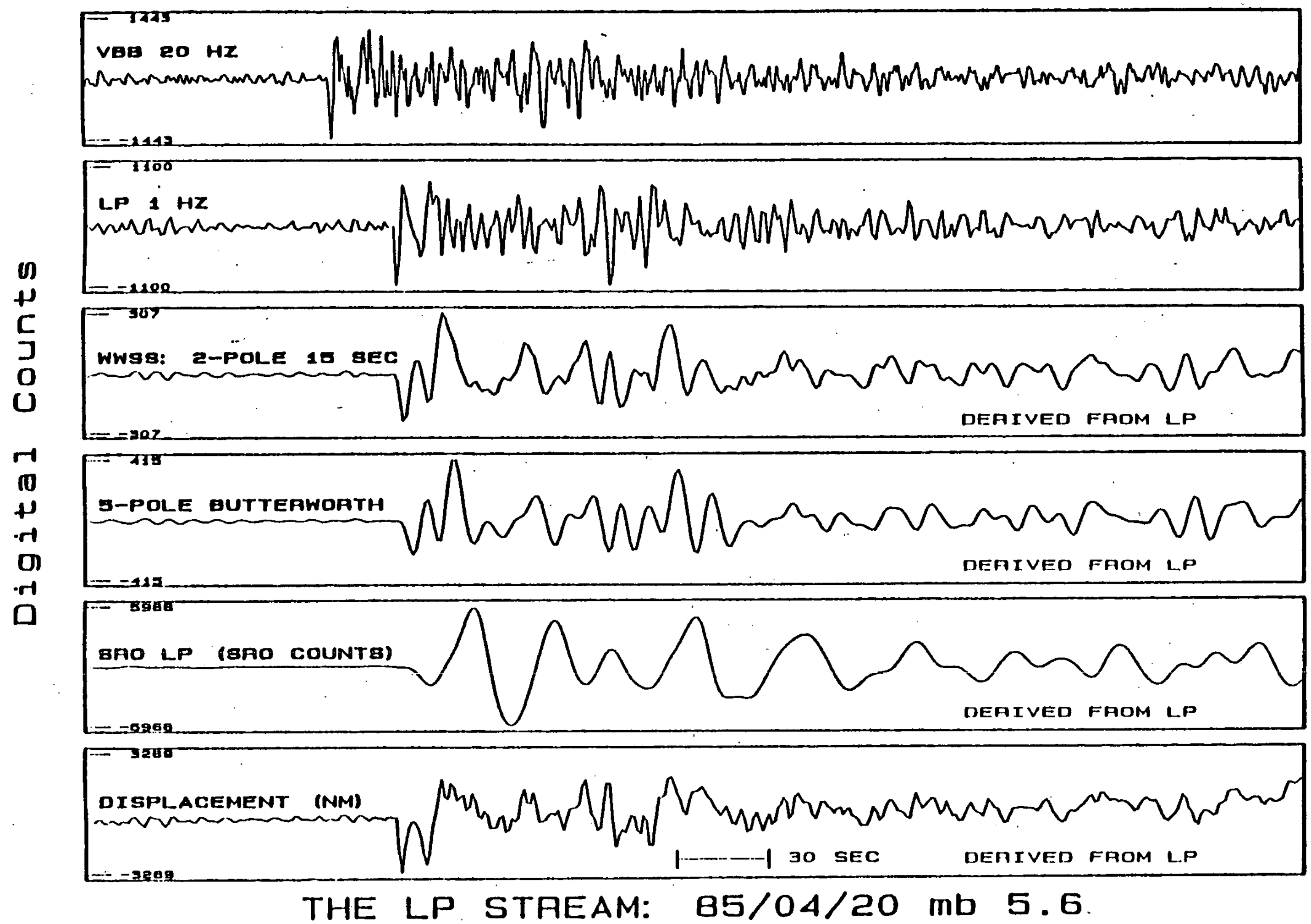

Digital Counts
1443
VBB 20 HZ
-1443
1100
LP 1 HZ
-1100
307
WWSS: 2-POLE 15 SEC
DERIVED FROM LP
-307
415
5-POLE BUTTERWORTH
DERIVED FROM LP
-415
5988
SRO LP (SRO COUNTS)
DERIVED FROM LP
-5988
3288
DISPLACEMENT (NM)
30 SEC
DERIVED FROM LP
-3288
THE LP STREAM: 85/04/20 mb 5.6

Figure 4

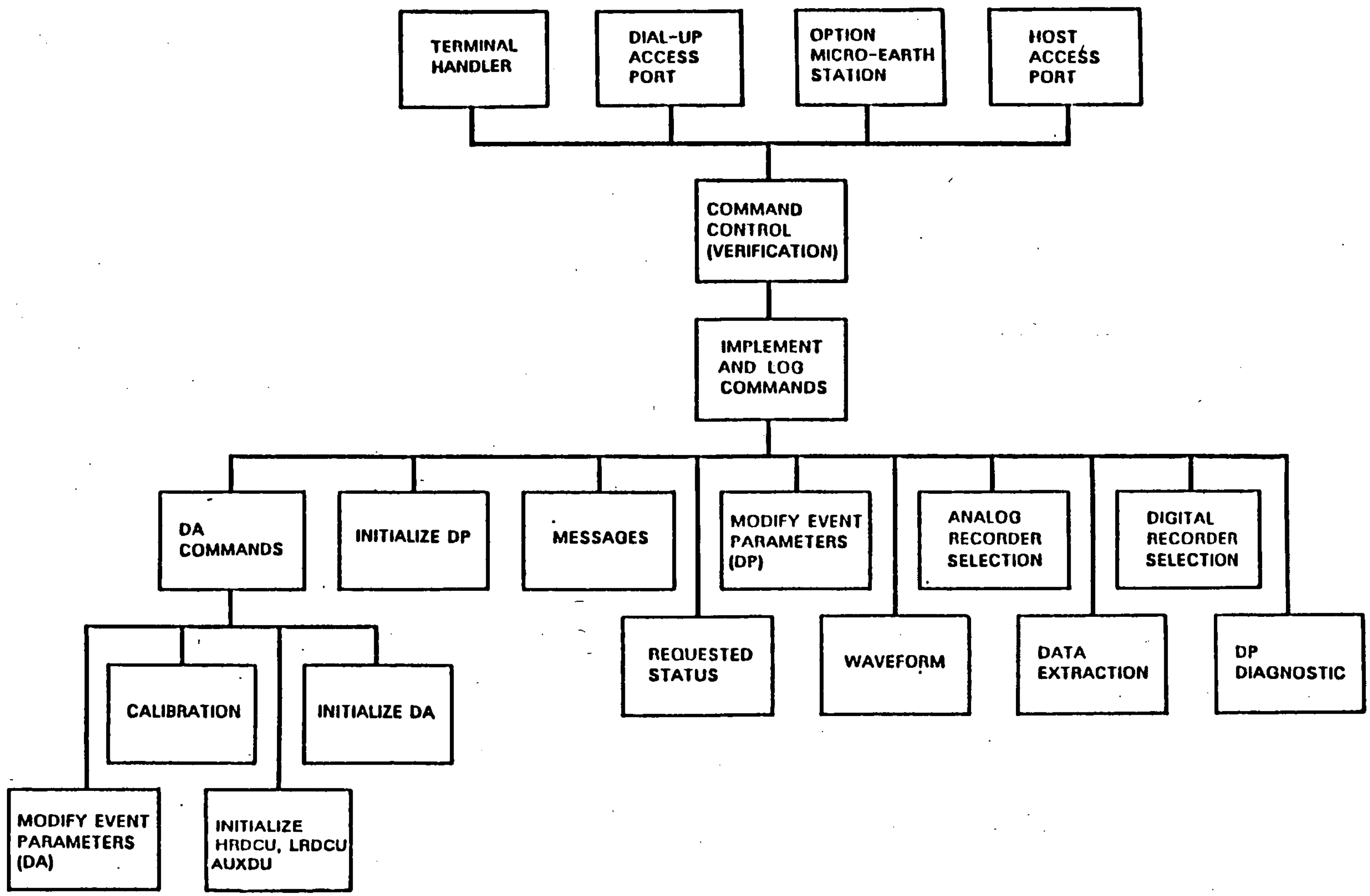

TERMINAL HANDLER
DIAL-UP ACCESS PORT
OPTION MICRO-EARTH STATION
HOST ACCESS PORT
COMMAND CONTROL (VERIFICATION)
IMPLEMENT AND LOG COMMANDS
DA COMMANDS
INITIALIZE DP
MESSAGES
MODIFY EVENT PARAMETERS (DP)
ANALOG RECORDER SELECTION
DIGITAL RECORDER SELECTION
REQUESTED STATUS
WAVEFORM
DATA EXTRACTION
DP DIAGNOSTIC
CALIBRATION
INITIALIZE DA
MODIFY EVENT PARAMETERS (DA)
INITIALIZE HRDCU, LRDCU AUXDU

Figure 5

Section B6: Other Measurements

1. The Ocean Drilling Program: R.P. Von Herzen
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THE OCEAN DRILLING PROGRAM

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The Ocean Drilling Program is a major geosciences project sponsored primarily by the National Science Foundation, and operated by Texas A&M University on behalf of a consortium of U.S. universities and oceanographic institutions. It also has the support and participation of 5 non-U.S. organizations in Great Britain, France, West Germany, European Science Foundation, and Japan. Scientific ocean drilling began about 2 decades ago when it first proved possible to drill holes into the seafloor with existing technology. Since that time about 750 holes (sites) have been drilled for scientific purposes in all the world oceans. Notable achievements include the deepest penetration in ocean sediments, about 1.7 km off the northwest African margin, and a maximum penetration of 1.2 km in basement rocks on the Costa Rica rift southwest of Panama. A small fraction of the sites (approximately 30) are so called re-entry sites, i.e., they have a funnel above the hole on the seafloor which allows the hole to be re-entered by pipe suspended from the drill ship, with the aid of underwater TV. Perhaps 12 to 15 of these holes are open to the depths to which they were originally drilled (Table 1)*. These holes represent a resource for scientific purposes, such as placing of seismometers down hole.

The technical capabilities of the present drilling vessel, the D/V JOIDES Resolution (SEDCO BP/471), are notable. More than 9 km of drill pipe can be suspended from the vessel. The hole diameter generally drilled is about 10 to 12 inches, but it is possible to drill holes with diameters up to 18 inches or even more. A casing (metal liner) can be emplaced in holes to improve stability for re-entry or other scientific purposes. The drill vessel operates without a riser (casing to the surface around the drill pipe), so that return circulation of the drilling fluids (normally seawater) is not possible. The drill pipe inside diameter, and clearances to enter it through the top drive and rig floor, are about 4-1/8 inches. Normally this allows about 3-1/2 inch to 3-3/4 inch diameter tools to clear through the inside of the drill pipe, depending on their length. The drill ship has dynamic positioning, i.e., it employs thrusters under computer control to maintain position over the seafloor within about 1% of the water depth in almost all weather conditions, using a bottom acoustic beacon for reference. As part of the pipe handling system, a heave compensator allows vertical ship motion up to 7 to 8 m amplitude caused by ocean swells to be nearly cancelled. Coring and geophysical logging by lowering tools with cables through the drill string have received much attention are are relatively routine. Shipboard computing facilities include a VAX 11/780, with many terminals available for time shared use. High latitude regions have been drilled with the assistance of an ice-picket vessel to survey for ice, even diverting small icebergs from interference with drilling.

As mentioned above, the drilling vessel is operated on behalf of an international consortium of partners. Planning for the vessel operations is centered around an executive and planning committee of the Joint Oceanographic Institutions for Deep Earth Sampling (JOIDES), which is normally chaired by the chairman of the planning committee (presently R. Moberly, University of Hawaii). Proposals for vessel use are directed to the planning committee, which usually refers them to appropriate panels which are in place to represent various regions and themes of the drilling program. The drilling vessel programs and regions of operations begin to be planned on the order of 4 to 5 years in advance, and are firmed up for specific drilling legs 1 to 2 years before drilling. Legs are normally about 50-55 days in duration, according to contractual agreements with the ship operators.

Instrumental installations in boreholes have been encouraged by the JOIDES organization, particularly where they contribute to increased knowledge of the earth. For example, a DARPA-funded program to install seismometers in existing drillholes in the Pacific was proposed and carried out successfully with seismometer packages deployed from the end of the drill pipe.

*Table 1 from report of workshop "Science opportunities created by wireline reentry of deep sea boreholes" (co-convenors M.G. Langseth, F.N. Spiess), Scripps Institution of Oceanography, 1987.

TABLE 1: CONDITION OF EXISTING JOIDES SEAFLOOR RE-ENTRY HOLES: (June 1987)

146 Leg 15, 15° 06.99'N, 69° 22.67'W (Caribbean). Water depth 3957 m. 762 m sub-bottom, 22 m sub-basement. Hole open. Old style cone. No logging or downhole experiments done.

288A Leg 30, 5° 58.35'S, 161° 49.53'E Steward Basin and adjacent Ontong Java Plateau). Water depth is 3030 m and penetration below seafloor is 998.5 m. From 0 m to 550 m are Cenozoic nannofossil-foraminiferal chalk and ooze with some chert. Apparently, this hole is bridged at 105 m below seafloor.

319A Leg 34, 13° 01.04'S, 101° 31.46'W: (Bauer Deep between the East Pacific Rise and the Galapagos Spreading Center). Water depth is 4296 m (pipe) and penetration below the sea floor is 157 m. From 110 m to 0 m are Miocene to Quaternary nannofossil forminefera ooze and clay. From 157 m to 110 m, basalt. The drill pipe was torquing when the site was abandoned; there may be loose basalt pieces or formation bridging in the hole.

320B Leg 34, 9° 00.4'S, 83° 31.8'W: The site is on the east edge of the Nazca Plate (Peru Basin). Water depth 4487 m and penetration bsf is 183.5 m. From 155 m to 0 m are Oligocene to Quaternary nannofossil and radiolarian ooze. From 183.5 m to 155 m, basalt. The hole was abandoned after the drill pipe was sticking in the hole, which was possibly caused by loose pieces of basalt falling into the hole, thus the hole may not be completely open.

332B Leg 37, 36° 52.8'N, 33° 38.6'W (Mid-Atlantic Ridge near FAMOUS area). Water depth 1841, hole depth 721 m, 582 m into basement. Casing is parted, little likelihood of re-entry.

333A Leg 27, 36° 50.45'N, 33° 40.05'W (MAR, FAMOUS Area). Water depth ~1666 m. 529 m sub-bottom, 310 m sub-basement. Top 260 m of hole in basement open, but bottom 50 m plugged with stuck pipe. No logging or downhole experiments done.

395A Legs 45 and 78B, 22° 45.35'N, 46° 04.90'W (MAR). Water depth 4485 m. 664 sub-bottom, 576 m sub-basement. 120 m casing (cemented into basalt). Hole is open; drilling terminated because of excessive torquing; sinker bar now at bottom. Extensive program of logging and downhole experiments completed on Leg 78B and Leg 109.

396B Leg 46, 22° 59.14'N, 43° 30.90'W (MAR). Water depth 4465. 405.5 m sub-bottom, 255 m sub-basement, 153 m casing (cemented into basalt). Hole open; no gear in hole, but lost because of sand and gravel in bottom. Logged but no positioning devices used. French plan to re-enter this hole in 1988.

415A Leg 51, 31° 01.7'N, 11° 39.1'W (Atlantic margin, off Morocco). Water depth 2817 m. 1074.5 m sub-bottom, basement not reached. Cone buried or flush with seafloor, hole filled with cavings below 900 m, barite mud above. No downhole experiments run but hole logged through pipe with natural gamma tool.

416A Leg 51, 32° 50.2'N, 10° 48.1'W (Atlantic margin off Morocco). Water depth 4203. 1624 m sub-bottom, basement not reached. Hole filled with drilling mud, occasional bridges. Logged but no downhole experiments run.

417D Leg 51 and 52, 25° 06.69'N, 68° 02.82'W (Bermuda Rise). Water depth 5489 m; 708.5 m sub-bottom, 363 m sub-basement. Top 250 m of hole in basement open, but bottom 100 m plugged by lost BHA. BHA possibly fishable. Sediments and top 125 m of basement logged (except for density). Oblique seismic experiment run in to 230 m of basement.

418A Legs 52 and 53, 25° 02.08'N, 68° 03.45'W (Bermuda Rise). Water depth 5519 m; 868 m sub-bottom, 544 m sub-basement. Hole revisited on Leg 102, bridges in the hole, but no logging tool was in hole as suspected. Oblique seismic experiment run.

433C Leg 55, 44° 46.0'N, 170° 01.26'E (Suiko Seamount). Water depth is 1861.8 m. Penetration below the sea floor is 550.5 m. From 550 m to 163 m are basalt flows with minor sand layers. From 163 m to 53 m are Oligocene reefal sand and minor sandy mud. From 52 m to 0 m are Miocene to Pliocene siliceous nannofossil chalk ooze. Hole 433C is probably open, but may have zones of fractured loose basalt which may make re-entry difficult.

442B Leg 58, 28° 59.04'N, 136° 03.43'E (Shikhoku Basin). Water depth is 4634.5 m, penetration 455.0 m. From 455 m to 355 m are pillow-basalt flows overlain by 5 m of Miocene limestone. From 375 m to 0 m are Miocene to Quaternary mud and clay, with minor ash. This hole is probably open, however, it was abandoned after the drillstring began torquing, thus there could be a bridge somewhere in the hole. Needs a surface casing.

462A Leg 61, 7° 14.5'N, 165° 01.9'E (Nauru Basin). Water depth 5186 m, penetration 1068.5 m to 665 m dolerite sills and basalt flow, 665 m to 450 m Maastrichtian volcanoclastic sediment, 450 m to 390 m Eocene chalk with minor chert, 300 m to 0 m calcareous siliceous ooze. Sonic, caliper, natural gamma and neutron logs were run. Hole was visited in Leg 89 and found to be open to bottom. The hole was deepened on Leg 89.

482D Leg 65, 22° 47.31'N, 107° 59.51'W (12 km E of East Pacific Rise axis). Water depth 3008 m, penetration 186.5 m. From 186.5 m to 137 m are massive flows and pillow basalts. 137 m to 0 m Pleistocene hemipelagic clay. Upper 39 m of hole is open, bottom part of hole contains BHA. Hole 482C contains the HIG seismometer.

483B Leg 65, 22° 53.0'N, 108° 14.8'W (Tamayo Fracture zone). Water depth 3084, total depth of hole 267 m and basement penetration is 160 m. The re-entry cone is blocked, but it may be easy to clear.

504A Leg 69, 1° 13.6'N, 83° 43.95'W (Costa Rica Rift flank). Water depth 3458 m, penetration 277 m. 277 m to 264 m basalt flows, pillows and breccias, 264 m to 145 m Miocene to Pliocene limestone and chert, nannofossil chalk, 145 m to 0 m Pliocene to Pleistocene ooze. Hole is open but drill bit cone is in the bottom of the hole. Density, caliper, gamma ray, and temperature logs have been run.

504B Legs 69, 70, 83, 92 and 111, 1° 13.61'N, 83° 43.81'W (S. flank Costa Rica Rift). Water depth 3460 m, penetration 1562.3 m. 1562.3 to 264 m basalt basement, flows and pillows overlying stock work and sills at depth. Sediment section same as 504A. Many logs have been run in the hole, oblique and VSP seismic experiments have been run. The hole is open, but there is a diamond drill bit in the bottom of the hole.

534A Leg 76, 28° 20.6'N, 75° 22.9'W (Atlantic, E. of Blake Escarpment). Water depth 4976 m, penetration 1666.5 m, 27.5 m basement. Gamma, caliper, sonic, density and temperature logs run. Hole appears to be in good condition.

547B Leg 79, 33° 46.8'N, 9° 21.0'W (off Morocco). Water depth 3952 m, penetration 1030 m. 1030 m to 924 m mudstone, 924 m to 422 m Cretaceous-Jurassic mudstone and limestone, 422 m to 0 m Tertiary ooze. Hole is probably in good condition.

553A Leg 81, 56° 05.32'N, 23° 20.61'W (Rockall Bank). Water depth 2339 m, penetration 682.5 m, basement 500 m bsf. 682.5 to 500 m basalt, 500 m to 230 m Eocene volcanoclastic breccia, tuff, ash, etc., 230 m to 0 m Pleistocene to Eocene ooze. Bottom 27.8 m of hole filled by sediment. Noisy sonic, gamma ray, dual induction, density, caliper and temperature logs run. Hole probably open.

595A Leg 91, 23° 49.34'S, 165° 3.61'W (near Tonga Trench). Water depth 5615 m, total penetration 124 m. Hole is probably open. The instrument for the University of Hawaii seismic experiment is still in the hole.

597C Leg 92, 18° 48.40'S, 129° 46.22'W (Galapagos Spreading Center). Water depth is 4164 m, penetration 143 m, 91 m into basalt. Caliper and two televiewer logs. Hole is apparently in good condition.

638C Leg 103, 42° 09.2'N, 12° 11.8'W (Galicia Margin). Water depth 4673 m, depth of hole 547 m. Hole conditions are apparently bad with sand in the hole. The hole was filled with mud before leaving.

642E Leg 104, 67° 13.2'N, 2° 55.8'W (Voring Plateau, Norwegian Sea). Water depth 1289, hole depth 1229 m. The tope of the cone is at the sea floor.

645E Leg 105, 70° 27.4'N, 64° 39.3'W (Baffin Bay). Water depth 2018 m, hole depth 1147 m. No surface casing, condition of hole is questionable.

648B Legs 106 and 109, 22° 55.3'N, 44° 56.8'W (on the axis of the Mid-Atlantic Ridge). Water depth 3341, hole depth is not known, probably about 45 m. A guidebase was installed at this hole. Last action in the hole was to detonate a charge at about 50 m sub-bottom in an attempt to free the BHA.

UTILIZATION OF BOREHOLE TELEVIEWER LOGGING FOR STRESS DETERMINATION AND BOREHOLE STABILITY ANALYSIS

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We propose a systematic program of logging with a borehole televiewer (BHTV) the holes to be drilled or reoccupied for emplacement of downhole sensors to create permanent ocean bottom observatories. The BHTV will be used to (l) determine the direction and bound the magnitudes of the principal horizontal stresses, (2) determine conditions in the hole affecting sensor emplacement, and (3) delineate the lithostratigraphy and characterize the fractures intersecting the wellbore. The results will both increase the scientific yield of the ocean bottom observatory program and improve the data recorded by the downhole sensors.

The BHTV is an acoustic well-bore scanning device that produces oriented images of wellbore reflectivity and shape as a function of depth (Figure 1). Digital processing techniques greatly enhance the results that can be obtained from the televiewer data. The presence of pillows and the location and orientation of fractures and cross-bedding can be detected in the processed data (Figure 2). Where mechanical and hydrologic properties are controlled by fracturing or structure, sections of weaker rock or greater permeability can therefore be inferred. Intervals of stress-induced wellbore spalling (break-outs) can be detected in the televiewer images (Figure 3) and analyzed quantitatively to determine the direction and bound the magnitudes of the maximum and minimum horizontal principal stresses (Zoback et al., 1985; Barton et al., 1988). Using the digitization and analysis programs developed at Stanford, breakouts have been detected in numerous wells drilled on land, and also in DSDP Holes 597C and 504B (Newmark et al., 1984; Morin et al., submitted).

Successful operation of sensors emplaced in a borehole depends on the state of the hole and of the surrounding material at the point of emplacement. Thus it is critically important to determine the mechanical properties of the rock and the size and shape of the hole, before the sensors are installed. Analysis of data collected in boreholes often requires the assumption that the rock behaves elastically at the borehole wall, and that the wellbore is cylindrical in cross-section. Neither is true within breakout zones.

The orientation of bedding in sedimentary sections, and the presence of fractures or pillows in oceanic basement, can also influence the behavior of downhole sensors. In particular, such properties as time-dependent strain and elastic-wave particle motion are strongly affected by the mechanical properties of the surrounding rock. Thus geophones or strain gauges must be sited to avoid fractures or heterogeneous units; the data we obtain with the BHTV is critically important to optimize the instrument emplacement.

The information obtained from BHTV data is also of more fundamental interest to seismic observations. The preferred orientations of fractures have been related to seismic anisotropy (Little and Stephen, 1985). Furthermore, plate kinematics affect both the elastic stresses in the brittle crust and the orientation of asthenospheric anisotropy. As both the direction of applied stress and the orientation of fractures intersecting a well can be determined from BHTV data, these effects can be studied through use of the televiewer in the holes drilled and/or re-occupied for downhole sensor emplacement.

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ELECTROMAGNETIC INVESTIGATIONS OF DEEP EARTH CONDUCTIVITY

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Much like seismic velocity and attenuation, the electrical conductivity of the Earth is an indicator of subsurface structure and composition. At shallow crustal depths (T<600°C) conductivity is primarily related to conduction in fluid-filled pores and cracks, while at mantle depths (T>1000°C), conductivity is dominated by conduction along connected paths of partial melt along grain boundaries. This mantle conduction process is dependent, to varying degrees, on temperature, pressure, composition, volatile content, etc. As such, knowledge of the spatial distribution of conductivity is highly complimentary to the effort to constrain subsurface structure, composition and dynamics using measurements of seismic parameters. Furthermore, the experimental procedures for collecting passive electromagnetic data useful for bounding mantle conductivity are similar in many respects to the corresponding procedures used for collecting seismic data. It is therefore possible, in principal to simultaneously undertake seismic and electromagnetic observations at a single site.

Overview of Signal Sources

For deep Earth studies, electromagnetic observations may be classified on the basis of the signal source. The geomagnetic field is dominated by signals of purely internal origin at periods longer than one or two years (the Secular Variation). At periods of about 0.25 days to one or two years, the equatorial magnetospheric partial ring current (3-4 r\e?) is the dominant signal source at most latitudes. At higher frequencies, the complex system of ionospheric and magnetospheric electrical currents, as well as the velocity field of the world oceans, all play a role in generating electromagnetic source signals.

The depth of penetration of magnetic field oscillations into the Earth is dependent both on the conductivity structure, and on the frequency, with greater penetration at lower frequencies. Because of this effect, as well as the shielding effects of the abrupt jump in conductivity observed globally at midmantle depths, it is necessary to make use of the purely internal geomagnetic field variations observed at very long periods for lower mantle and outer core studies.

Deep Mantle Conductivity and Outer Core Velocity

This work involves estimation of the attenuation of the internal secular variation by the conductive lower mantle; the use of the observed regional accelerations in the secular variation as a source of conductivity information;modelling the geodynamo process, as well as constraining certain linear functionals of the outer core velocity field by making use of surface observations of the internal geomagnetic components, etc.

Midmantle Conductivity and Constraints on Partial Melt, Temperature, etc.

At depths roughly coincident with the 670 km discontinuity, and typically spanning the range of 300-1200 km, Geomagnetic Depth Sounding may be used to map both lateral heterogeneities in conductivity structure, as well as the vertical gradient in conductivity on a

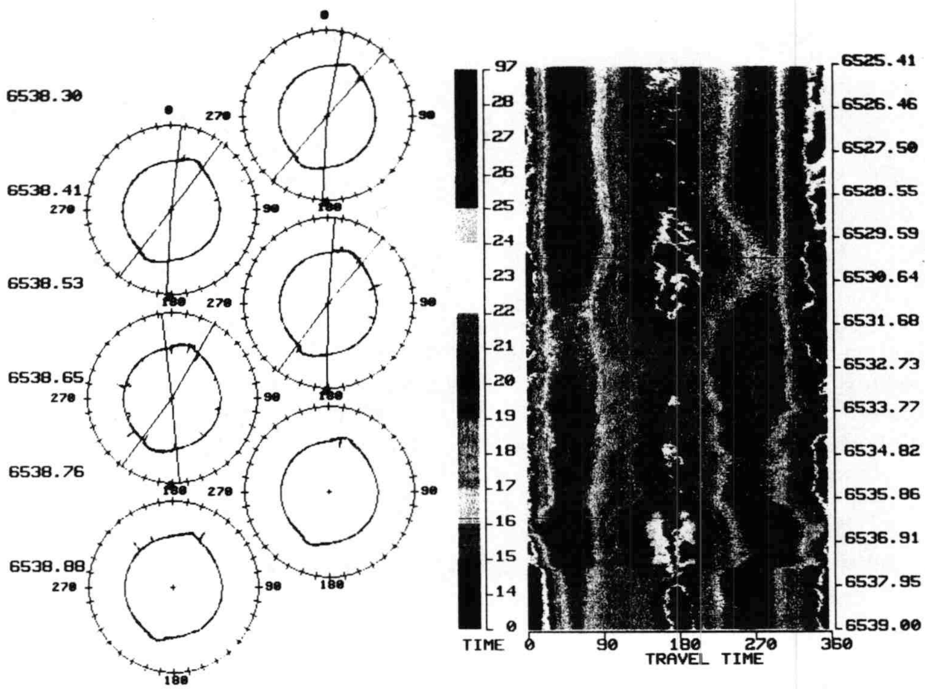

Figure 3: BHTV travel-time data from the Cajon Pass Research Well showing images of breakouts and breakout cross-sections from digital borehole televiewer data.

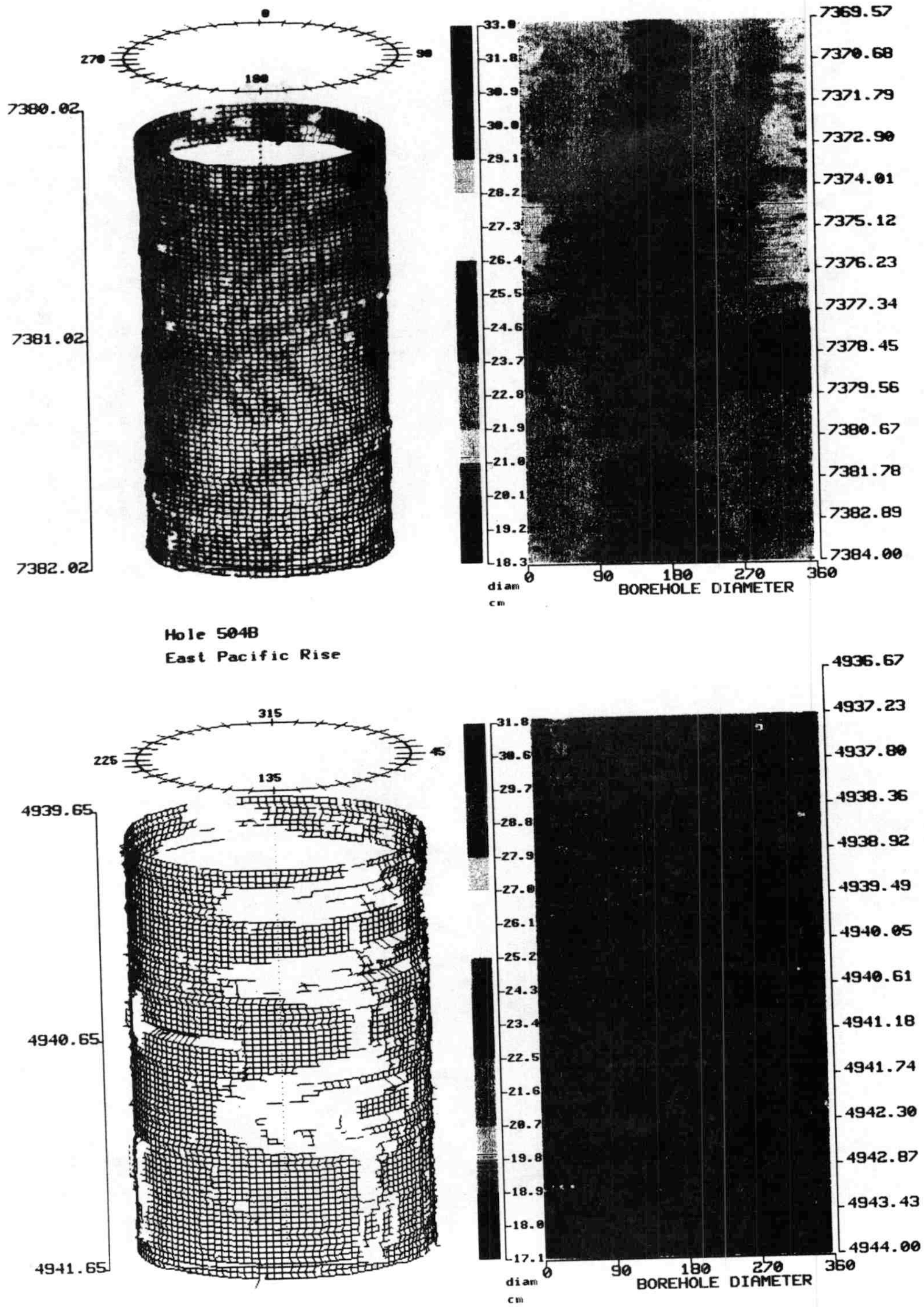

Figure 2: Examples of data showing (a) cross-bedding and fracturing in sandstones, and (b) pillow basalts in DSDP Hole 504B from digitally enhanced BHTV data. The cylindrical projections are wire frame images of the inside of the borehole, and are plotted at approximately one-to-one scale.

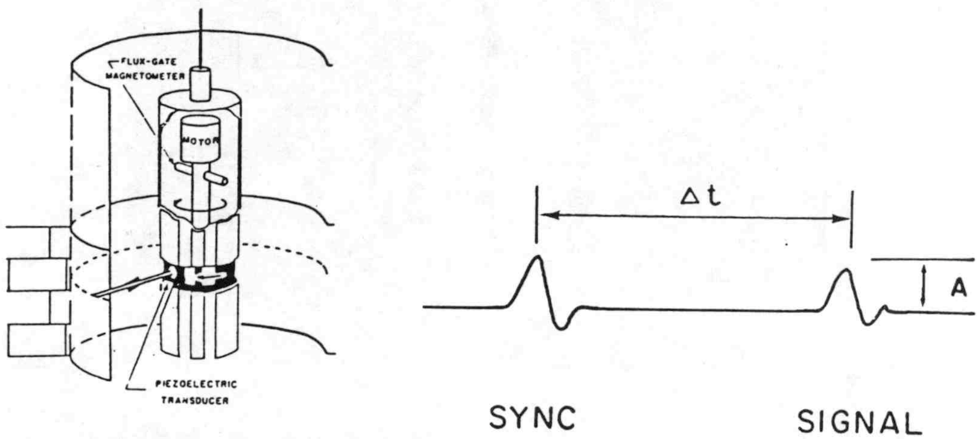

Figure 1: Schematic of the borehole televiewer, showing the principal data extracted digitally from each acoustic pulse. The transducer rotates 3 times per second and fires 1800 times per second; the travel-time and amplitude of each reflected pulse is calculated from the data digitized at 1 microsecond intervals, and displayed as a function of position and azimuth w.r.t. magnetic N.

global scale. This makes use of the magnetospheric partial ring current source field, and involves estimating the internal response of the Earth to this external excitation. In one formulation, the geomagnetic field is expanded in terms of a spherical harmonic representation. The ratio of internal and external parts of the field (as a function of frequency) may be determined for each degree and order of the expansion. These response functions provide information about subsurface electromagnetic impedance, and may be used for estimating conductivity by either inversion or forward modelling.

At present, the global distribution of observatories where there are reliable continuous records of vector magnetic variations in the period range 0.25 days to one or two years is extremely limited. The distribution resembles the WWSSN, and is strongly biased in favor of northern hemisphere continental sites. There are a number of island observatories (typically of poor quality), but long period seafloor observatories have not yet been put into operation. This severe spatial aliasing seriously limits the degree and order of the field expansion.

Reliable internal/external response functions have been obtained only for the P_{-1}^{0} term - limiting classical GDS studies to radially symmetric conductivity models. This restriction to global lD models has been eased somewhat by the recent development of perturbation techniques permitting moderate levels of lateral heterogeneity - but the poor distribution of observatories remains a significant hurdle to improving the resolution of these heterogeneities. This grows increasingly important since recent work indicates that consistent broad scale regional upper mantle heterogeneities may be seen on a global basis. Work is also underway on constraining the abruptness of the conductivity gradient in the midmantle. A sharp orders-of-magnitude increase in conductivity is globally observed at depth ranges of 400-800 km, but the slope and depth of this gradient varies from site to site. Much is therefore to be gained from filling in the gaps in our global magnetic field coverage by instrumenting the seafloor with multi-year deployments of magnetometers.

Upper Mantle Structure and Partial Melt Mapping

At depths of perhaps 30-300 km, magnetotelluric investigations may be used to obtain both relatively high resolution conductivity information (tens of km), as well as information on anisotropy. MT involves the simultaneous measurements of natural magnetic and electric field variations. On the seafloor, MT is band-limited by the presence of the ocean. The conductive seawater shields the seafloor from external magnetic field variations of periods shorter than a few minutes. At periods much longer than one day, induction due to the dynamo interaction of the oceanic velocity field dominates the spectrum. At extremely long periods (in the range of global GDS studies) it is unknown if the oceanic induced signal continues to dominate the seafloor EM spectrum. In any event, seafloor MT is a valuable tool for upper mantle studies at one range of frequencies, and an equally effective tool for physical oceanographic studies at a lower range of frequencies.

The possibility exists that at extremely long periods, seafloor MT may play an important role in midmantle studies. Recent work in very long period MT combined with a GSA study at Tucson has revealed the presence of a well localized peak in conductivity at about 400 km depth, superimposed on the background global midmantle conductivity gradient. This increase in resolution over previous studies was made possible by the addition of very long period electric field data. The collection of seafloor MT data, as well as GDS data, promises to provide global coverage of mantle conductivity structure with unprecedented resolution. The correlation of mantle conductivity anomalies with seismic parameters such as slowness is an exciting possibility that needs to be pursued within the framework of any new initiative in global geophysical instrumentation.

Also note that while slowness may vary by a few %, conductivity may vary by many orders of magnitude (as much as 7 decades) - so it is quite a sensitive indicator of changes in bulk averaged properties.

STRAINMETERS FOR DEPLOYMENT IN DEEPSEA DRILL HOLES

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Borehole strainmeters are a relatively new class of instruments, which have been used to measure seismic and tectonic strains at world-wide locations since 1971. Several kinds of instruments currently are in use. Originally proposed by Benioff (1935), the first practical borehole strainmeter was the Sacks-Evertson dilatometer (Sacks et al., 1971, Evertson, 1977). In the dilatometer pressure changes in a cylindrical volume of silicone oil (Fig. 1) are related to volume strain the host rock. More recently, borehole strainmeters that are capable of resolving more than one component of the strain tensor have been devised (Moore et al., 1974; Agnew, 1986). The Sakata instrument (Sakata, 1981; Shimada et al., 1987), for example, is a hydraulic design that has three separate sensing volumes. By intercomparison of the pressures in the three volumes, the horizontal components of the normal and shear strains can be determined. The Gladwin instrument (Gladwin, 1984; Gladwin and Hart, 1985) also measures the horizontal normal and shear strain components, but unlike the Sacks-Evertson and Sakata designs it uses diametrically opposed displacement transducers at three azimuths with the instrument case (Fig. 2).

Borehole strainmeters are broadband instruments, capable of producing usable data from nearly 0 Hz to seismic frequencies. At low frequencies the instruments respond to secular changes in the strain field due to tectonic processes (McGarr et al., 1982), to tidal strains (King et al., 1976), and to the most grave free oscillation modes. In the Sacks-Evertson dilatometer the useful high frequency response is limited to about 1 Hz by the design of the pressure transducer. The Gladwin instrument is probably capable of significantly higher frequencies than the hydraulic instruments, because it uses a different type of transducer. The high frequency limits of the instrument have not been fully explored, however, as many of the existing installations are for the purpose of earthquake prediction research where stable performance at long time scales is emphasized. A comparison is presented in Figure 3 for teleseismic signals in the frequency range 0.04-0.01 Hz that were recorded using a Sacks-Evertson dilatometer and an accelerometer at the same station. It is evident that the strain records are in many days similar to the acceleration records, but close examination of the reveals numerous differences in the amplitudes and phases of individual waveforms.

The phase should differ by $\pi/2$ between the acceleration record and the strain record. A single, harmonic component of a plane wave has the form $s = A \sin(kx - \omega t + \phi)$. The strainmeter responds to the first space derivative, $s' = kA \cos(kx - \omega t + \phi) = kA \sin(kx - \omega t + \phi - \frac{\pi}{2})$, while the accelerometer responds to the second time derivative $s = \omega^2 A \sin(kx - \omega t + \phi)$. This difference of $\pi/2$ in phase can be seen for the surface wave in Figure 3, where the signal-to-noise ratio is greatest. The amplitude relationships are more complicated, as the strain amplitudes are related to the acceleration amplitudes by the phase velocity $c=\omega/k$. For dispersed waves, an intercomparison of the spectral amplitudes for the acceleration and strain records provides a basis for determining the phase velocity dispersion curve (Mikumo and Aki, 1964; Sacks et al., 1976). The interpretation of such a dispersion curve in terms of earth structure applies within a relatively small neighborhood (dimensions on the order of one wavelength) of the observatory, so that significantly better lateral resolution may be achieved than if an array of widely spaced seismographs is used to make phase or group velocity measurements. In general dilatometers respond to P-waves, Rayleigh waves, and other modes of propagation that have Rayleigh eigenvectors. It is also common to observe S-waves on dilatometer records, since the instruments are installed near a boundary where S-to-P conversion occurs.

The apparent body wave amplitude on a strain record is a more complicated issue than surface wave amplitude. In the first place a disturbance that can be viewed as a plane wave in the host rock probably does not appear to the strainmeter as a plane wave, because the wave motion near the strainmeter is perturbed by the presence of the strainmeter itself. The dimensions of the region in which the wave is perturbed are rather comparable to the diameter and length of the strainmeter itself than to the seismic wavelength. In the second place, a body wave traveling horizontally, perpendicular to the long axis of the borehole, will produce a different response from the strainmeter than it would traveling vertically, parallel to the borehole. In general, the response of a borehole strainmeter to body waves is strongly dependent upon the angle of emergence, or the angle between the axis of the instrument and the direction of wave. It is possible to establish calibration curves for strainmeter response versus angle of emergence, which are specific to a particular station, by finite element modeling of the strainmeter and host rock as a mechanical system (Williams, 1986).

Strainmeters generally supplement, not replace, the more traditional pendulum instruments. Although either kind of instrument can be used for certain studies, measurements of group velocity dispersion for example, having strainmeters and pendulum instruments together in the same observatory would produce the maximum amount of information. A possible configuration for a marine borehole seismic observatory would be a single instrument package, which contains individual sensor modules that measure:

* three components of velocity
* four strain components (dilatation, horizontal normal and shear strains)
* two tilt components
* temperature

Both pendulum seismometers and strainmeters have a history of successful operation in continental boreholes, and much of the existing technology can be applied to strainmeters and tiltmeters for use in marine boreholes. However, the borehole strainmeters currently in operation are all at relatively shallow depths, on the order of 100 meters or less, and a significant development effort would be required to produce the sensor package for a marine borehole.

A preliminary estimate of the size and of the down-hole instrument package can be made using the results from a previous workshop "Science Opportunities Created by Wireline Re-entry of Deepsea Boreholes." Based on the report from that workshop (Langseth and Spiess, 1987), a likely size is a stainless steel cylinder, of 20 cm diameter and 7 m length, and weighing about 350 kg. The downhole package would contain the transducers and a minimum of electronics. A larger electronics package, including data recording or telemetry equipment and a power source, would connect via cables to the downhole package.

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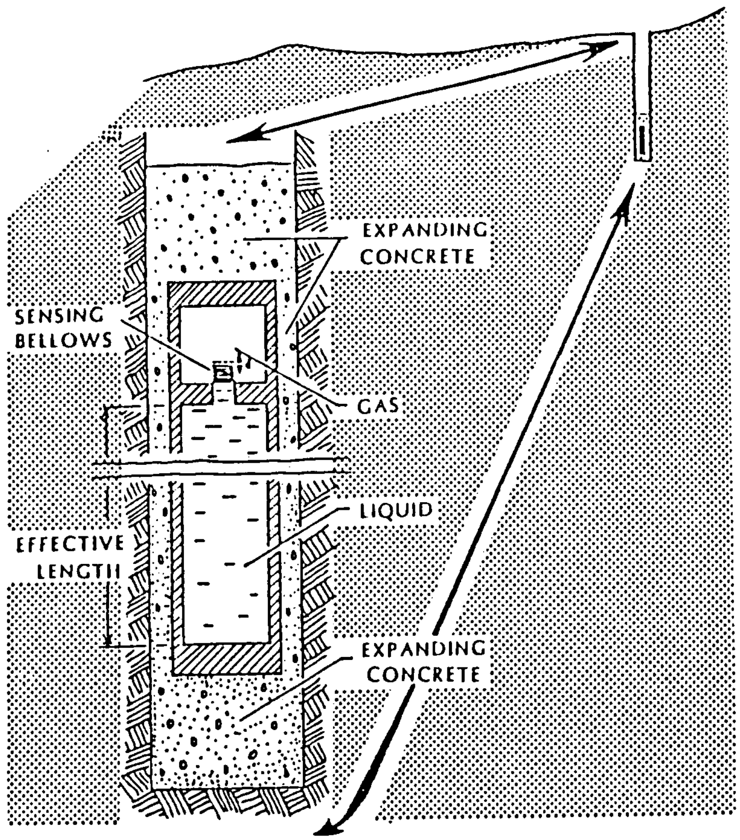

Fig 1. Schematic diagram (no scale implied) of a Sacks-Evertson dilatometer installed in a borehole.

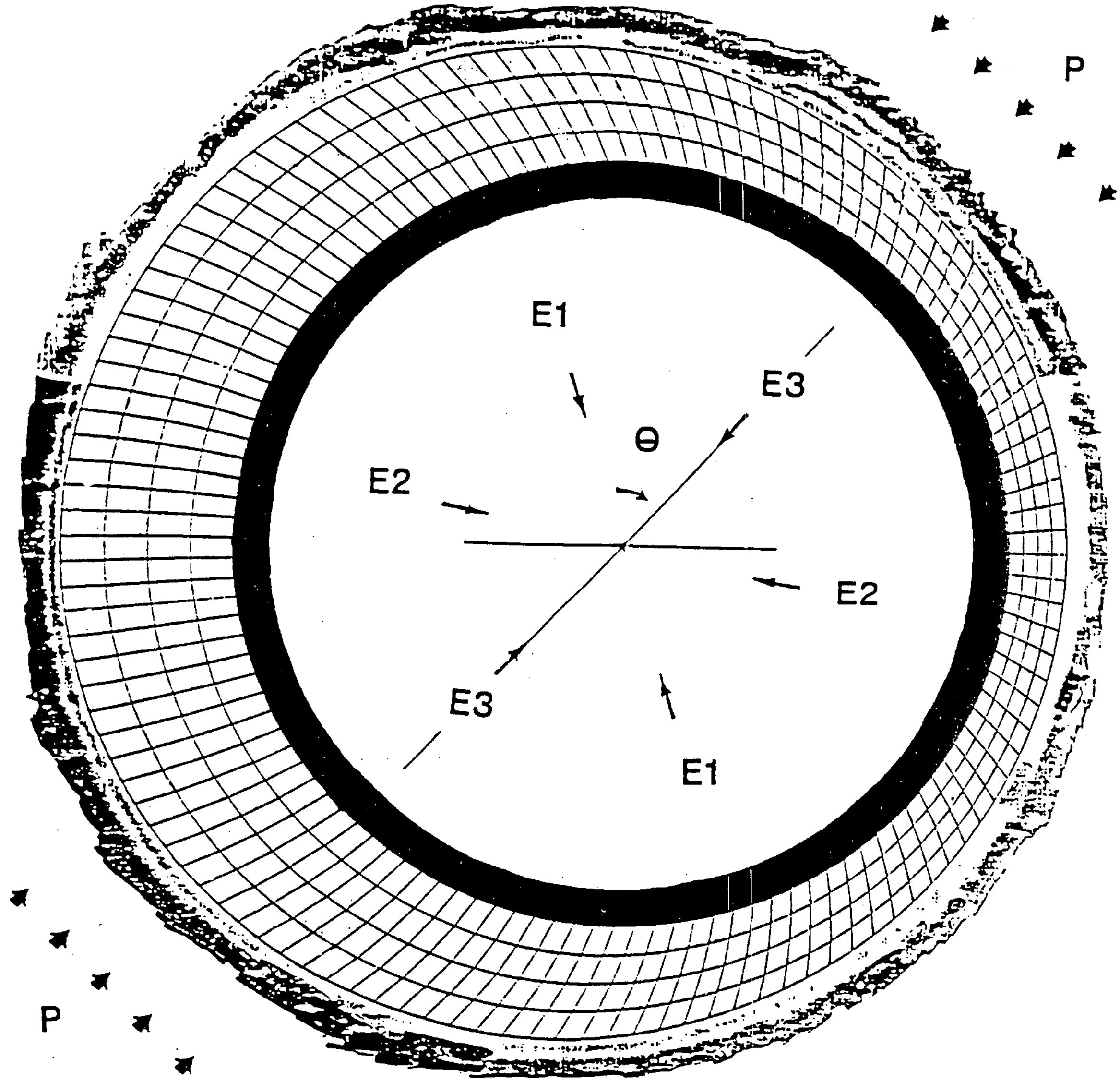

Fig 2. Gladwin vector strainmeter, shown to scale in plan view. The steel instrument housing is depicted by the black ring. It is coupled to the well bore by expanding concrete (grid). From measurements of the well bore elongation in three directions (E1, E2 & E3) the direction (θ) and magnitude (P) of the horizontal stress components can be determined. The diameter of the well bore as shown is 190 mm, and the instrument is offset from the center of the well by 10 mm.

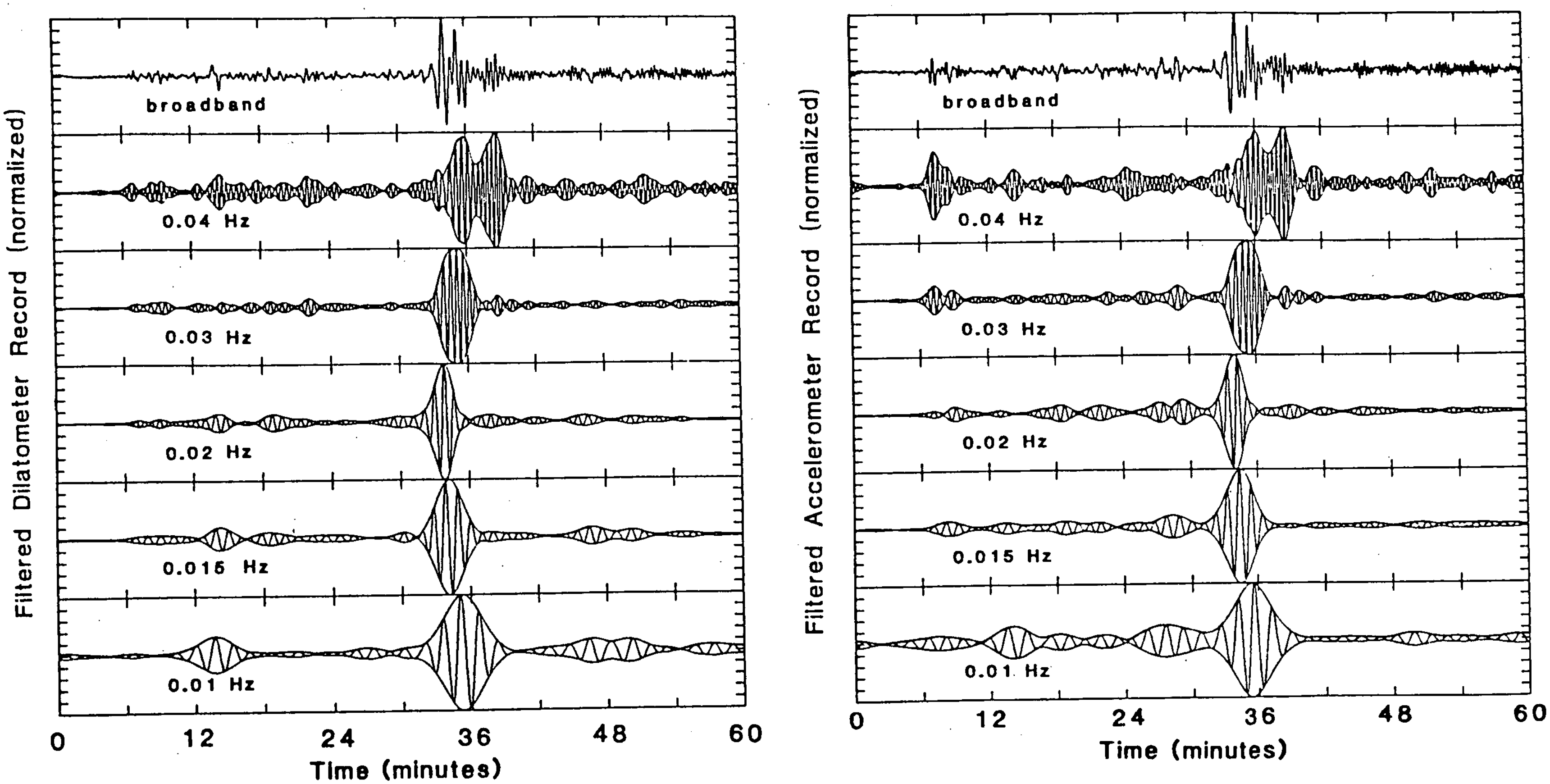

Fig 3. Broadband and narrow-band filtered dilatometer (left) and accelerometer (right) records of the 15 November 1984 Loyalty Islands earthquake (M_b=6.2). The instruments were installed at Washington, DC. The narrow-band data were filtered using constant Q Gaussian filters, Q=1.5. Either data set could be used to measure the group velocity dispersion, but comparison of the filtered Rayleigh wave data reveals an expected phase difference of $\pi/2$ between the two instruments (see text).

Ocean Floor Seismic (OFS) OBSERVATIONS FOR NUCLEAR DISCRIMINATION RESEARCH

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I. INTRODUCTION

Ocean bottom seismic stations could play an essential role in nuclear discrimination research. This stems from the likelihood that the number and location of ocean bottom stations, located in international waters, will not be subject to regulation. Significantly, the world's most tectonically active areas, where discrimination is most difficult, are near the deep ocean. These areas include the entire Pacific margin (i.e., "the Ring of Fire"), the Alpine-Himalayan tectonic belt extending from the Mediterranean through the Middle East to Indonesia and, of course, the oceanic ridges. In fact, nearly all of this entire area is within 5° of international waters. Even the Himalayan tectonic belt is less than 20° from the ocean. Clearly internal land stations, whose number and location are likely to be restricted by a host nation, are not necessary to study these high seismicity areas. In any event, available land station should not be wasted here. They would be better used to those internal regions whose geologic structure would lend itself to decoupling-type evasion techniques.

The use of ocean bottom stations for regional discrimination research is predicated, of course, on the ability of such stations to detect seismic events as effectively as alternative land stations. Unfortunately, adequate information is not available to assess the relative performance of ocean bottom stations in a near-in, or regional context. Previous experiments sponsored by ARPA in the early 1960's indicated that ocean bottom stations would not be as effective as land stations for global teleseismic surveillance. This judgment was based on the observation that the background noise level, measured at several sites by short period seismometers, was much higher than at land stations. The instruments were deployed by simply dropping them onto the seafloor mud and ooze. Although these instruments were not able to examine the long period noise which earlier workers suggested was comparable to land observations, extrapolation of the short period results suggested that long period noise levels would also be higher than land station levels. Accordingly, the concept of using ocean bottom stations for teleseismic monitoring was abandoned in the late 1960's.

Today such a judgment may no longer be valid in view of recent comparisons of short period S/N ratios of OBS and borehole instruments. It appears that modern digital seismographs rigidly emplaced beneath the seafloor by deep sea drilling ships, manned submersibles, or remote controlled manipulator ships can provide inexpensive ocean bottom stations useful for both long period teleseismic and short period regional monitoring. In fact, by incorporating modern acoustic satellite and seafloor cable data telemetering techniques, it should be possible to have essentially real-time monitoring comparable to land stations.

This report summarizes the role that ocean floor observatories might play in modern discrimination research. Below, we discuss the advantages to using ocean floor stations that have come about from recent advances in ocean technology.

II. ADVANTAGES OF OCEAN BOTTOM STATIONS (OBS's)

A. Ocean floor Observations vs Land Stations

Aside from their obvious political desirability, ocean floor seismic observations may have important scientific and technical advantages over land stations for nuclear discrimination research. Some of these are:

1. Increased Signal Amplitude

Although most previous experiments have probably not provided a faithful portrayal of true ground motion on the seafloor because of poor coupling and high noise, it is clear that signal amplitudes from earthquakes observed on the ocean floor are generally higher than those seen by a land station. For example, the ARPA-sponsored field tests off the Aleutian Islands in 1968 showed that OBS-calculated m_b values averaged 0.2 unit greater than land station calculated values. This was thought to result from the fact that the rays travelled a slightly shorter path to the ocean bottom stations and, more important, they did not have to propagate through a low Q continental crust. Also, more recent work has shown attenuation along oceanic lithosphere paths to be extremely small. Q values are estimated to exceed 6000. For the higher frequency band which will be particularly useful in regional monitoring, this signal enhancement could be significant.

2. Lower Noise

Deep ocean sites far from land would be relatively isolated from the sources of background noise likely to affect seismic stations. These are: cultural noise, breaking surf on coastlines, storm microseisms, and local sea surface waves or tidal current effects. In practice this appears to be the case. Both the early work of Lamont and the later ARPA-sponsored Texas Instruments field tests showed decreased noise with increasing water depth and increasing distance from land. The dominant noise source is believe to be surf microseisms propagating out from the coastline as Rayleigh waves in the water mass and along the water-seafloor interface. The latter path is particularly energetic.

It must be emphasized at this point that, after considering the field method used to make background noise observations, the apparent high noise levels may not have been a true measure of solid-earth motions. In the T.I. system and for that matter with most systems: (i) the seismometers simply rested on surficial, unconsolidated seafloor sediments and ooze whose seismo-acoustic properties were not much different than the overlying water mass; OBS position 1), and (ii) most systems used tall vertical frames which protruded into the water mass to house their seismometers. These facts suggest that such devices probably recorded ocean water as well as solid-earth ground motion. Significantly, those seafloor instruments with seismometers well coupled to the solid earth and isolated from the "wind-like" fluid motions induced on the housing frame by ocean currents and turbulence as well as the tilt of soft sediment have shown short period noise levels approaching the 1-10 mμ levels observed at the better land SRO stations. These include the early Lamont devices of the 1950's (=1mμ) and the more recent Japanese and British OBS's (=25-50mμ). Notably, most OBS devices have been plagued by much higher noise levels. In fact, some show strong frame-sediment resonance.

Unfortunately, no recent studies of long period seafloor noise have been made. In fact, only the two long period OBS's developed at Lamont in the early 1960's have had significant recording durations. One of these, a shore cable connected system off California, operated from the mid-1960's until the mid-1970's. Unfortunately no definitive analysis of noise observations from this device has been reported. However, information from a short duration recording (8.5 days) off Bermuda show noise amplitudes about 2 orders of magnitude greater than today's typical land SRO stations (i.e., 5 μ vs 50 mμ). Again, these measurements are subject to the same doubts expressed about the short period observations in that these devices were essentially free-fall instruments resting on the sediment surface of the seafloor.

In summary it appears that the better free-fall OBS's existing today have short period noise levels approaching land stations. These low noise levels combined with the expected higher signal amplitude on the ocean floor suggest that even these simple seafloor-type, pop-up OBS instruments have a S/N ratio useful for regional discrimination research. With their improved coupling of borehole seismometers to the solid earth and their isolation from water motions and resonance effects, it is probable that ocean-floor borehole seismic observatories may have long and short period S/N ratios higher than island and OBS stations could be developed for both teleseismic and regional discrimination research.

3. More Uniform Crust and Mantle Structure

The crust and mantle structure beneath the ocean basins is now known to be much simpler than that beneath the continents. The seafloor spreading hypothesis, generally accepted by ocean scientists to account for the formation of the seafloor, predicts that nearly horizontal subplanar rock layers underlie most of the deep ocean. Only at mid-ocean ridges, deep trenches, and oceanic islands will there be significant lateral inhomogeneity of earth structure. Deep sea drilling, seismic refraction, and gravity and magnetic measurements support this hypothesis.

Such simple layering implies that large aperture arrays could be deployed to further improve the S/N ratio by beamforming or velocity filtering. In fact, many of the signal processing problems caused by the near-field complexity of earth structure at land large aperture arrays (e.g., LASA, NORSAR) should not be encountered. By sharply reducing signal-generated coherent noise, a closer approach to the ideal $\sqrt{N}$ signal to noise ratio improvement might be realized. The widespread uniformity of earth structure beneath the deep ocean basins also implies that much larger arrays than those practical for land installation could be built.

4. Simplified Operation

The operational advantage of remote OBS stations may be significant since no on-site personnel are involved. Also, by employing acoustic and buoy/satellite data telemetering techniques, ship costs for data retrieval or costs for interconnecting cables or satellite telemetry necessary for real-time data links are eliminated. In fact, deep sound channel (SOFAR) hydro-phones like the Air Force's Missile Impact Locating Systems (MILS) could provide a reliable quasi-real time monitoring capability on a global scale. Of course, moored satellite telemetering buoys with only local short range acoustic links to the ocean floor observatories or a completely hard-wired cable system could be used. Unfortunately, these real-time monitoring approaches would probably made ocean floor observatories cost more comparable to land internal stations.

5. Greater Geographic Coverage

Since oceans occupy 70% of the earth's surface and many areas have no islands suitable for seismic observatories, it is difficult to infer the seismic structure beneath the ocean basin areas using only land stations. In any event, observations made from island stations are likely to be unrepresentative of the broad ocean crust and mantle structure. A well distributed network of ocean bottom seismographs and/or large aperture arrays would fill this gap in our knowledge. Also, the low level seismicity of such important features as trenches and mid-oceanic ridges which are only accessible with ocean floor observatories can be examined. These observations could have important ramifications for general earthquake research as well as for nuclear discrimination research.

B. Borehole Seismometers vs Seafloor OBS's

The most effective method to improve both the signal noise ratio and discrimination performance of ocean floor seismic observatories is to rigidly mount the seismometer beneath the soft sediment layer.

1. Better Coupling to the Solid Earth

Competent semi-consolidated sedimentary materials are generally found a few tens of meters beneath the unconsolidated surficial seafloor sediments. These deeper, lithified layers show sharp increases in both compressional and shear wave velocities and bulk density. In face, the hard, crystalline igneous rocks of the High Q oceanic crust are usually covered by less than a few hundred meters of sedimentary materials in most areas. Accordingly, a borehole-type seismometer installation, much like the present SRO stations, which is employed in the hard sediment or on/within the oceanic basement rocks should provide signal coupling vastly superior to free-fall devices resting on the soft sediment/water interface. Also, the signal amplitude can be maximized by simply adjusting the overall seismometer case density to match the acoustic impedance of the surrounding borehole rock materials.

2. Lower Noise

The depth of burial necessary for the seismometer to attain a significant noise reduction is probably only a few tens of meters due to the sharp gradient in the seismo-acoustic properties of the soft seafloor sediments. A buried seismometer is not only isolated from the wind-like noise induced by the ocean currents and turbulence on the housing frame but the soft overlying surficial material may act much like a soundproofing layer which will absorb ocean-generated noise. In addition, the air-water and seafloor-water interfaces will serve as efficient reflectors which effectively trap any propagating waves within the water volume. Tests conducted by Woods Hole/Miami scientists in shallow water showed more than a factor of ten reduction in local wind/current-generated noise on a vertical component, short period seismometer buried only a few meters beneath the seafloor. Notably, some of the early Lamont OBS's which reported very low noise levels had their seismometers in probe-like legs which penetrated the seafloor a few meters.

The coherent microseism noise propagating as Rayleigh waves in the ocean water and along the seafloor sediment/water interface is also markedly attenuated with increasing depth since the shear wave velocity of the surficial sediments is only about 0.2 km/sec. Thus, the microseism noise which is generated beneath breading surf on distant beach surfaces does not penetrate very deeply into the soft sediment layer. Seismometers buried 300-600 meters beneath the seafloor would be virtually shielded from this dominant source of seafloor noise.

3. Reduced Signal Contamination

Seismic signals received at a seafloor-type, pop-up OBS travel through the ocean water as well as through the solid earth beneath the station. Those rays arriving at the OBS which are reflected from the local air-sea surface interface, particularly, contaminate the direct seismic arrival phases. This signal-generated noise not only introduces apparent complexity and reverberation in the wave train coda but it also tends to generally reduce signal amplitude because of interference. Accordingly, by locating the seismometer in a borehole beneath the seafloor surface, seismic signals entering the overlying water mass and soft sediments will be effectively trapped much like noise initially generated in the ocean. In fact, seismometers near the oceanic basement-hard sediment interface should be virtually free of reflected arrivals returning from overlying interfaces.

APPENDIX 1

WORKSHOP OUTLINE

DOWNHOLE SEISMOMETERS IN THE DEEP OCEAN

Clark Building, 5th Floor, Quissett Laboratories
Woods Hole Oceanographic Institution

<u>TUESDAY</u> - 26th April

8:00 a.m. - 8:45 a.m.	Registration, coffee and pastry
8:45 a.m. - 9:00 a.m.	Introduction to Workshop - Purdy and Dziewonski
9:00 a.m. - 12:00 p.m.	PLENARY SESSION 1: SCIENTIFIC REQUIREMENTS Moderator: Dziewoński
9:00 - 9:30 a.m.	Ocean Floor Stations and Resolution of 3D Structure for the Whole Earth: Woodhouse
9:30 - 10:00 a.m.	Ocean Floor Stations and Source Mechanism Studies: Kanamori
10:00 - 10:30 a.m.	Ocean Floor Stations and the Measurement of Anisotropy of the Oceanic Lithosphere: Tanimoto
10:30 - 11:00 a.m.	Coffee
11:00 - 11:30 a.m.	Ocean Floor Stations on the Measurement of Regional Surface Wave Dispersion and Attenuation: Mitchell
11:30 - 12:00 p.m.	Body Wave Studies Using Ocean Floor Data: Solomon, Orcutt and Jordan
12:00 - 12:30 p.m.	Application of Ocean Floor Stations to Tsunami Prediction: Burton
12:30 p.m. - 2:00 p.m.	Lunch
2:00 p.m. - 6:00 p.m.	PLENARY SESSION 2: REVIEWS OF EXISTING AND PLANNED PROJECTS Moderator: Purdy
2:00 - 2:30 p.m.	FDSN Organization and Current and Future Plans: Berry Station Siting: Engdahl
2:30 - 2:50 p.m.	IRIS: Butler
2:50 - 3:10 p.m.	GEOSCOPE: Romanowicz

3:10 - 3:30 p.m.	POSEIDON: Hamano
3:30 - 3:45 p.m.	RIDGE: Purdy
3:45 - 4:15 p.m.	Coffee
4:15 - 4:40 p.m.	Opportunities for use of TPC 1: Nagumo
4:40 - 5:00 p.m.	The Ocean Drilling Project: Von Herzen
5:00 - 5:20 p.m.	Downhole Seismic Measurements in the Deep Ocean: Past, Present and Future: I Duennebier and McCreery
5:20 - 5:40 p.m.	Downhole Seismic Measurements in the Deep Ocean: Past, Present and Future: II Stephen and Orcutt
5:40 - 6:00 p.m.	Long Period Measurements on the Ocean Floor: Sutton and Dorman
6:00 - 6:30 p.m.	Refreshments
6:30 p.m.	Buffet Dinner and Poster Session

WEDNESDAY - 27th April

7:30 a.m. - 8:00 a.m.	Coffee and Pastry
8:00 a.m. - 12:00 p.m.	PLENARY SESSION 3: THE TECHNICAL ISSUES Moderators: Dziewonski and Purdy
8:00 - 8:20 a.m.	U.S. National Broad Band Network: Masse
8:20 - 8:40 a.m.	Broad Band Sensors: Wielandt
8:40 - 9:00 a.m.	GSN Data Acquisition System: Dziewonski
9:00 - 9:20 a.m.	PASSCAL Data Acquisition: Phinney
9:20 - 9:40 a.m.	Continental Downhole Installations: Hutt
9:40 - 10:00 a.m.	Ambient noise measurements at the new IRIS/IDA oceanic island sites and some preliminary results from Guralp borehole sensors: Berger
10:00 - 10:20 a.m.	Coffee
10:20 - 10:40 a.m.	LFASE Downhole Seismometer Array: Stephen
10:40 - 11:00 a.m.	Satellite Telemetry in Oceanography: Frye and Briscoe
11:00 - 11:20 a.m.	A Geophysical Downhole Observatory: Williams

11:20 - 11:40 a.m.	Utilization of Televiewer logging for stress determinations and borehole stability analysis: Zoback and Moos
11:40 - 12:30 p.m.	Formation of Subgroups and Definition of their Tasks
12:30 p.m. - 2:00 p.m.	Lunch
2:00 p.m. - 6:00 p.m.	Subgroup 1: Scientific Objectives: Forsyth
	Subgroup 2: Experiment Designs and Technical Issues: Orcutt
6:00 - 6:30 p.m.	Refreshments - FENNO HOUSE
6:30 p.m.	Buffet Dinner - FENNO HOUSE

<u>THURSDAY</u>- 28th April

8:00 a.m. - 8:30 a.m.	Coffee and Pastry
8:30 a.m. - 9:30 a.m.	Subgroups meet separately for review.
10:00 a.m. - 12:30 p.m.	PLENARY SESSION 4: PLANS Moderators: Dziewonski and Purdy
10:00 - 10:45 a.m.	Report from Subgroup 1 and Discussion: Forsyth
10:45 - 11:30 a.m.	Report from Subgroup 2 and Discussion: Orcutt
11:30 - 12:30 p.m.	General Discussion and Integration of Subgroup Findings Moderators: Dziewonski, Purdy, Forsyth, Orcutt

End of Formal Part of Workshop

2:00 p.m. - 6:00 p.m.	Text Generation by Steering Committee and Subgroup Leaders

APPENDIX 2
DOWNHOLE SEISMOMETERS IN THE DEEP OCEAN WORKSHOP
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