Evolution of the ocean crust: results from recent seismic experiments

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SUMMARY: We present results from recent analyses of seismic refraction and sea-floor microseismicity studies in the Pacific and Atlantic oceans which lend support to the hypothesis that processes responsible for the construction of ophiolite suites are similar to phenomena extant at mid-ocean ridges. Seismicity at fast-spreading ridges is characterized by very low magnitude (0–1) and shallow (<2–3 km) microearthquakes and long-term oscillations or harmonic tremor. A detailed seismic-refraction experiment on a fast-spreading portion of the East Pacific Rise supports the hypothesis of the existence of a crustal magma chamber. Analyses of these data indicate that the chamber is largely unperturbed by the presence of conjugate spreading centres and the inverted delta-shaped zone of partial melt is characterized by a half width in excess of 6 km. Finally, a new approach for obtaining upper-crustal velocities when applied to data characterized by the presence of shear waves provides several counter-examples to the hypothesis that the shallow crust evolves with time.

During the past few years, marine seismology has proved to be an effective tool in studying the detailed elastic structure of the oceanic crust. Concomitant advances in the quality and quantity of physical properties' data from ophiolite suites has permitted detailed comparisons between these allochthonous terranes and the oceanic lithosphere. A properly testable 'ophiolite hypothesis' for the genesis, evolution and structure of the oceanic crust has arisen from this work. We present two new data sets which, in general, support this hypothesis. These consist of ocean-bottom seismograph seismicity and refraction studies of the East Pacific Rise which require a substantial crustal magma chamber. This chamber, by implication, is responsible for the bulk of the fractionation processes which construct a vertically heterogeneous crustal column. Finally, a recently developed method for accurately determining the compressional velocity of the shallow crust casts some doubt on the evolutionary systematics attributed to 'Layer 2A' and the oceanic crust in general.

Microearthquakes on the East Pacific Rise at 21°N

The fast-spreading East Pacific Rise is strikingly quiescent teleseismically and observed events are clearly confined to transform faults. Microearthquake surveys conducted near the rise crest have been no more successful than the teleseismic studies in detecting events which clearly lie along the spreading axis (Reid *et al.* 1977; Lilwall *et al.* 1981; Prothero & Reid 1982). This inactivity contrasts dramatically with the teleseismic and local observations along the Mid-Atlantic Ridge (e.g. Forsyth 1975; Weidner & Aki 1973; Lilwall *et al.* 1981). Riedesel *et al.* (1982) have recently exploited the growing body of knowledge of spreading-centre fault lengths obtained by detailed surveys on slow- and fast-spreading ridges to place upper bounds on the depth extent of the earthquake fault planes. Whereas faults along the slow-spreading Mid-Atlantic Ridge likely extend through the entire oceanic crust, East Pacific Rise events which escape detection on the worldwide seismic network must be characterized by faults less than 2 km in vertical extent.

Following the successes of the Rivera Submersible Experiment (RISE Project Group 1981) and the discovery of an extraordinarily active hydrothermal field at 21°N on the East Pacific Rise, we planned and executed a microseismicity survey of the rise axis in the vicinity of the hydrothermal vents (Riedesel *et al.* 1982). Five ocean-bottom seismographs (OBS) were deployed within and near (<2 km) the neovolcanic zone and located with an absolute accuracy of better than 100 m. Altogether more than 100 distinct events were recorded in the period 16 July 1980–8 August 1980. These events consisted of regional and local earthquakes as well as phenomena interpreted as volcanic tremor caused by the movement of magma or hydrothermal fluids.

The recorded microearthquakes allowed a geologically interesting estimate of the extreme depth of occurrence. The microearthquakes were extremely impulsive and the measurement of S-P times was easily performed. The earthquakes were quite small, the seismic moments were order 10^{17} dyne-cm corresponding to a magnitude 0–1. For these moments and a stress drop of one bar (Prothero & Reid 1982), the source area was on the order of 35 m^2. Eleven of these microearthquakes were recorded on three or more OBSs and their locations were determined to accuracies from 1 to several km. Even though many of the events appear to be close to the hydrothermal vents within the neovolcanic zone, the probable errors do not permit an unambiguous location at

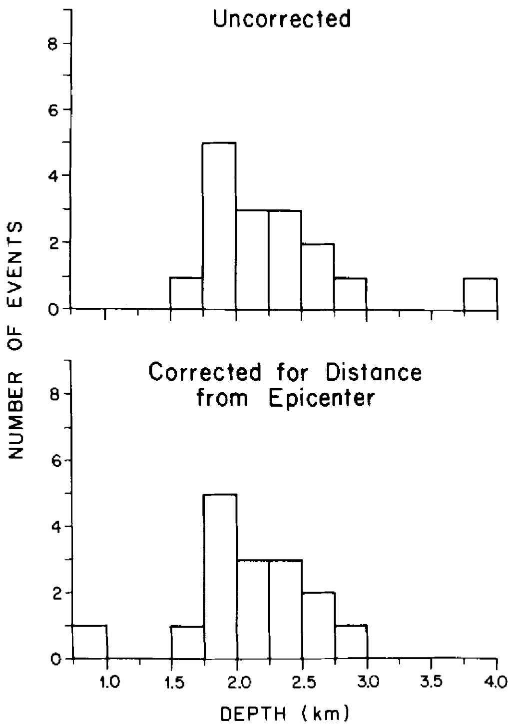

FIG. 1. A histogram of the depths for some earthquakes close to the vents. The S-P time at a given instrument is used to calculate the distance of the event. The top graph assumes that every event recorded at a given seismograph is directly beneath that OBS. The bottom graph corrects this for those few earthquakes with well-determined hypocentral depths.

the vents vis-à-vis the normal faults which bound the axis.

The S-P times alone however, require that the earthquakes be quite shallow. Figure 1a plots the maximum depth for the microearthquakes computed by assuming each event occurred beneath the recording station. Since the locations are not necessary, a large number of events recorded on only one or two instruments can be included in the study. The accurate epicentral locations of a number of these events were known so that their maximum hypocentral depths can be more accurately estimated. When these locations are employed, the corrected plot in Fig. 1b results. A maximum depth of 2–3 km appears to be reasonable for the microearthquakes although the results should not be interpreted to indicate a clustering between depths of 2 and 3 km.

Observed microearthquakes and brittle fracture clearly occurred at depths shallower than 3 km; this may be construed as a maximum depth of hydrothermal circulation and crustal embrittlement. This maximum depth coincides with the depth to the top of the crustal magma chamber at 21°N earlier reported by Reid *et al* (1977). We have closely examined the excitation of Stoneley or interface waves and conclude that the microearthquakes not only occur at depths of less than 3 km, but must occur very near the sea floor.

Two thirds of the events recorded apparently correspond to events interpreted as volcanic tremor. These events were very emergent, rising above the microseismic noise level and decaying, after a few minutes, below this noise level. The spectra were peaked between 3 and 5 Hz, frequencies somewhat higher than observed at subareal volcanoes such as Mt St Helens. The events are devoid of identifiable phases although the seismograms for a given source region were extremely repeatable from day to day. Furthermore, the temporal behaviour of individual events bore a strong resemblance to recordings obtained at Mt St Helens prior to the 18 May 1980 eruption.

This detailed experiment has provided strong evidence for considerable small-scale seismic activity at spreading centres with morphologies characteristic of fast-spreading rates. The seismicity is quite shallow, in contrast with the Mid-Atlantic Ridge, and seismic activity formerly associated only with subareal volcanoes is quite common on the sea floor. These observations are, furthermore, consistent with the presence of a shallow crustal magma chamber previously outlined by seismic-refraction studies and required by the 'ophiolite model' of the oceanic crust.

A detailed rise-crest refraction experiment—the MAGMA expedition

A seismic refraction experiment conducted on the East Pacific Rise at 8°N provided the first geophysical data which could be interpreted in terms of a crustal magma chamber associated with the genesis of the oceanic crust (Orcutt *et al.* 1975). Subsequently, a number of experiments on the East Pacific Rise provided further evidence for such a magma body (Reid *et al.* 1977; Bibee 1979; Herron *et al.* 1980; McClain & Lewis 1980; Hale *et al.* 1982) although similar experiments in the Atlantic were largely negative (Keen & Tramontini 1970; Poehls 1974; Whitmarsh 1975; Francis *et al.* 1977). Bibee (1979), although accumulating strong evidence for a rise-crest magma chamber, interpreted the data as requiring a sub-crustal rather than a crustal magma chamber. Such a structure was not only quite different from previous interpretations, but was not consistent with ophiolite models requiring a

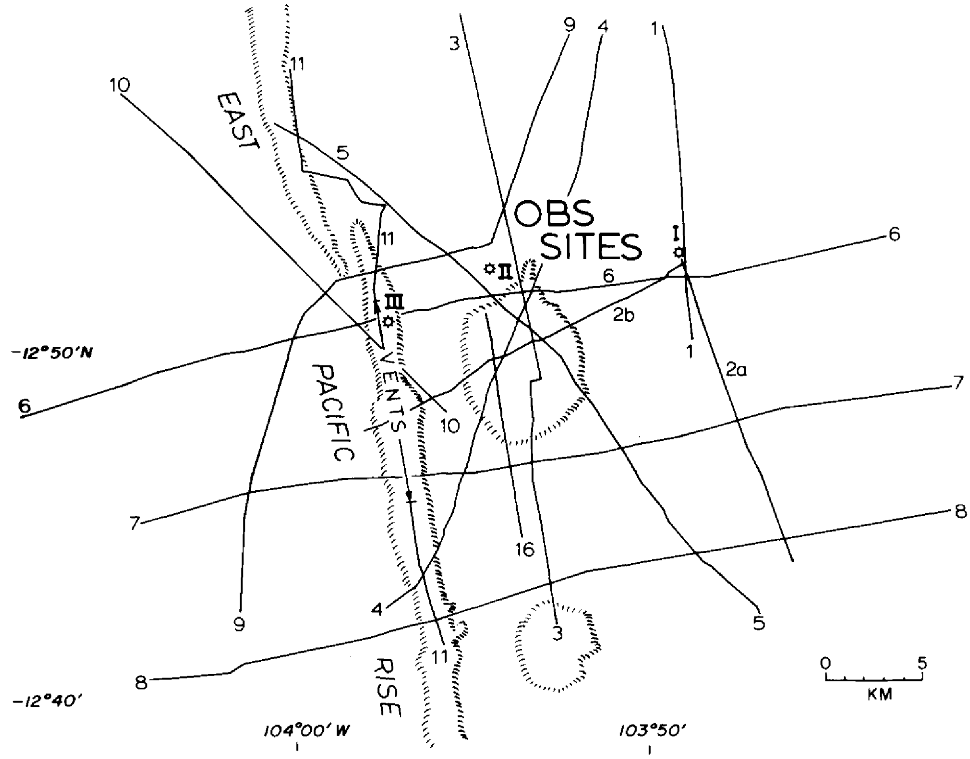

FIG. 2. Location of three OBS sites (I, II, and III) with respect to the East Pacific Rise and associated vent or 'Black Smoker' field. Seamounts are small-scale features less than 150 m in height. The conjugate-spreading centres are outlined near 12°50'N. Topography is adapted from SEABEAM chart provided by R. Hekinian and mapping conducted by R/V Melville during MAGMA expedition. Short lines up to 18 km in length consist of small shots (1–3 lb) fired every 230 m. Lines, originally intended to radiate from the OBS sites, suffer from the usual navigation problems encountered at low latitudes at sea.

crustal magma chamber for fractionation processes. In order to provide a data set capable of unambiguously constraining the depth and lateral extent of the crustal magma chamber, we conducted an extensive seismic-refraction experiment on the East Pacific Rise at 12°50'N in June and July 1982.

The East Pacific Rise near 13°N is characterized by a central high between 1 and 2 km in width. Hydrothermal vents have been observed using bottom photography within this central high between 12°45'N and 12°53'N (R. Hekinian, pers. comm. 1982) and SEABEAM profiling in the area reveals a small discontinuity in the rise axis (Fig. 2). The eastern–southern ridge dies out to the north of the western–northern branch. The overlapping nature belies a classical fracture zone, although the region is roughly in line with the O'Gorman 'Fracture Zone' described by Klitgord & Mammerickx (1982). A number of these small offsets have recently been observed along the East Pacific Rise (Macdonald & Fox 1983) and have been given the name 'conjugate-spreading centres' or 'degenerate transform faults'. The feature of 13°N is the smallest of the degenerate transform faults with an offset of less than 2.5 km. In fact, the feature was noted during the MAGMA expedition while conducting refraction profiles along the rise axis. The nature of the bathymetry was relayed to Macdonald & Fox on board the SEABEAM-equipped R/V Washington and the original French charts were extended to the north to include this sea-floor feature.

The MAGMA experiment utilized OBSs at three sites to record near-surface explosive sources. The sites were located on the ridge (age = 0), 6 km to the east (age = 0.11 Ma) and 16 km east (age = 0.30 Ma) (see Fig. 2). The OBSs were microprocessor-controlled digital instruments which record four components (vertical, two horizontals and a hydrophone) at 128 samples s^{-1} (Moore *et al.* 1981). A total of nine (eight successful) OBS deployments were made at the three sites. The single OBS to fail suffered from an erratic oscillator which caused time to

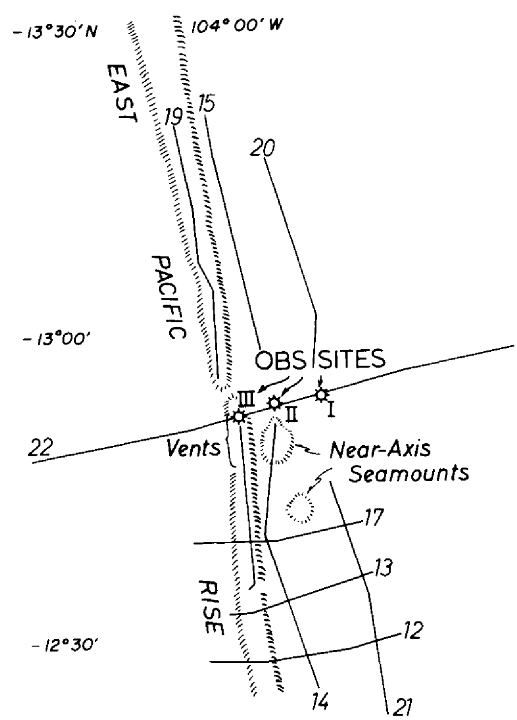

FIG. 3. Long refraction lines using conventional shots to 300 lb in size fired approximately every 0.9 km. The degenerate transform fault is sketched without the observed overlap (see Fig. 2).

advance too rapidly; the OBS was, however, successfully recovered. The redundancy of multiple OBSs at each site was required to store the large volume of data and to provide multiple coverage of the most essential lines. Not only were the OBSs turned on to record lines shot through the instruments, but OBSs at the other sites were frequently turned on to provide a broad three dimensional arrangement of sources and receivers. Each bottom seismograph was programmed to record 30 or 45 s of data at predetermined times. As many as 698 windows were used for an individual OBS, requiring that a very detailed shooting plan be worked out in advance. A total of 2003 shots were fired yielding some 4633 shot receiver pairs or 18,532 individual seismograms.

The refraction experiment, which extended over a 23-day period on station, was broken into two phases. In the first phase, short (15–17 km) lines were shot using small (1–3 lb) charges suspended from floating balloons at a depth of 10 m. This shooting technique provided a strikingly repeatable source. A grid of lines was shot in this manner (Fig. 2) with a shot spacing along each profile of about 230 m. The shots provided a dense areal coverage of the upper crust at the ridge axis and adjacent regions. The second phase of the refraction programme consisted of larger sources (15–300 lb) shot over an expanded two-dimensional grid (Fig. 3) with a lower shot density (about one shot every 900 m).

Data interpretation is in a very early stage, but a number of inferences can already be made. Profiles parallel to the ridge axis (Figs 4 and 5) exhibit dramatic differences between the ages of 0 and 0.30 Ma. The profiles over the older crust (bottom frames in Figs 4 and 5) are characterized by clear, impulsive arrivals to ranges in excess of 50 km. On the other hand, the axial profiles reveal weak emergent arrivals beyond ranges as small as

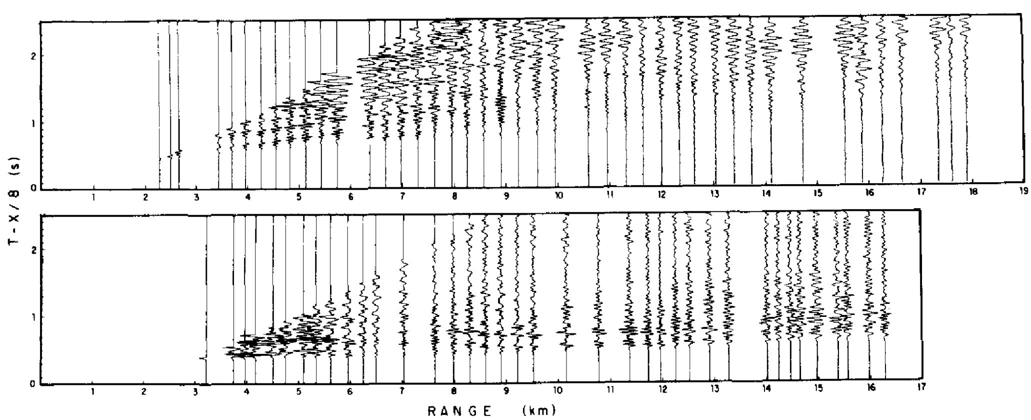

FIG. 4. Short refraction lines extending to a range of 16 km on 0.30-Ma crust and 18 km on zero-age crust. Profiles are reduced at 8 km s^{-1} and amplitudes are scaled as $(x/10)^{1.75}$. No topographic corrections have been applied. When the arrivals are impulsive the bubble pulse of the source is clearly seen. Note the extreme attenuation and delay of arrivals beyond a range of less than 10 km on the rise crest profile when compared to the generally impulsive 0.30-Ma crust.

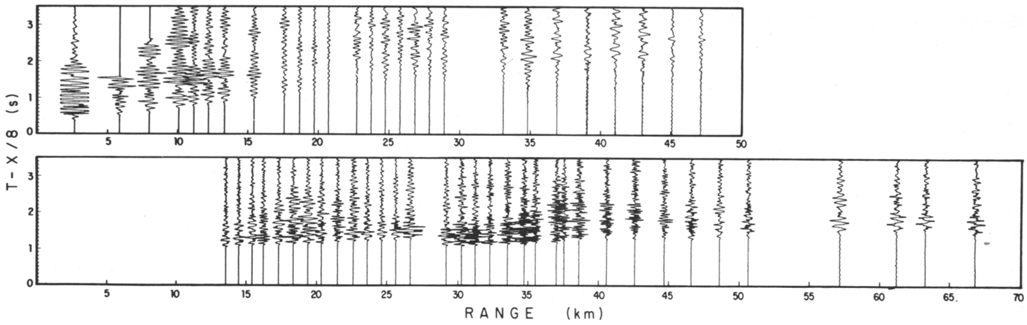

FIG. 5. Long refraction lines extending to a range of 64 km on 0.30-Ma crust and 45 km on zero-age crust. Profiles are plotted as in Fig. 4. Again note the significant attenuation of the zero-age profile vis-à-vis the profile collected on crust with an age of 0.30 Ma. The frequency content of the rise-crest line is restricted to quite low frequencies.

7 km. The contrast could not be the result of attenuation exclusively, but requires a low-velocity medium at a fairly shallow depth beneath the sea floor.

The effect of a crustal low-velocity zone is shown in Figs 6 and 7. The velocity–depth profile in Fig. 6 is the FF2 model of Spudich & Orcutt (1980). The ray-trace diagram and accompanying travel-time curve illustrate the effects of high-velocity gradients in the shallow and deep crust. Few rays turn within the fairly homogeneous 'plutonic' lower crust so we would expect the arrivals from this portion of the crust to be relatively weak. Fig. 7 illustrates the effect of introducing a shallow low-velocity zone into the structure. Now, as rays penetrate deeper than approximately 1 km into the crust, the wavefronts are refracted downward away from the sea floor and a substantial 'shadow zone' is generated. The expected effect would be a significant attenuation of seismograms beyond some cut-off distance. We propose this effect is responsible for the rapid attenuation of amplitudes in Figs 4 and 5 beyond ranges of 7 km. This extremely short range

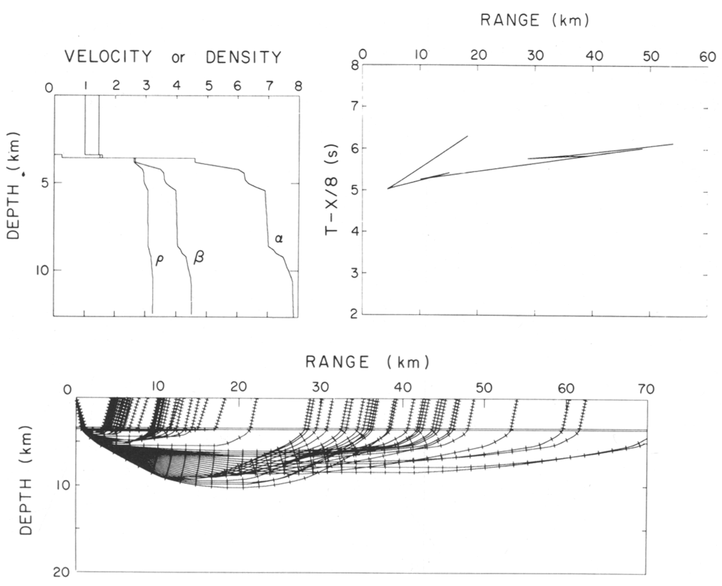

FIG. 6. Ray-trace diagram corresponding to the elastic profiles plotted in upper left-hand corner. The velocity–depth profile is adapted from the work of Spudich & Orcutt (1980). The large triplication in the travel-time curve in the upper right-hand corner is caused by the Moho transition. Note rays return to the surface along the entire length of the profile to produce profiles similar to the 0.30 Ma crustal sections in Figs 4 and 5.

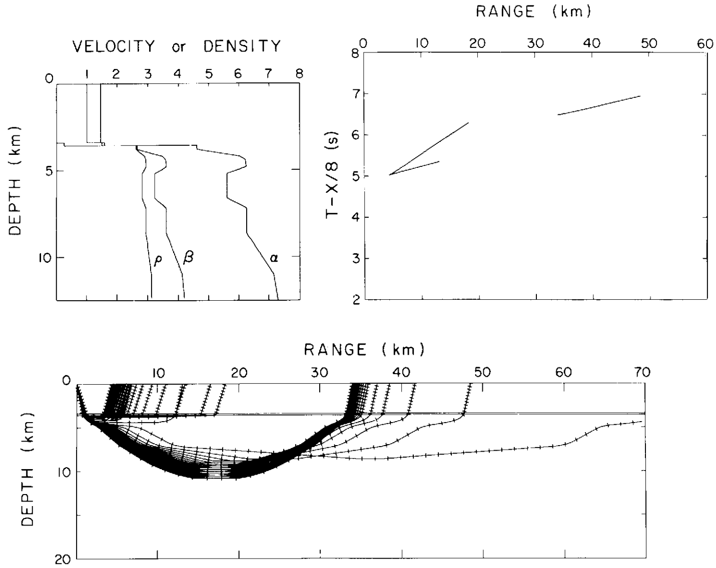

FIG. 7. Figure plotted as in Fig. 6 except the velocity–depth profile includes a pronounced low-velocity zone. The ray-trace diagram illustrates the downward refraction of rays as they touch and enter the low-velocity zone. Note the 'shadow zone' and delay introduced in the travel-time curve. The cut-off of clear, impulsive arrivals at the shadow is diagnostic of the depth to the top of the low-velocity zone.

requires that the magma chamber responsible for the low-velocity zone be very shallow in the crust, perhaps no deeper than 1, at most 2 km.

The data collected on lines parallel to the rise axis at a distance of 6 km most closely resemble the 0.30-Ma profiles in that distant arrivals are quite clear and impulsive. However, as illustrated in Fig. 8, a clear shadow zone and arrival delay are encountered at ranges on the order of 7–11 km. We currently interpret this to indicate

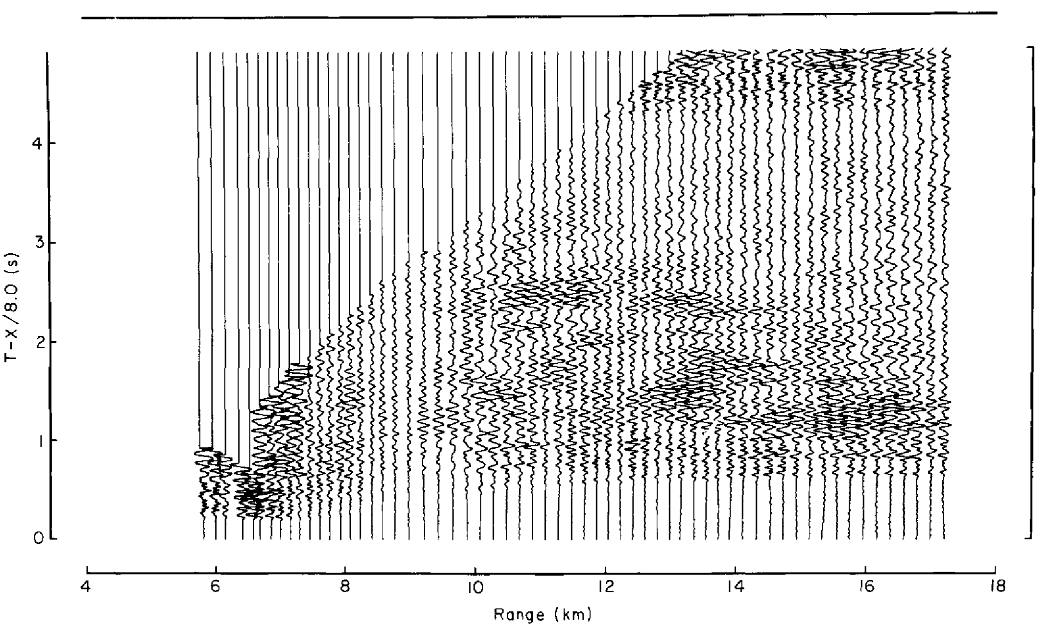

FIG. 8. Short refraction lines parallel to the rise axis on 0.11-Ma crust. Note the pronounced shadow zone appearing at a range of 8–11 km and the relatively impulsive arrivals at greater ranges.

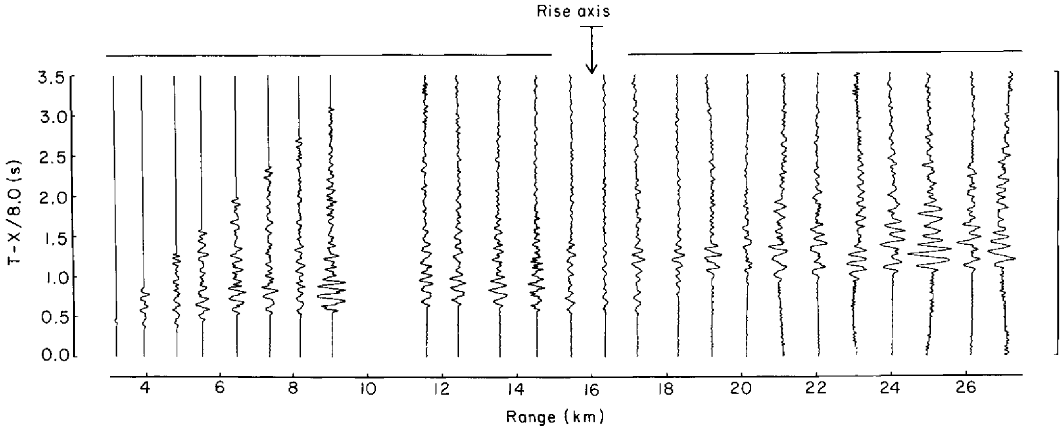

FIG. 9. Refraction line crossing rise axis as recorded on the hydrophone channel from Site I 16 km to the east. Note the time delay and change in wave form which occurs about two shots beyond the rise axis.

the solidification of the lower portion of the chamber as a function of age with a distal wing of melt remaining shallow in the crust.

Data on the cross-ridge profile, Fig. 9, indicate a much more subtle effect over the rise—i.e. a small time delay which appears to result from anomalous material in the lower crust. Examination of data from these various profiles indicates the low-velocity zone (magma chamber) is somewhat greater than 6 km in half width and morphologically resembles the model proposed by Pallister & Hopson (1981) based on observations of the Semail ophiolite in Oman.

Evolution of the oceanic crust

Several studies in the past have provided evidence that the oceanic crust evolves as a function of age. Le Pichon *et al.* (1965), Shor *et al.* (1971), Goslin *et al.* (1972), Christensen & Salisbury (1975) and Woollard (1975) conducted studies which indicated that the oceanic crust thickened regularly with age. Mechanisms proposed to explain the thickening included regular ongoing intrusion of the lower crust from below (Christensen & Salisbury 1975) and serpentinization of the uppermost mantle (Lewis 1978). The ophiolite model for the oceanic crust, on the other hand, presumes that the crust is formed at the rise axis and moves away with the underlying lithosphere. Although intraplate volcanism is widespread there is no good evidence for the regular migration of basaltic melts to the base of the crust. Substantial serpentinization is unlikely since little evidence exists for large volumes of hydrothermal circulation within the lower oceanic crust.

McClain (1981) has reviewed available evidence for crustal thickening and, after a careful winnowing of the data set employed by most of the authors cited above, concluded there is not compelling evidence for crustal thickening. He pointed out that the bulk of the thin crustal solutions at small ages were clustered north of the Mendocino fracture zone where the slow-spreading Gorda Rise has historically produced very thin oceanic crust. He eliminated these solutions from the data set asserting that any thickening pattern should persist when the remainder of the extensive data set was examined. He also eliminated stations which were directly over fracture zones, seamounts or the volcanoes of the Hawaiian Archipelago. Finally McClain pointed out that the non-tangency of the ground waves and the bottom-reflected, water-borne phases were usually corrected assuming the delay was caused by propagation through sediments. This was done (recall that many of the profiles were collected in the 1940s and 1950s before sea-floor spreading was understood) even when the stations were near the rise crests where little sediment can be found. McClain corrected his data set for this effect by recomputing crustal thicknesses for crust less than 6 m.y. of age where sedimentation rates are very low. Rather than use the assumed sediment velocity of 2.15 km s^{-1} he substituted a 'Layer-2A' velocity of 3.5 km s^{-1}. This, of course, increased the total crustal thickness. The compound effect of these very reasonable modifications was to cast considerable doubt on the hypothesis of crustal thickening beyond an age of a few million years.

Spudich & Orcutt (1980) pointed out that the compressional velocity of the sea floor acts as a notch filter for the transmission and conversion of shear waves. Figure 10 depicts the compres-

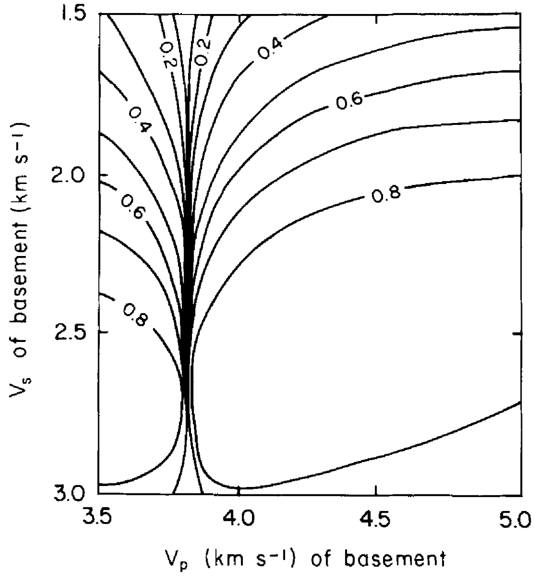

FIG. 10. Magnitude of the energy normalized PS or SP conversion coefficient at the sediment–basement interface. Phase velocity of the incident wave is 3.8 km s^{-1}, a typical crustal S-wave velocity. A basement velocity of 3.8 km s^{-1} will clearly filter out crustal shear waves with this phase velocity.

sional to shear conversion coefficient for an ocean overlying the basaltic crust where the phase velocity of the incident plane wave is taken as 3.8 km s^{-1}. The coefficient is quite large for basement velocities above and below 3.8 km s^{-1} and for high shear velocities. However, a notch appears near 3.8 km s^{-1} and broadens as the shear velocity drops. Clearly shear waves cannot exist in the crust for phase velocities near the compressional velocity of the sea floor. Spudich & Orcutt (1980) exploited this notch filter to constrain the compressional velocity of the shallow crust in this study. The compressional velocity was required to exceed 4.3 km s^{-1} and values in the range 4.4–4.6 km s^{-1} provided the best results.

We have subsequently applied this technique to the analysis of a set of Marine Seismic System (MSS) borehole data collected in the Atlantic during Leg 78B of the Deep Sea Drilling Project. These data were characterized by large amplitude shear waves and the compressional velocity at the site, on 9-Ma crust, was 4.4 km s^{-1}. Clowes & Au (1982) were also able to use the notch-filter effect to resolve high compressional velocities near 4.4–4.7 km s^{-1} at the sea floor for small crustal ages of 1–3 Ma.

Houtz & Ewing (1976) previously published an extensive treatment of sonobuoy data which outlined a systematic increase of shallow crustal velocity (Layer 2A) as a function of time. The crust near the rise axis is, statistically, typified by a velocity of near 3.3 km s^{-1}, while crust at an age of 35 Ma possesses a velocity of 4.5 km s^{-1}. The increase in velocity is attributed to a decrease in rock porosity by infilling with sediments and alteration products produced by hydrothermal circulation.

Figure 11 is modified from Fig. 8 of Houtz & Ewing (1976) where the triangles (Pacific) and squares (Atlantic) represent averaged values for crust of various ages. The straight line is a least-squares' regression of these data and indicates that the Layer-2A compressional velocity does, indeed, increase with age. However, the results from the studies of Spudich & Orcutt (1980) and Au (1981) and the MSS data, when plotted on the same figure, lie at much higher velocities. These results are inconsistent with the ageing hypothesis and serve as counter-examples to the interpretation of the sonobuoy data. Furthermore, recent ocean-crust, multichannel expanding-spread data collected during the Lamont Doherty, University of Texas, University of Rhode Island, Woods Hole Oceanographic Institution TRANSECT programme between the Mid-Atlantic Ridge and Blake-Bahama Plateau are typified by strong shear wave propagation from short ranges to ranges in excess of the mantle triplication (Peter Buhl, pers. comm. 1982). These observations will add considerably to this data set which is very sensitive to the upper-crustal velocity and will

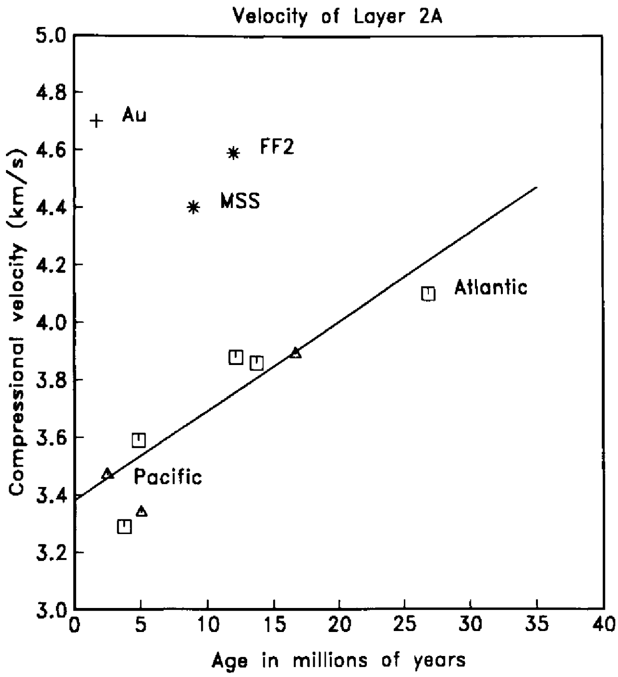

FIG. 11. Regressions of 'Layer-2A' velocities in the Pacific (triangles) and Atlantic (squares) as a function of age from Houtz & Ewing (1976). Asterisks and plus signs are taken from studies cited in the text. These latter velocities serve as well-constrained counter-examples to the velocity–age systematics proposed for 'Layer-2A'.

provide a clearer picture of the nature of the evolution of the shallow oceanic crust. Nevertheless, it has been possible to document sites in which the simple Layer-2A ageing phenomenon does not apply and the previously apparent variation of upper-crustal velocity with age may more aptly be described by regional or even local variations in crustal genesis.

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