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# International Deployment of Accelerometers: A Network for Very Long Period Seismology

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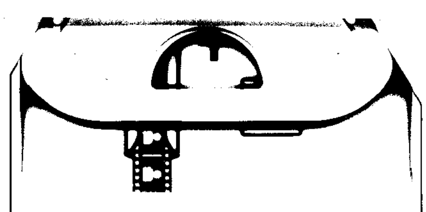
## Introduction

The value to seismology of a global net of seismometers has long been recognized. Indeed, the first proposal for such a net came in 1895, only 6 years after the discovery that earthquake waves could be recorded at long distances, and the first net of standardized instruments was in existence by 1898 [*Dewey and Byerly,* 1969; *Wood,* 1942; *Milne,* 1899]. A true network, involving not only standardized high-quality instruments but also the exchange and ready availability of data, did not come until the establishment of the World-Wide Standard Seismograph Network (WWSSN) in the middle 1960's [*Oliver and Murphy,* 1971]. The importance of this to general seismology cannot be overstated; not only has it improved the quality of traditional research areas, but the availability of the data has suggested new types of investigations. In the decade since the WWSSN was built, two small networks have been set up: the High-Gain Long-Period instruments (HGLP) [*Savino et al.,* 1972] and, very recently, the Seismic Research Observatories (SRO) [*Peterson and Orsini,* 1976]. Both of these networks use careful shaping of the instrument response to improve the detection of surface waves from very small events, which is important in lowering the magnitude threshold for discriminating earthquakes from explosions.

Another development of the past decade has been the improvement of instruments for measuring very long period (VLP) ground motions: those with periods of more than 100 s (frequencies less than 10 mHz). The initial impetus for this came





from physicists who wished to detect the excitation of the earth's free oscillations by gravitational radiation; this resulted in the construction of a fedback instrument by Moore in 1965 [*Moore,* 1966; *Block and Moore,* 1966]. Since then, the design of such instruments has been refined [*Moore and Farrell,* 1970], and the electronics needed for the instrument itself and for digitally recording its output has become much more reliable and inexpensive. The idea of installing a worldwide network of instruments to record VLP phenomena is now practicable. We have begun such a network (called IDA, an acronym for International Deployment of Accelerometers) and have already installed the stations shown in Figure 1.

## Uses of VLP Data

Though gravitational radiation is not now considered likely to be detected by using seismic instruments [*Press and Thorne,* 1972], a network of VLP instruments could still contribute to gravity physics. Certain theories of gravity (though not general relativity) predict anomalous earth tides [*Nordtvedt and Will,* 1972]. Worldwide gravity tide observations might put smaller bounds on the size of such tides than are currently set by single-station observations [*Warburton and Goodkind,* 1976] and so limit the size of possible nonrelativistic effects. Observations of earth tides are interesting geophysically for providing constraints on the behavior of the ocean tides [*Farrell,* 1973]. Measurements at stations outside Europe and North America would be particularly useful in providing constraints on models of ocean tides in such poorly known areas as the southeast Pacific.

From the standpoint of solid earth geophysics the most useful observations to be made at very long periods are those of the free oscillations of the earth that have periods of 1 hour or less. These oscillations are largely unaffected by local structure, so that observations of their frequencies give information about the distribution of elasticity and density with depth, averaged over the whole earth. These oscillations were first observed following the 1960 Chilean earthquake on instruments designed to detect VLP motions, but most of our knowledge of them has come from the processing of data from a great many WWSSN stations [*Mendiguren,* 1973; *Gilbert and Dziewonski,* 1975]. The power of using a network is shown by these investigations, which have increased the number of observed free oscillations from 150 [*Derr,* 1969] to about 1000, despite the poor response and small signal to noise ratio (in this frequency range) of the WWSSN instruments. If quiet instruments with a good response were used, free oscillations can be observed from earthquakes of much smaller size, perhaps as small as magnitude 6.0 [*Block et al.,* 1970] rather than the 7.5 required for WWSSN observations, so that data can be obtained more rapidly. At present, only 20% of the free oscillation multiplets with frequencies less than 20 mHz have been observed and identified. If data from different earthquakes as well as from different stations [*Gilbert and Dziewonski,* 1975] are combined, it should be possible to determine the frequencies of many more multiplets and improve our determinations for those already known. Such data will allow refinements in the models of the elastic structure of the earth.

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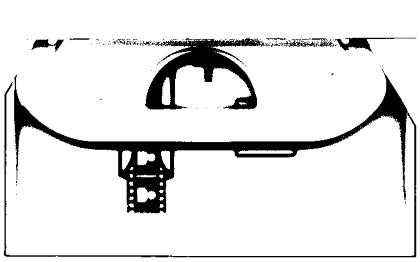
*'No longer need we wait for truly large and very infrequent earthquakes to add to our observational knowledge of the earth's free oscillations.'*

Better data should also stimulate some investigations which have so far not been possible. For example, for an idealized earth model all free oscillations with the same radial and polar displacement functions have the same frequency, while in the real earth the frequency depends slightly on the azimuthal displacement function [*Dahlen,* 1968]. Observations of this 'splitting' can give information on the derivative with respect to depth of the density and elastic parameters; such information would be extremely valuable in constraining models of the earth's structure. Measurement of splitting would also allow accurate determinations of the resonance width ($Q^{-1}$) of free oscillations, which can be used to find the dis-





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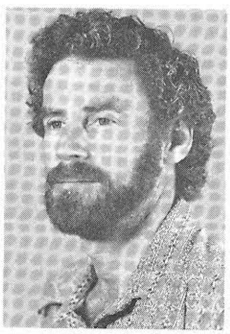


Jonathan Berger was educated at Dalhousie University and the University of California, San Diego. He has been involved for a number of years in the low end of the geophysical spectrum, developing laser interferometers to monitor tectonic motion along the San Andreas fault system and, more recently, developing the instrumentation for Project IDA.

tribution of anelasticity in the earth and also to see how the anelasticity depends on frequency at periods longer than can be investigated by using body waves [*Anderson and Archambeau*, 1964]. Measurements of splitting may also provide information about the degree of lateral heterogeneity in the earth [*Dahlen*, 1974]. A somewhat more speculative possibility is that low-noise instruments might allow observations of so-called 'core modes'; these include translational oscillations of the inner core [*Slichter*, 1961], planetary and gravity waves in the fluid outer core [*Dahlen and Smith*, 1975], and elastic modes in the inner core [*Dziewonski and Gilbert*, 1973]. Observations of these would add greatly to our knowledge of the core. For example, any observations of gravity waves would settle the vexed question of whether or not the outer core is stably stratified (if it is, the impossibility of convection would restrict possible mechanisms for the geomagnetic dynamo). It is not known if core modes are ever excited enough to be observed. Most of them have periods longer than an hour, so that any attempt to detect them will require instruments which are both sensitive and quiet in this period range. Another speculative possibility is that there may be earthquakes which occur very slowly. These would excite only VLP ground motions and so would be detectable only by VLP instruments.

Long-period data can also be used to retrieve earthquake source mechanisms as a function of frequency [*Gilbert and Dziewonski*, 1975]. This technique allows checking of assumptions made in conventional short-period mechanism studies, and knowledge of the focal mechanism allows data from several earthquakes (as well as many stations) to be 'stacked' for the determination of free oscillation frequencies. *Gilbert and Dziewonski* [1975] depended on a dense global network (the WWSSN), but recently, a new technique, matched filtering, has been developed [*Buland and Gilbert*, 1976], which can be used with a sparse global network. In this technique the observed spectrum for a particular record is represented as a linear combination of six synthetic spectra, each one being the unit response for one of the six elements of the symmetric moment tensor of the source. For each frequency band the six elements can be determined by linear regression. Experiments indicated that this method is accurate, reliable, and stable for as few as 10 stations. This technique depends on our ability to compute realistic synthetic seismograms (Figure 2).

Observations with very long period instruments are not confined to free oscillations. Improved instruments would allow the study of very long period surface waves from earthquakes as small as magnitude 5.5. Observations of such surface waves, either as multiple arrivals at a single station or along paths between two stations, may be used to determine the regional structure of the upper mantle and might answer the question of how deep the difference between continents and oceans persists. Determination of earthquake mechanisms from 20- to 60-s surface waves is now fairly commonplace; improved data would allow extension of this technique to longer periods.

## An Ideal VLP Array

A worldwide array of instruments to measure very long period ground motions would be valuable to geophysical research. What, ideally, would be the nature of such an array? Clearly, since it is to measure primarily worldwide phenomena, the stations should be distributed over the earth as evenly as possible. The spacing would depend on the number of stations. Ideally, there can never be enough, but in reality, there are economic limits. The coverage of a sparse network can be somewhat improved by combining data from many earthquakes as well as from many stations, but in order to do this there must be enough stations in the first place to determine the source mechanism.

Each station would ideally record three components of ground motion and also atmospheric pressure.

Since pressure is coherent with horizontal ground motion [*Savino et al.*, 1972], barometric data could be used to reduce the recorded ground noise. This presumes that the instruments are quiet enough to detect ground noise, an obvious requirement but also a difficult one to meet. Ground noise at very long periods has been little studied. *Haubrich* [1970] showed that vertical noise is flat in acceleration and has a power density of approximately $10^{-18}$ m$^2$ s$^{-3}$ over the frequency range from 1 to 15 mHz. A very long period instrument must have a flat response to acceleration from 0 to 15 mHz; at periods of more than about 1 hour the instrument will be responding as much to changes in the gravitational potential as to actual ground acceleration. With this bandwidth, for quiet sites, the rms fluctuations on the instrument output during quiet periods would correspond to about $10^{-10}$ m s$^{-2}$ (0.01 $\mu$Gal). The acceleration from mantle Rayleigh waves from a magnitude 8½ earthquake 90° away could be up to $10^{-4}$ m s$^{-2}$ [*Brune and King*, 1967]. The instrument would thus need to have a dynamic range of about 120 dB; this means that the data would have to be recorded digitally, which is desirable in any case because virtually any interpretation of this sort of data requires digital processing. It is also necessary that the instrument response be linear over this range. Finally, the instruments must be calibrated and the response well determined in both gain and phase. The accuracies required vary from user to user. For some seismological studies, 10% accuracy would be adequate, while for studies of ocean tide loading, 0.1% accuracy would be needed to constrain the load effects [*Slichter*, 1972].

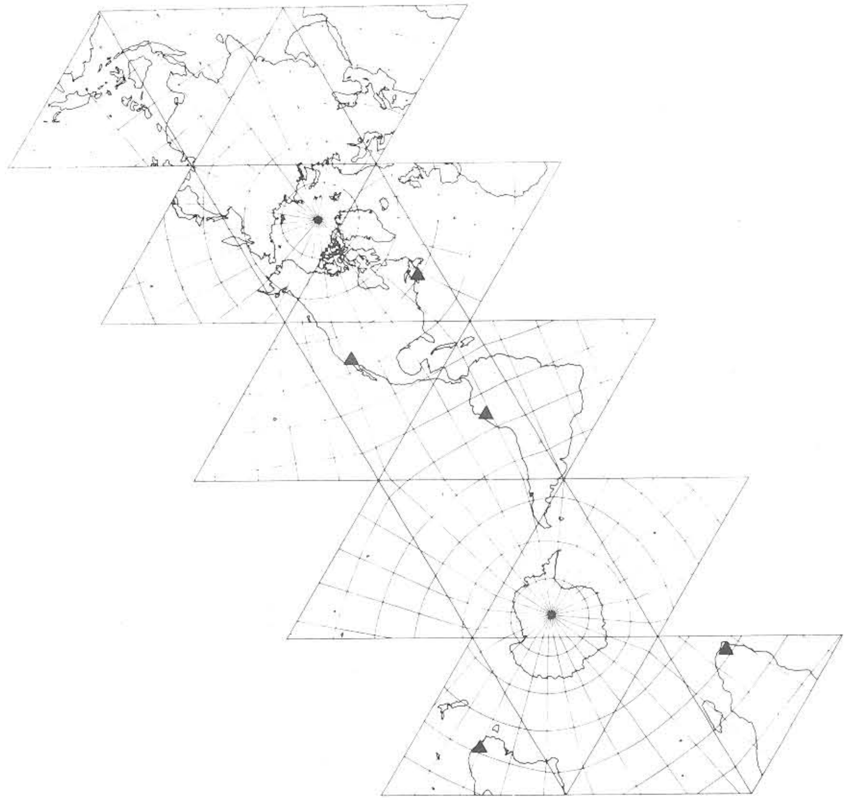


Fig. 1. The icosahedral mosaic. The surface of the earth has been projected onto the 20 faces of a circumscribed icosahedron by using the gnomic projection. The centers of each face are 41.81° apart. The mosaic illustrated is the one whose centers best fit the present five-station network, in a least squares sense. The rms deviation is 13.3°.

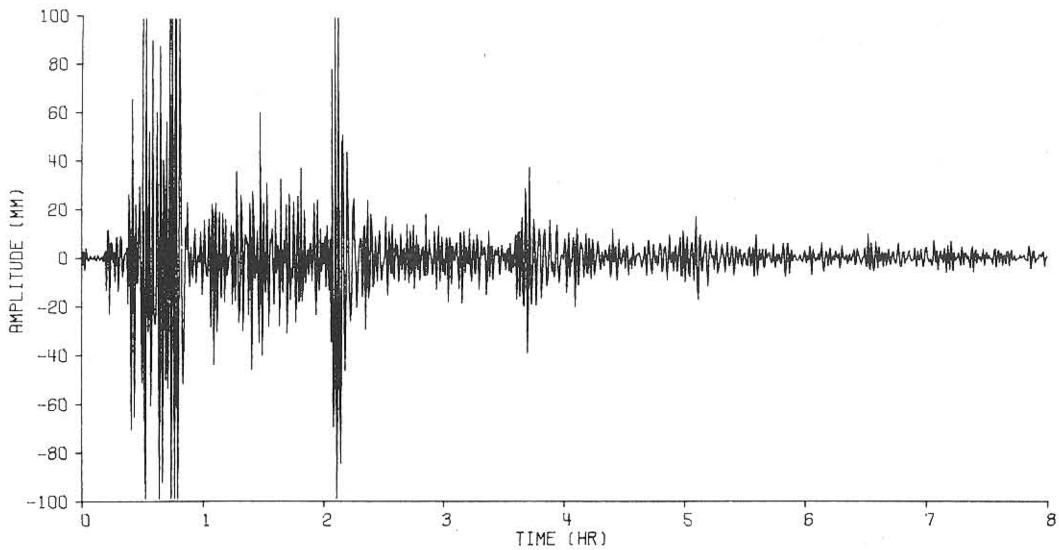
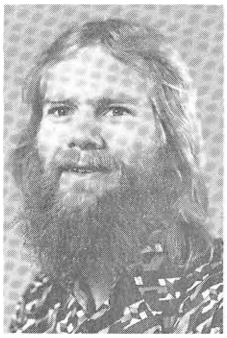



Fig. 2. A vertically polarized synthetic seismogram generated by a point source low-angle thrust with a moment of $1.5 \times 10^{29}$ dyn cm. The synthetic record was amplified by a WWSSN long-period (amplitude only) response and was low-pass filtered. Note that R1 and R2 have been clipped at $\pm$ 100 mm.



Raymond P. Buland was born in California in 1948 and graduated from the California Institute of Technology in 1970. He is a Ph.D. candidate at the University of California, San Diego. Research for his Ph.D. dissertation, conducted at the Institute of Geophysics and Planetary Physics, is concerned with the spectral characteristics of earthquake sources.

## IDA: A Real Array

The network that we have been creating falls short, inevitably, of the ideal but not, we trust, by too much. For economic reasons we do not intend to have more than 20 stations; we feel that routine maintenance and data processing from a larger network would be beyond our present means. An even spacing of 20 stations around the earth is illustrated in Figure 1. The exact siting of stations is largely determined by operational constraints. We prefer to place instruments at places where there is an existing seismic station with available vault space, personnel to operate the instrument, and reliable power. In addition to the Piñon Flat Observatory in California, we have installed instruments at Canberra, Australia; Ñaña, Peru; Sutherland, Republic of South Africa; and Halifax, Nova Scotia.

Economic limits have also resulted in our measuring only vertical ground motion at this time. Horizontal ground noise is greater than vertical ground noise and can be reduced to comparable size only by putting the instrument underground [*Peterson and Orsini*, 1976]. This is not to deny the value of horizontal components, and we hope to add them to the network eventually.

The vertical seismometer is a La Coste-Romberg gravimeter. We use this instrument because of its small size, low noise, and the very low drift of its zero-length spring. The gravimeter has a three-plate capacitor, the two outer plates fixed and the center plate moving with the mass. This capacitor provides for both the position sensing and the feedback forcing of the mass. The system design is essentially that of *Block and Moore* [1966]. Figure 3 shows a block diagram of the electronics used. The essential feature of the position detection system is that detection is performed with a narrow band ac signal. A 5-kHz reference voltage is applied symmetrically to the outer plates, and the voltage induced on the center plate, after amplification, is the input to the lock-in amplifier. This device selectively amplifies a narrow frequency band centered at 5 kHz and converts this

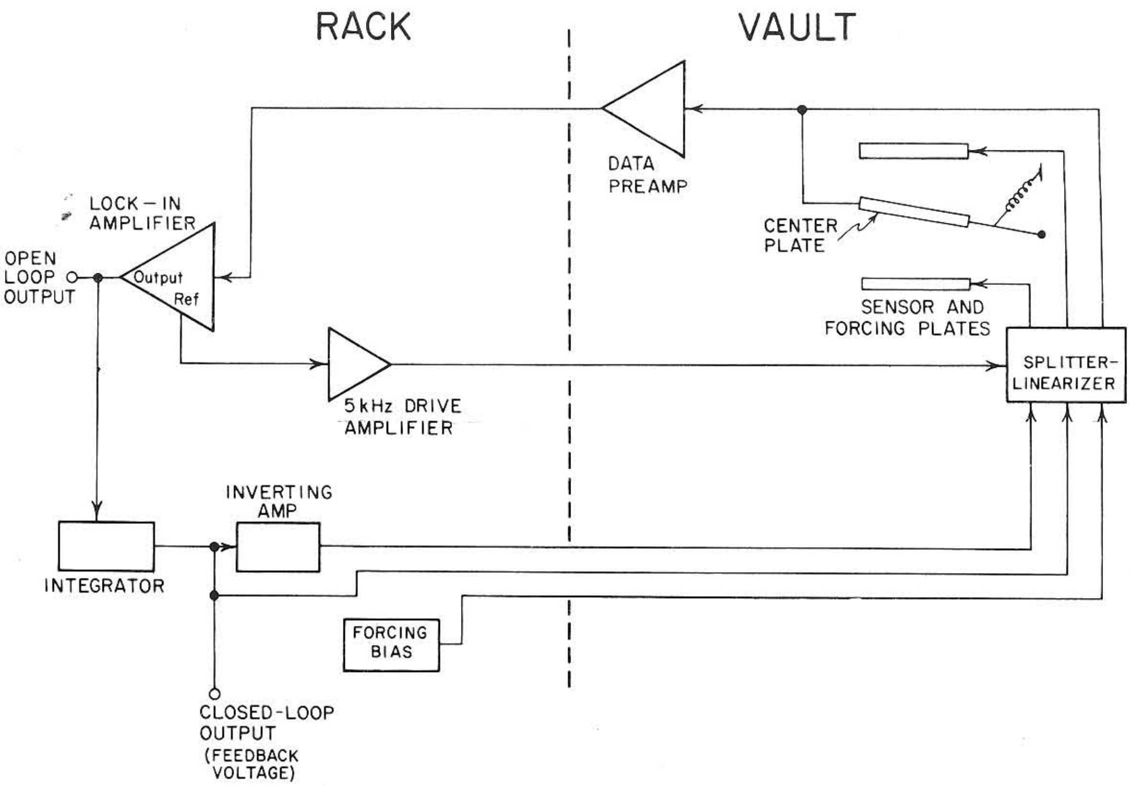



Fig. 3. Block diagram of sensing and feedback electronics for the IDA system.

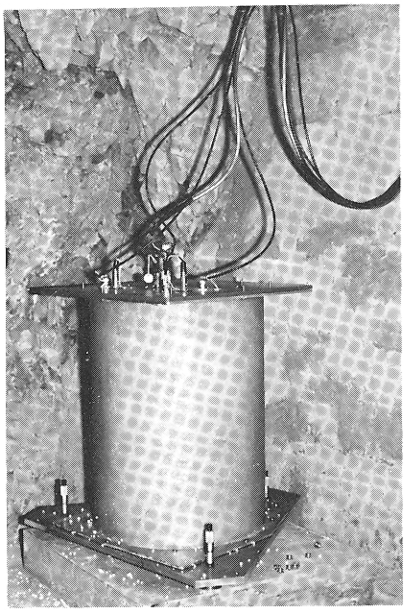


Fig. 4. The La Coste-Romberg gravimeter in its temperature-controlled can. The micrometers on the baseplate are for initial leveling and subsequent tilt calibration.

narrow band ac information to an equivalent bandwidth at dc. Thus, for example, if we wished to have position information in the band 0–10 mHz, we could use the ac band 5 kHz ± 5 mHz, an equivalent relative bandwidth of $10^{-6}$. Using such an extremely narrow band ac signal minimizes the problems of electronic noise in the circuits which handle very small signals. The essential element in the lock-in amplifier is a phase sensitive detector, which generates a voltage whose magnitude is proportional to the 5-kHz component in the input signal and whose sign depends upon the phase lead or lag of the input relative to the reference signal.

The output of the lock-in amplifier, which is thus proporional to the mass displacement, is then fed to an integrator circuit; the output of this (and −1 times it) is applied to the outer capacitor plates. A constant bias voltage is put on the center plate. These voltages set up electrostatic forces on the center plate and are arranged to keep it centered. Negative feedback is thus used to center the mass; the integrator is needed to keep the loop stable (*Moore*, 1966]. Using feedback has a number of advantages. The motion of the mass is less than it would be without feedback by a factor equal to the loop gain (about 1000), and this decreased motion makes the instrument that much more linear. The feedback voltage, which is the output of the instrument, is directly proportional to the force on the mass and hence to the acceleration of the instrument. The constant of proportionality depends only on the capacitance of the parallel plate arrangement and the bias voltage, so that it is easy to keep the calibration of the instrument constant.

Rigid environmental control is also needed to assure low noise operation. The instrument is hermetically sealed inside a thick-walled can which has resistance wire wound around the outside. A thermistor sensor and feedback circuit control the current through this wire so that the temperature of the instrument varies by less than $5 \times 10^{-4}$ °C. The temperature is kept close to the 'inversion point,' at which the spring is, to first order, insensitive to temperature fluctuations. This temperature is between 50° and 60°C. To reduce power requirements for heating and to cushion the instrument from severe thermal shocks, the inner can sits inside a larger one filled with expanded polystyrene beads and is attached to the instrument baseplate by stainless steel tubing. This outer can (see Figure 4) is 46 cm in diameter and 60 cm high, and its lid holds those parts of the electronics which must be close to the actual in-

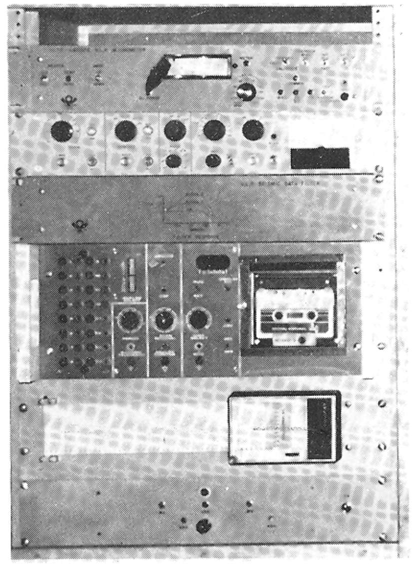


Fig. 5. The system electronics rack including, from the top, the feedback and bias control chassis, the lock-in amplifier, the mode filter, the cassette recorder, the monitor strip chart recorder, and the power supply battery charger.

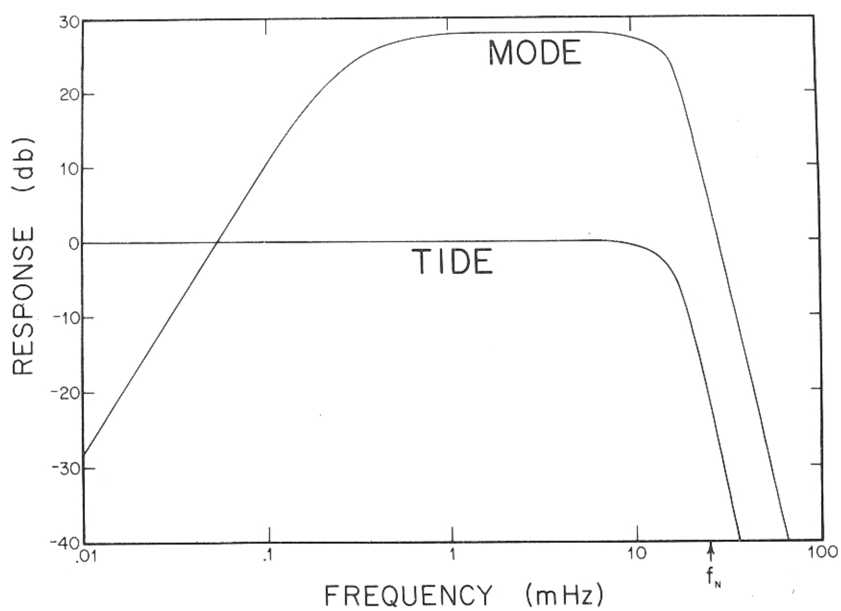
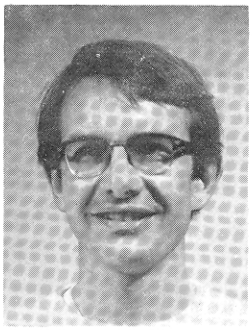



Fig. 6. The system amplitude response. There are two data channels: one has a response that is flat in acceleration from dc to 15 mHz with a nominal gain of $6.5 \times 10^{-7}$ m/s² V, rolling off at higher frequencies to eliminate aliasing at the recorder Nyquist frequency of 25 mHz. The mode channel has an enhanced sensitivity of $2.6 \times 10^{-8}$ m/s V in the band 0.3 mHz to 15 mHz.



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strument. Most of the electronics is in a rack (Figure 5), which may be up to 100 m away.

The system power supply operates from batteries which are continuously charged from the local ac source, and thus the system is independent of ac power for as long as a day. A large dynamic range requires digital recording, but in order to get the needed range and still use an inexpensive analog to digital converter some analog prefiltering is necessary (Figure 6). The filter outputs go to the digital data logger, which records the data on standard Phillips-type cassettes. (The output of the tide channel also goes to a chart recorder for monitoring instrument drift.) We record two channels every 20 s, at which rate the cassettes last over 10 days. The analog to digital converter in the recorder has 12-bit resolution, and this in conjunction with the two filter channels gives the instrument a dynamic range of 100 dB. The recorder has built into it a crystal-controlled clock, which records the time and is kept set to the nearest 0.5 s; this is more than adequate in the VLP frequency range. The cassettes are sent back by air mail to La Jolla, where they are unpacked and reformatted with a Digital Equipment Corporation PDP-11 minicomputer. Tapes written by this system are then sent to the Lawrence Berkeley Laboratory (LBL) computer system at the University of California, Berkeley, where all data are put on a photo-digital mass storage system. Tapes of a wide range of densities and formats can be written on the LBL system so that the data will be easily accessible to other institutions.

Calibrating the instrument requires knowledge of both the gain and the phase shift as a function of frequency and of the output in volts for a change in acceleration (the dc response). The latter is more difficult to determine accurately. It may be found by tilting the instrument to simulate a change in gravity, which gives an absolute calibration to within 2%. Each instrument is also run at Piñon Flat Observatory for comparison with the superconducting gravimeter [*Prothero and Goodkind,* 1972], which is calibrated to within 0.5%. Gain and phase are determined by inserting a random telegraph signal (generated by a circuit in the data logger) into the feedback loop; cross correlation of this input with the filter outputs gives the complete instrument response. The accuracy possible with this method is limited only by the length of time that this test is run [*Moore and Farrell,* 1970].

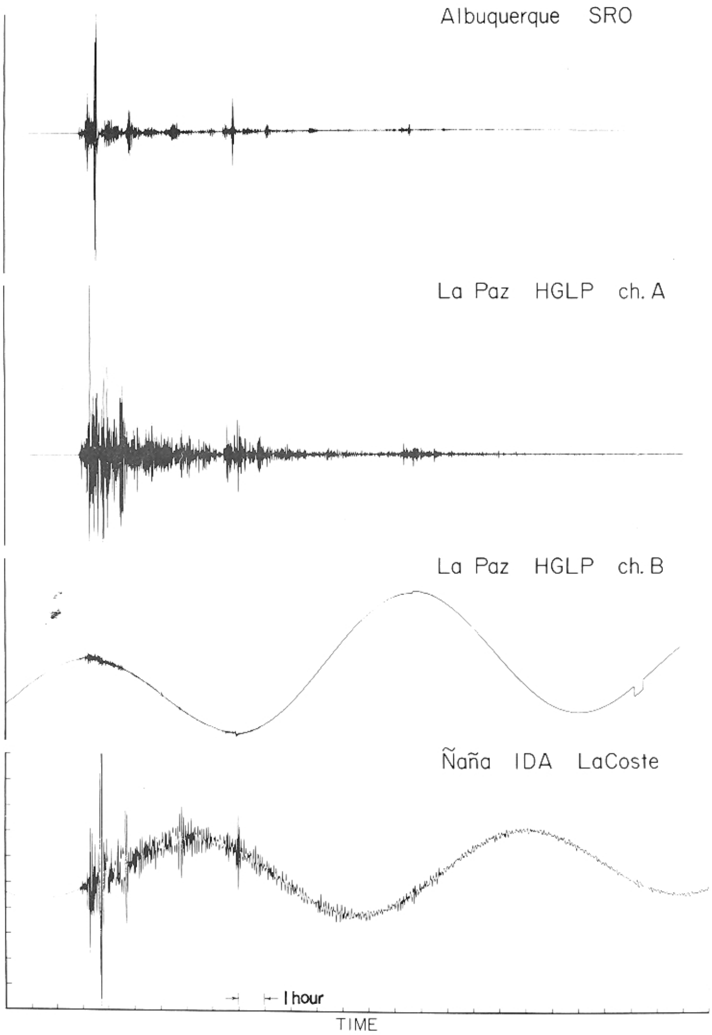


Fig. 7. The Solomon Islands earthquakes of July 20 and 21, 1975. The two largest events had a surface wave magnitude of 7.8 and 7.6. The SRO and HGLP channel A systems have a response peaked near 30 s (33 mHz). The records have been low-pass filtered to remove energy above 10 mHz. The HGLP channel B is a very low gain output designed primarily to monitor the mass position for releveling. One division on the IDA instrument corresponds to $1.3 \times 10^{-8}$ m/s² in the passband of 0.3 mHz to 15 mHz.

## Conclusion

As an example of the data quality, Figure 7 shows a sequence of earthquakes recorded by the SRO, HGLP, and IDA systems. All stations are at approximately the same epicentral distance, and the HGLP and IDA stations are within 10° of each other. What is quite apparent from these seismograms is the very different responses of the various systems. The IDA system, which is designed specifically to provide data in the VLP band, shows continued low-frequency activity for many hours after the event.

An examination of the Fourier spectrum of the event in Figure 8 gives a clear indication of the excitation of normal modes down to ${}_0S_3$ (0.47 mHz) and possibly even ${}_0S_3$ (0.31 mHz), the gravest elastic gravitational mode. The signal to noise ratio for this record is nearly 30 dB throughout most of the VLP band.

The ability to observe modal activity from such relatively frequent events will greatly increase the rate of data acquisition. Since modal excitation is linearly proportional to earthquake moment, every 10-dB decrease in noise level results in a twofold increase in the frequency of events with observable modal activity [*Chinnery and North*, 1975]. No longer need we wait for truly large and very infrequent earthquakes to add to our observational knowledge of the earth's free oscillations.

Finally, a comparison of the noise levels at Ñaña the day before the events with the results of *Haubrich* [1970] demonstrates that over this frequency band the Ñaña noise levels are slightly lower than those recorded at Tonto Forest Observatory at Payson, Arizona, with a similar system. We believe that these are the lowest noise levels ever recorded in the band and indicate not only the excellence of this site but also the low noise of the IDA system.

## Acknowledgments

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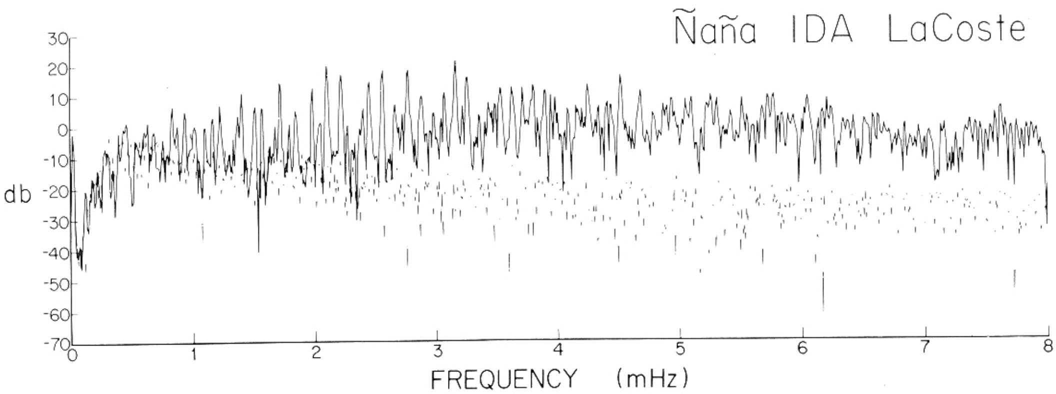


Fig. 8. Fourier spectrum of the Solomon Islands earthquakes (top plot) and the day before (bottom plot). Here, 5000 samples of 20-s data have been processed. The R1 pulse from the largest event has not been included in the analysis. The point 0 dB corresponds to a power level of $10^{-16}$ m$^2$ s$^{-3}$ [(m/s$^2$)$^2$/Hz].

Miller. We wish to acknowledge the artistry of R. L. Parker for his orientation and production of the icosahedral mosaic.

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