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SEISMIC STRUCTURE

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Introduction

Seismic exploration of the oceans began in earnest in the 1950s. The early seismic experiments were refraction in nature using explosives as sources. The principal data were first arrival, P-wave travel times, which were analyzed to produce primarily one-dimensional models of compressional velocity as a function of depth. Within a decade, the results of these experiments had convincingly demonstrated that the crust beneath the ocean crust was much thinner than continental crust. Moreover, the structure of the deep ocean was unexpectedly uniform, particularly when compared with the continents. In light of this uniformity it made sense to talk of average or 'normal' oceanic crust. The first compilations described the average seismic structure in terms of constant velocity layers, with the igneous crust being divided into an upper layer 2 and an underlying layer 3.

Today, the scale and scope of seismic experiments is much greater, routinely resulting in two- and three-dimensional images of the oceanic crust. Experiments can use arrays of ocean bottom seismographs and/or multichannel streamers to record a wide range of reflection and refraction signals. The source is typically an airgun array, which is much more repeatable than explosives and produces much more densely sampled seismic sections. Seismic models of the oceanic crust are now typically continuous functions of both the horizontal and vertical position, but are still principally P-wave or compressional models, because S-waves can only be produced indirectly through mode conversion in active source experiments.

In spite of their greater resolving power, modern experiments are still too limited in their geographic scope to act as a general database for looking at many of the questions concerning oceanic seismic structure. The main vehicle for looking at the general seismic structure of the oceans is still the catalog of one-dimensional P-wave velocity models built up over approximately 40 years of experiments. The original simple layer terminology, with slight elaboration, is by now firmly entrenched as the means of describing the principal seismic features of the oceanic crust; despite the fact that the representation of the underlying velocity structure has changed significantly over time. The next section discusses the evolution of the velocity model and the layer description. Subsequent sections discuss the interpretation of seismic structure in terms of geologic structure; the seismic structure of anomalous crust; and the relationship of seismic structure to such influences as spreading rate and age.

Normal Oceanic Crust

Table 1 reproduces one of the first definitions of average or 'normal' oceanic crust by Raitt (1963). Even in this era before plate tectonics, Raitt excluded from consideration any areas such as oceanic plateaus that he thought atypical of the deep ocean. Today, compilations count as normal crust formed at midocean ridges away from fracture zones. The early refraction experiments typically consisted of a small set of widely spaced instruments. They were analyzed using the slope-intercept method in which a set of straight lines was fitted to first arrival travel times. This type of analysis naturally leads to stair-step or 'layer-cake' models consisting of a stack of uniform velocity layers separated by steps in velocity. Although their limitations as a description of the earth were recognized, these models provided a simple and convenient means of comparing geographically diverse data sets. Raitt divided the oceanic crust into three layers and included a fourth to represent the upper mantle. The top layer (layer 1), was a variable thickness sedimentary layer. Below this came the two layers that together comprised the igneous oceanic crust, a thinner more variable velocity layer (layer 2), and a thicker, more uniform velocity layer (layer 3). Layer 3 is the most characteristically oceanic of the layers. Arrivals from this layer are the most prominent arrivals in typical refraction profiles. The uniformity of high velocities within this layer mark layer 3 as being compositionally distinct from continental crust. At the base of layer 3 is the Mohorovic discontinuity or Moho, identifiable as such because the velocities of layer 4 were comparable to those seen in the upper mantle beneath continents.

As refraction data sets with better spatial sampling became available, the systematic errors inherent in fitting a few straight lines to the first arrival travel times became more noticeable. This was especially true for layer 2 first arrivals, which appear over a relatively short-range window, but have noticeable curvature because of the wide range of layer 2 velocities. The initial resolution of this problem was to fit more lines to the data and divide layer 2 into smaller, constant velocity sublayers termed 2A, 2B, and 2C. However, there was a more fundamental problem: the layer-cake models were not consistent with the waveform and amplitude behavior of the data. This flaw became apparent when, instead of just using travel times, the entire recorded wavefield began to be modeled using synthetic seismograms. Waveform modeling led to a recasting of the one-dimensional model in terms of smoothly varying velocities, constant velocity gradients, or finely layered stair-steps. Large velocity steps or interfaces are now included in the models only if they are consistent with the amplitude behavior. The stair-step representation is a tacit admission that there is a limit to the resolution of finite bandwidth data. A stair-step model is indistinguishable from a continuous gradient provided the layering is finer than the vertical resolution of data, which for refraction data is some significant fraction of a wavelength. Today, purely travel time analysis based upon densely sampled primary and secondary arrival times and accumulated knowledge can yield accurate models, but seismogram modeling is still required to achieve the best resolution.

The change in the style of the velocity models is illustrated in **Figure 1**, which shows models for a recent Pacific data set. A change in gradient rather than a jump in velocity marks the boundary between layer 2 and 3 in most modern models.

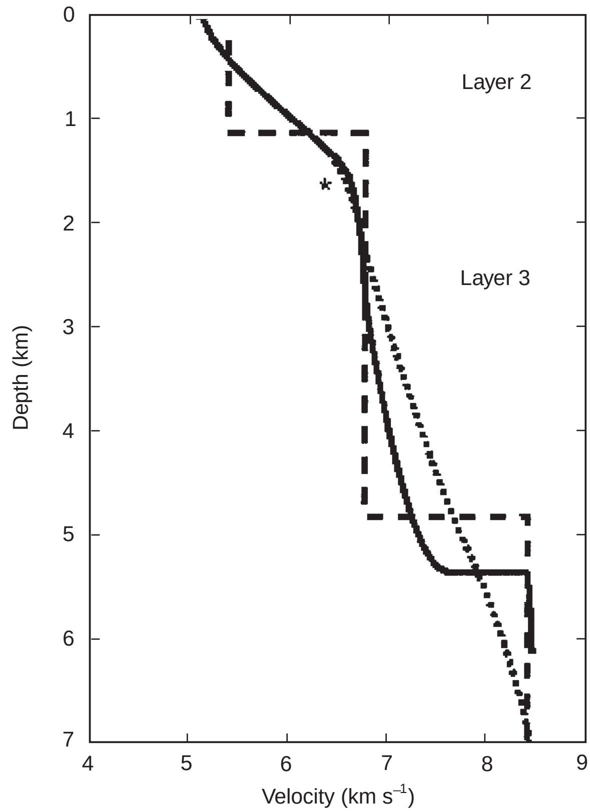

Figure 1 Velocity models for a recent Pacific data set.

Paradoxically, the jump in velocity at the Moho – present in the traditional layer-cake model – is not required by the first arrival times, but is required to fit the secondary arrival times and amplitude behavior of the data. The example also illustrates another general problem with layer-cake models, which is that they systematically underestimate both layer and total crustal thickness.

Table 1 also presents a more modern summary of average oceanic crustal structure. All models included in this compilation were the result of synthetic seismogram analysis. A range of velocities and typical gradients now characterizes layers 2 and 3. Layer 2 is a region of rapidly increasing velocity at the top of the crust, with typical gradients from 1–3 s^{-1}, while layer 3 is a thicker region of more uniform velocity with gradients between 0 and 0.2 s^{-1}. The layer thicknesses are little changed from the Raitt compilation but are systematically thicker and thus total crustal thickness is also larger.

A virtue of the layer description is that it captures the main features of the seismic data and results without being too precise. So, while the style of velocity model has moved away from one that is strictly layered, the description in terms of layers persists both for historical continuity and linguistic

Table 1 Traditional and modern summaries of average oceanic crystal structure

Parameter	*Traditional*[a]		*Modern*[b]		
	Velocity ($km\,s^{-1}$)	*Thickness (km)*	*Velocity ($km\,s^{-1}$)*	*Thickness (km)*	*Representative gradient (s^{-1})*
Layer 2 (igneous crust)	5.07 ± 0.63	1.71 ± 0.75	2.5–6.6	2.11 ± 0.55	1–3
Layer 3 (igneous crust)	6.69 ± 0.26	4.86 ± 1.42	6.6–7.6	4.97 ± 0.90	0.1–0.2
Layer 4 (upper mantle)	8.13 ± 0.24		> 7.6		
Total igneous crust		6.57 ± 1.61		7.08 ± 0.78	

[a]From Raitt (1963).
[b]Modified from White et al. (1992).

convenience. Without a simple layer-cake model, defining the layer 2/3 transition can be somewhat problematic. Ideally the velocity function will show a small velocity jump or resolvable inflection at the boundary, but often the layer 2/3 boundary is taken as being either at a change in the general velocity gradient of the model or at a particular velocity just below representative layer 3 velocities, sometimes both. The chosen velocity is typically somewhere between 6.5 and $6.7\,km\,s^{-1}$.

The use of layers 2 and 3 is almost universally and consistently applied when summarizing the basic features of both seismic data sets and models. More problematic is the use of layers 2A, 2B, and 2C to describe subintervals layer 2. While the use of these layers is widespread in the literature, their application is more variable and has evolved in conjunction with changes in model style and resolution. As a result caution is needed when comparing models from disparate experiments, particularly across tectonic and geographic regions. Today, many authors subdivide layer 2 into 2A and 2B only. Layer 2A is widely recognized as a well defined surfical layer in young oceanic crust near ridges, being associated with velocities $< 3\,km\,s^{-1}$ and a transition to velocities $> 4\,km\,s^{-1}$ at its base. In this division, layer 2B is simply the lower part of layer 2.

Interpretation of Seismic Structure

The nature of the relationship between the seismic and geologic structure of the oceanic crust is the subject of ongoing debate. The adoption of plate tectonics and seafloor spreading provided a framework for understanding the initially surprising uniformity and simplicity of ocean seismic structure. It also gave rise to the hope that there would be a correspondingly simple and universal interpretation of the seismic layering in terms of geologic structure. However, this expectation has receded as the complexity of the seismic models has increased and our understanding of the diverse magmatic, tectonic, and hydrothermal processes shaping crustal structure, both spatially and temporally, has improved. Fundamentally, there is no unique, unambiguous interpretation of seismic velocity in terms of rock type or geologic structure. Having coincident P- and S-wave velocity models plus other geophysical data can considerably reduce this ambiguity but ultimately cannot eliminate it. Ideally, reference drill holes through the full oceanic crustal section would be used to calibrate seismic and other geophysical results. These would allow the dominant processes, controlling, for example, the layer 2/3 boundary or the nature of layer 2A to be identified in different tectonic settings. Unfortunately, to date only a limited number of drill holes have penetrated a significant depth into the oceanic crust, and none have penetrated a full crustal section. As a result there has been only limited opportunity for direct comparison between seismic and *in situ* structure. When interpreting seismic results, we must still rely heavily on inferences that draw upon a number of less direct sources including seafloor observations, analogy with ophiolites, and laboratory measurements on dredged rock samples.

The simplest, most straightforward interpretation of seismic velocity is to assume that velocity is dependent on composition and that different velocities indicate different rock types. This reasoning has guided the traditional interpretation of the Moho boundary. In seismic models, the characteristic signature of the Moho is an increase in velocity of between ~ 0.5 and $1.0\,km\,s^{-1}$ to velocities $> 7.6\,km\,s^{-1}$. The increase may occur either as a simple interface (at the resolution of the data) or as a transition region up to a kilometer thick. In reflection data, the Moho is often observed as a low frequency, $\sim 10\,Hz$, quasi-continuous event. In general, layer 3 is considered to be predominantly gabbros, and the most common interpretation of the

Moho is as the boundary between a mafic gabbroic crust and an ultramafic upper mantle. The observed Moho structures reflect either a simple contact or a transition zone of interleaved mafic and ultramafic material. These interpretations are supported by observations within ophiolites, reference sections of oceanic lithosphere obducted onto land. However, partially serpentinized, ultramafic peridotites can have P- and S-wave velocities that are for practical purposes indistinguishable from gabbros, when the degree of alteration is between 20 and 40%. For lesser degrees of alteration, the serpentinized rocks will have velocities that are intermediate between gabbros and peridotite and are thus distinguishable. Although widespread serpentinization is unlikely, particularly at fast and intermediate spreading rates, it may be locally important at segment boundaries at slow spreading rates where faulting and extension expose peridotites to pervasive alteration by deeply circulating sea water.

A further complication in interpreting the Moho is the fact that the igneous crust may contain cumulate ultramafic rocks such as dunite, which crystallized as sills from a melt. Seismically, these are indistinguishable from the residual upper mantle harburgites that yielded the crustal melts. Thus a distinction is sometimes made between Moho defined seismically, and a petrologic Moho separating cumulate rocks from source rocks.

There is very little intrinsic difference in composition or velocity to the basaltic rocks – pillow lavas and sheet flows, sheeted dikes, and gabbros – that typically constitute the upper part of the oceanic crust. Certainly not enough to explain the large range of layer 2 velocities or the difference between layers 2 and 3. Instead the seismic character is attributed primarily to the cracks, fractures, and fissures that permeate the upper crust. These reduce the effective stiffness of the rock matrix, and hence the velocity of the upper crust at seismic wavelengths. The relationship between velocity and the size, shape, orientation, and distribution of cracks is a complex nonlinear one that affects P- and S-wave velocities differently. However, at least informally, it is often crack volume or porosity that is taken as the primary control. The large velocity gradients within layer 2 are then seen as being the result of a progressive closure of cracks or reduction in porosity with depth, as confining pressure increases. Once most cracks are closed, velocities are only weakly pressure-dependent and can have the low velocity gradients characteristic of layer 3. An analogous velocity behavior, but with a smaller velocity range is observed in individual rock samples subject to increasing confining pressure.

From this perspective, there can only be a structural or lithologic interpretation of the seismic layers if there is a structural dependence to the crack distribution. Support for such an interpretation comes from composite velocity profiles through ophiolites, constructed using laboratory measurements on hand samples at suitable confining pressures. These profiles showed a broad agreement with oceanic results and led to the standard ophiolite interpretation of seismic structure in which layer 2 is equated with the extrusive section of pillow lavas and sheeted dikes and layer 3 with the intrusive gabbroic section. Moreover, the relative and absolute thicknesses of the extrusive and intrusive sections in ophiolites are comparable to those of the oceanic seismic layers. At a more detailed level, measurements on ophiolites often show a reduction in porosity at the transition from pillow basalts to sheeted dikes: an observation used to bolster the inference that layer 2A is equivalent to pillow lava section in young oceanic crust.

As the only available complete exposures of oceanic crustal sections, ophiolites have had a historically influential role in guiding the interpretation of seismic layering. However, a number of cautions are in order. First, there are inherent uncertainties in extrapolating seismic velocities measured on hand samples to the larger scales and lower frequencies characteristic of seismic experiments. Second, the seismic velocities of ophiolites could have been modified during the obduction process. Finally, most ophiolites are thought to have been produced in back arcs or marginal basin settings and thus while valuable structural analogs may not be representative of the ocean basins as a whole. Ultimately, the ophiolite model can only be used as a guide, albeit an important one, for interpreting oceanic structure, seismic or otherwise and conclusions drawn from it must be weighed against other constraints.

The traditional ophiolite model is a convenient and widely used shorthand for describing seismic structure, that is useful provided that too much is not asked or expected of it. The porosity interpretation of upper crustal velocities gives the basic layer 2/layer 3 division of seismic models a sort of universality that transcends, within bounds, changes in the underlying lithologic structure, and emphasizes the need for taking tectonic setting into consideration when interpreting seismic structure. Any process that either resets or significantly modifies this crack/porosity distribution of the upper crust will imprint itself on the seismic structure. For example, near fracture zones, fracturing and faulting are usually inferred to be dominant controls on layer

2 structure. At Deep Sea Drilling Project Hole 504B, a deep penetration hole in 5.9 million year old crust formed at intermediate spreading rates, the base of seismic layer 2 is found to lie within the sheeted dike section. It is thought that progressive filling of cracks by hydrothermal alteration processes has, over time, raised the depth of the layer 2/3 boundary.

Anomalous Crust

Definitions of normal seismic structure focus on oceanic crust formed at midocean ridges and specifically exclude structures such as fracture zones or oceanic plateaus as anomalous. Particularly at slow spreading rates, fracture zones and the traces of segment boundaries are part of the warp and weft of ocean fabric, making up about 20% of the seafloor. These have been most extensively studied in the northern Atlantic, where the ridge is segmented on scales of 20–100 km, and most segment boundaries are associated with attenuated crust. The degree of crustal thinning shows no simple dependence of segment offset, although the large-offset fracture zones may indeed contain the most extreme structure. Fracture zones and segment traces typically exhibit an inner and outer region of influence. In the outer region, there is gradual thinning of the crust towards the trace extending over a distance of perhaps 20 km. This region is marked by a deepening of the seafloor and a simultaneous shoaling of the Moho. Within the outer region, seismic structure is a thinned but recognizable version of normal crust, and the most extreme structure is associated with the inner region. Here, within a ~ 10 km wide zone, the crust may be < 3 km thick and in one-dimension may appear to be all layer 2 down to the Moho. Looked at in cross-section, there is a coalescence of a gradually thickening layer 2 with a Moho transition region, and the elimination of layer 3. The structure at segment boundaries can be explained as a combination of reduced magma supply, much of it feeding laterally from the segment centers, and pervasive faulting, possibly low angle in nature.

Velocities intermediate between crust and mantle values, 7.1–7.4 $km\,s^{-1}$, are observed beneath small offset transforms and nontransform offsets below about 4 km depth, most likely indicating partial serpentinization of the mantle by deeply circulating water. The upper limit of serpentinization is difficult to determine seismically because of the overlap in velocity between gabbros and altered peridotites at high degrees of alteration. But, from seafloor observations, at least some serpentinites must lie at shallower depths.

In the fast spread Pacific, fracture zones – which are spaced at intervals of a few hundred kilometers – affect only a relatively small fraction of the total crust. In addition, seismic studies suggest that the crustal structure of fracture zones is essentially a slightly thinner version of normal crust with a well-defined layer 3. In addition there can be some thickening and slowing of layer 2 in the vicinity of the transform associated with upper crustal faulting.

The term oceanic large igneous provinces (LIPs) provides a convenient umbrella under which to group such features as oceanic plateaus, aseismic ridges, seamount groups, and volcanic passive margins. They are massive emplacements of mostly mafic extrusive and intrusive material whose origin lies outside the basic framework of seafloor spreading. Together they account for much of the anomalous structure apparent in maps of seafloor bathymetry. At present, the rate of LIP emplacement, including the continents, is estimated to be equal to about 5–10% of midocean ridge production. However, during the formation of the largest LIPs, such as the Ontong Java plateau, off-axis volcanism was a significant fraction of midocean ridge rates. Many LIPs can trace their origin to either transient or persistent (hot-spot) mantle plumes; as such they provide a valuable window into the dynamics of the mantle. Where plumes interact with ridges, they can significantly affect the resulting crustal structure. The most notable example of this, at present, is the influence of the Iceland hot-spot on spreading along the Reykjanes ridge, where the crust is about 10 km thick and includes an approximately 7 km thick layer 3.

Two general features of LIPs seismic structure are a thickened crust, up to 25 km thick, and a high velocity, lower crustal body, reaching up to 7.6 $km\,s^{-1}$. However, in detail, the seismic structure depends on the style and setting of their emplacement, including whether the emplacement was submarine or subareal, intraplate or plate boundary. For example, the Kerguelen-Heard Plateau (a province of the larger Kerguelen LIP) is estimated to have 19–21 km thick igneous crust, the majority of which is a 17 km thick layer 3 with velocities between 6.6 and 7.4 $km\,s^{-1}$. The plateau is inferred to have formed by seafloor spreading in the vicinity of a hot-spot similar to Iceland. If this is the case, the greater thickness and higher velocities of layer 3 can be attributed to greater than normal extents of partial melting within the upwelling mantle. An example of an intraplate setting is the formation of the Marquesas Island hot-spot. Seismic data reveal that in addition to the extrusive volcanism responsible for the islands, significant intrusive emplacement has created a crustal root beneath the previously existing oceanic crust. Combined, the total crust is up to 17 km thick. The crustal root, with velocities between 7.3 and 7.75 km s^{-1}, may be purely intrusive or a mixture of intrusive rocks with preexisting mantle peridotites.

Systematic Features of the Oceanic Crust

For the most part, seismic investigations of the oceanic crust tend to focus on specific geologic problems. As a consequence, the catalog of published seismic results has sampling biases that make it less than ideal for looking at certain more general questions. There are for example a relatively large number of good measurements of young Pacific crust and old Atlantic crust, but fewer on old Pacific crust and only a handful of measurements on crust formed at ultra-slow spreading rates. Older data sets analyzed by the slope-intercept method are often discounted unless they are the only data available for a particular region. Nevertheless there are a number of systematic features of oceanic crust that can be discerned from compilations of seismic results.

Spreading Rate Dependence of Average Crustal Thickness

Although the style of crustal accretion varies considerably between slow and fast spreading ridges, the average thickness of the crust produced including fracture zones is remarkably uniform at 7 ± 1 km for full spreading rates between 20 and 150 mm a^{-1}. This result indicates that the rate of crustal production is linearly related to spreading rate over this range. Crustal thicknesses are more variable at slower spreading rates, reflecting the more focused magma supply and greater tectonic extension. At ultra-slow spreading rates below 20 mm a^{-1}, there is a measurable and rapid decrease in average crustal thickness. This reduction is expected theoretically, as conductive heat loss inhibits melt production in the upwelling mantle.

Age Dependence of Crustal Structure

The clearest and strongest aging signal in the oceanic crust is the approximate doubling of surficial velocities with age from about 2.5 km s^{-1} at the ridge axis to 5 km s^{-1} off-axis. This increase in velocity was first reported in the mid-1970s based on compilations of surface sonobuoy data. Originally, the velocity signal was interpreted as being associated with a thinning of layer 2A over a period of 20–40 Ma. However, the same data can equally well be explained as simply the increase in velocity of a constant thickness layer, and a compilation of modern seismic data sets indicates that layer 2A velocities increase much more rapidly, almost doubling in < 10 Ma. While both of these inferences are supported by individual flowline profiles extending out from the ridge axis, the distribution bias of modern seismic data sets to the ridge axes makes it hard to assess the robustness of this result.

The increase in layer 2A velocity with age is due to hydrothermal alteration sealing cracks within the upper crust. There need not be a correspondingly large decrease in porosity, as alteration that preferentially seals the small aspect ratio cracks will produce a large velocity increase for a small porosity reduction. Given this mechanism, similar, albeit smaller increases in layer 2B velocities might be expected. Such an increase is not apparent in present compilations, although a small systematic change would be masked by the intrinsic variability of layer 2B and the variability induced by different analysis methods. There is though some indication of systematic change with layer 2B from analysis of ratios of P- and S-wave velocity and as noted in the previous section alteration is thought to have raised the layer 2/3 boundary at Hole 504B.

Anisotropic Structure

Two types of anisotropic structure are frequently reported for the oceanic crust and upper mantle. The P-wave velocities of the upper mantle are found to be faster in the fossil spreading direction, than in the original ridge parallel direction, with the difference being around 7%. This is due to the preferential alignment of the fast a-axis of olivine crystals in the direction of spreading as mantle upwells beneath the midocean ridge.

The other region of the crust that exhibits anisotropy is the extrusive upper crust, which has a fast P-wave propagation direction parallel to ridge axis at all spreading rates. The peak-to-peak magnitude of the anisotropy averages $\sim 10\%$. Like the velocity structure, this shallow anisotropy is generally ascribed to the crack distribution within the upper crust. Extensional forces in the spreading direction are thought to produce thin cracks and fissures that preferentially align parallel to the ridge axis.

See also

Mid-Ocean Ridge Tectonics, Volcanism and Geomorphology. Seismology Sensors.

Further Reading

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SEISMOLOGY SENSORS

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Introduction

A glance at the globe shows that the Earth's surface is largely water-covered. The logical consequence of this is that seismic studies based on land seismic stations alone will be severely biased because of two factors. The existence of large expanses of ocean distant from land means that many small earthquakes underneath the ocean will remain unobserved. The difference in seismic velocity structure between continent and ocean intruduces a bias in locations, with oceanic earthquakes which are located using only stations on one side of the event being pulled tens of kilometers landward. Additionally, the depths of shallow subduction zone events, which are covered by water, will be very poorly determined. Thus seafloor seismic stations are necessary both for completeness of coverage as well as for precise location of events which are tectonically important. This paper summarizes the status of seafloor seismic instrumentation.

The alternative methods for providing coverage are temporary (pop-up) instruments and permanently connected systems. The high costs of seafloor cabling has thus far precluded dedicated cables of significant length for seismic purposes, although efforts have been made to use existing, disused wires. Accordingly, the main emphasis of this report will be temporary instruments.

Large ongoing programs to investigate oceanic spreading centers (RIDGE) and subductions (MARGINS) have provided impetus for the upgrading of seismic capabilities in oceanic areas.

The past few years has seen a blossoming of ocean bottom seismograph (OBS) instrumentation, both in number and in their capabilities. Active experimental programs are in place in the USA, Europe, and Japan. Increases in the reliability of electronics and in the capacity of storage devices has allowed the development of instruments which are much more reliable and useful. Major construction programs in Japan and the USA are producing hundreds of instruments, a number which allows imaging experiments which have been heretofore associated with the petroleum exploration industry. This contrasts sharply with the severely underdetermined experiments which have characterized earthquake

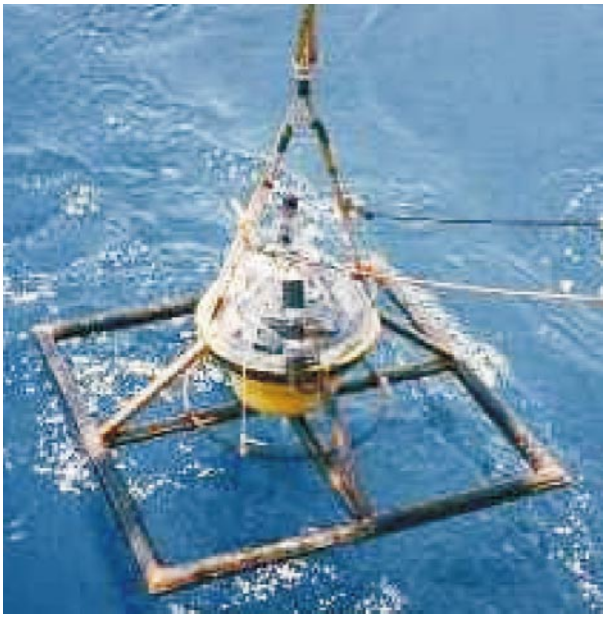

Figure 1 The UTIG OBS, a particularly 'clean' mechanical design, which has been in use for many years, with evolving electronics. The anchor is 1.2 m on each side. (Photograph by Gail Christeson, UTIG.)