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An Oceanic Long Range Explosion Experiment

A Preliminary Report

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Abstract. An oceanic long range explosion experiment has been conducted on well-dated, 70×10^6 year old lithosphere to a range of 600 km. The receiving instruments were an array of digital ocean bottom seismographs and chemical explosions up to 2 t in weight were used as sources. This paper reports the results of the travel-time analysis of the data from one station. The travel-time data were formally inverted using extremal and linearized inversion methods. The results indicate that considerable "fine structure" exists in the upper mantle as has been found in various continental profiles. The inversions do exclude velocities in excess of $8.4\,\mathrm{km \cdot s^{-1}}$ to a depth of 60 km.

Key words: Long line – Lithospheric profile – Explosion seismology – Oceanic lithosphere – Ocean bottom seismograph – Travel-time inversion.

Introduction

During recent years there has been considerable activity in the conduct of long range explosion seismology studies of the upper mantle. Project Early Rise (Barr, 1967; Green and Hales, 1968; Iyer et al., 1969; Lewis and Meyer, 1968; Masse, 1973; Mereu and Hunter, 1969; Warren et al., 1968) and similar experiments in the Soviet Union (Ryaboi, 1966) served as models for a subsequent series of detailed European experiments in western Europe across France (Hirn et al., 1973) and Great Britain (Bamford et al., 1976; Kaminski et al., 1976). Similar experiments in the oceans include a long line across the Gulf of Mexico to stations in Florida and Mexico (Hales et al., 1970) and a long line near the Marianas Trench (Asada and Shimamura, 1976).

Recently modern analysis methods and increasing data density have begun to reveal considerable "fine structure" in the upper 100 km of the earth's mantle. Specifically, evidence for quite high velocity material ($8.3–8.8\,\mathrm{km \cdot s^{-1}}$) has been found in this shallow depth range (Hales et al., 1970; Kosalos and

Meyer, 1975; Hirn et al., 1973, 1976; Kind, 1974; Ryaboi, 1966; Kennett, 1976; Sutton and Walker, 1972) and anisotropy has been demonstrated to play a large role in the detailed velocity structure of the upper mantle (Raitt et al., 1971; Bamford, 1973).

Several difficulties are evident in the past conduct of oceanic profiles. In the case of the Gulf of Mexico experiment the sources were at sea over oceanic lithosphere of questionable origin and the receivers were on continents. The interpretation of the data is, thus, necessarily clouded by the wave or ray modification in the ocean-continent transition. At sea, when only a few receivers (ocean bottom seismographs in the case of the Marianas experiment) are available it is necessary to expend large quantities of explosive in obtaining a few seismograms. The parsimonious distribution of data available in such a case leaves wide latitude in conducting the inverse problem of discovering the velocity at depth (Kennett and Orcutt, 1976).

In the conduct of this experiment the ocean bottom seismographs were placed on well-dated oceanic lithosphere and the line was extended parallel to magnetic anomaly 32 (generated approximately 70 mybp.). Shooting along an isochron minimizes the effects of heterogeneity due to lithosphere evolution. The line extended between the Clarion and Molokai Fracture Zones in the north-eastern Pacific to a range of nearly 600 km so that no major tectonic features were crossed. We have, thus, sought to minimize the effects of variations along the profile which have presented difficulties for previous long lines. In order to boost the data density we have expended nearly 35 t of chemical explosive within 600 km of the ocean bottom seismograph array.

Conduct of the Experiment

During the January, 1976 Scripps *Deepsonde* Expedition a long, split refraction profile was shot to a range of 600 km using charges up to 2 t in weight. The data to be examined were obtained with a closely-spaced ($\sim$1 km) linear array of three digital ocean bottom seismographs near the Clarion Fracture Zone in 5.5 km of water. The ocean bottom seismographs employed have been described in detail by Prothero (1974) and the use of these instruments for refraction has been discussed by Orcutt, Kennett and Dorman (1976). Figure 1 illustrates the data obtained between 100 and 600 km from one ocean bottom seismograph with the ocean bottom topography and free air gravity anomaly underlain. The arrows indicate the travel-time picks used in the interpretation. The amplitudes have been normalized to a shot size of 100 kg (Orcutt et al., 1976), band-pass filtered 3–9 Hz with a digital filter and, to amplify distant traces, are multiplied by a factor proportional to range. The travel times have been reduced by a velocity of 8 $\mathrm{km \cdot s^{-1}}$ and the water delay has been removed. A topographic correction has been applied assuming that the velocity contours down to a depth corresponding to the highest apparent velocity mirror the topography at the sea bed (Kennett and Orcutt, 1976). The analysis of the data to a range of 100 km has been described elsewhere (Orcutt, et al., 1976) and the velocity bounds and structure resulting from that analysis are illustrated in Figure 4

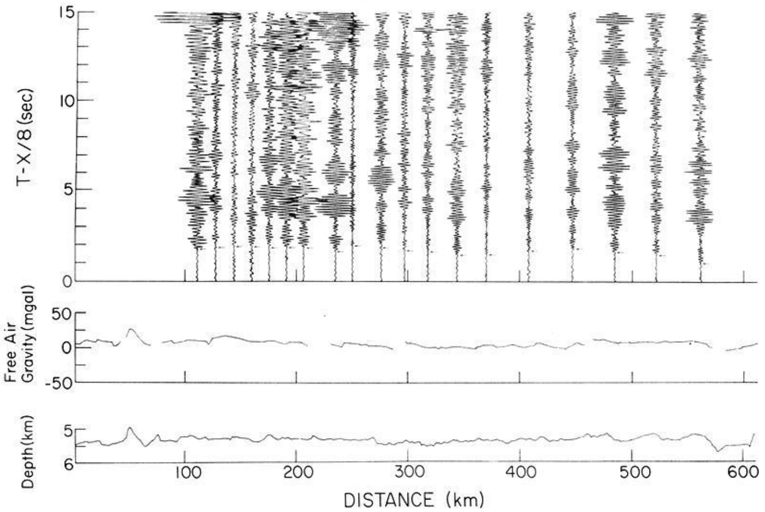
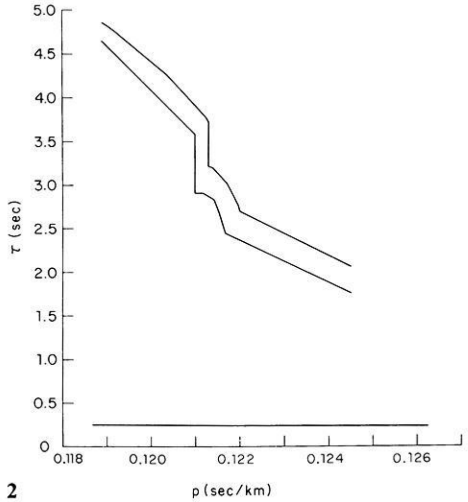


Fig. 1. Seismograms between 100 and 600 km for one of three closely-spaced ocean bottom seismographs. Travel times are reduced by 8 $\mathrm{km \cdot s^{-1}}$ and amplitudes are scaled to a single shot size of 100 kg. The data has been band-pass filtered 3–9 Hz and a topographic correction was applied. The free-air gravity anomaly in milligals and ocean bottom topography in km (depth) are also illustrated. The arrows represent the travel time picks used in the interpretation

for depths less than 10 km. One feature of the model which is immediately apparent is the well-developed Moho. No evidence for homogeneous, thick layering historically associated with the oceanic crust (Shor et al., 1970) was evident.

Data Analysis

The travel-time inversion methods employed have been previously described by Kennett and Orcutt (1976) and are based on the "τ-method" of Johnson and Gilbert (1972) and Bessonova et al. (1974). Figure 2 illustrates the result of the reparameterization (from $T(x)$ to $\tau(p)$) of the data beyond 100 km. The lowest velocity for which there is any evidence is 8.02 $\mathrm{km \cdot s^{-1}}$ (ray parameter $= p = 0.1246\ \mathrm{s \cdot km^{-1}}$) and the highest velocity is 8.4 $\mathrm{km \cdot s^{-1}}$). Because the data at ranges less than 100 km do not permit the bounds to be continued to zero delay (sea floor) 0.1 km of 1.6 $\mathrm{km \cdot s^{-1}}$ sediment and 0.3 km of 4 $\mathrm{km \cdot s^{-1}}$ basement were stripped off the data as shown by the nearly horizontal line at $\tau \sim 0.2$ s. Two important features should be noticed in the $\tau(p)$ curve. First of all, the reparameterization process indicates that the travel-time data between 250 and 410 km lie along a retrograde branch of the travel-time curve (Bessonova et al., 1974; Kennett and Orcutt, 1976) so that, in this region ($0.1212 \leqq p \leqq 0.122$) the $\tau(p)$ curve is concave downward. Also reparameterization provides evidence for a low velocity zone, a discontinuity in delay time of as much as 1 s, at a ray parameter of approximately 0.121 $\mathrm{sec \cdot km^{-1}}$. Figure 3 illustrates the rela-



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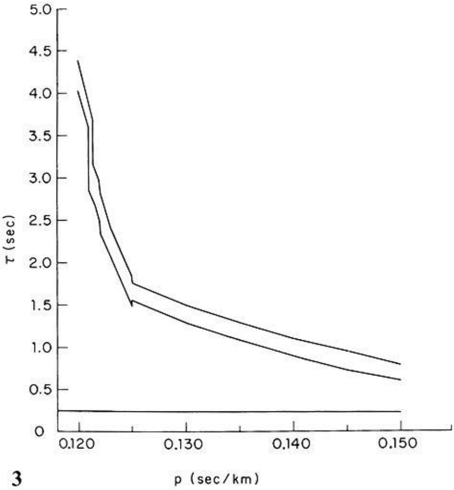

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Fig. 2. Delay time (τ)-ray parameter (*p*) data for seismograms beyond 100 km. The nearly horizontal line at the bottom represents the delay stripped from the data which is associated with very shallow (< 0.4 km) structure

Fig. 3. Delay time-ray parameter data from Figure 2 and its relationship to the data from the crustal structure

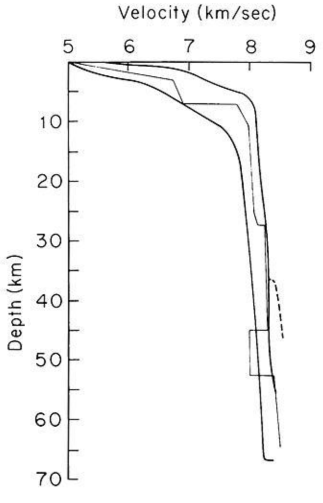

Fig. 4. Envelope in velocity depth space associated with the delay time data. The model represents the result of a linearized inversion of the delay time data and the associated travel times are plotted in Figure 5

tionship between the $\tau(p)$ curve derived for the distant data (> 100 km) and the $\tau(p)$ curve for the crustal data (Orcutt et al., 1976).

The results of inverting the $\tau(p)$ data are shown in Figure 4. The envelope was derived by an extremal inversion technique due to Bessonova et al. (1974). It was assumed that the minimum velocity within the low velocity zone was $8\ \mathrm{km \cdot s^{-1}}$ and the extremal bounds were adjusted accordingly. Because travel-time data provide only an integral constraint on the velocity structure within

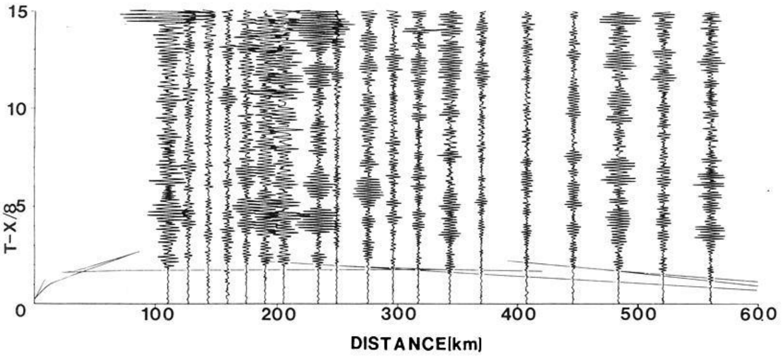

Fig. 5. The seismic data from Figure 1 with the travel time curves from the model in Figure 4 superimposed

the low-velocity zone the low velocity zone can, and does, penetrate the deeper bound. The dashed line on the shallow bound below 37 km illustrates the modified bound corresponding to the presumed wave guide minimum velocity (Bessonova et al., 1974; Kennett, 1976). Below 37 km the velocity is, thus, allowed to exceed the solid, shallow bound.

The model lying within the bounds was derived by a linearized inversion of the $\tau(p)$ data (Johnson and Gilbert, 1972; Kennett and Orcutt, 1976). At a depth of 50 km the "spread" of the resolving kernel (Kennett and Orcutt, 1976) is about 7 km and the error in velocity is 0.2 $\text{km}\cdot\text{s}^{-1}$. However, by a depth of 60 km the spread increases to over 50 km indicating that the data are providing no control of the structure at or below this depth.

The travel-time curve resulting from the linearized inversion model is superimposed upon the data in Figure 5. The travel-time curve does fit the data extremely well although the position of the caustic around 200 km is somewhat in doubt in that it should perhaps extend to shorter ranges. Because of the emergent nature of the arrivals between 100 and 200 km it will be necessary to employ synthetic seismogram techniques to work out the details of the velocity increase at a depth of approximately 28 km responsible for the triplication in the travel-time curve.

Discussion

The preliminary analysis of this profile illustrates the presence of the same sort of "fine structure" revealed by the various analyses of the recent European long lines. Formal inversions of the travel-time data alone indicate that the resolution lengths are an order of magnitude smaller than comparable resolution lengths obtainable from surface wave studies such as that of Forsyth (1975).

A very important result of the analysis of this data set is that, in this case under tightly controlled, nearly "ideal" (hopefully laterally homogeneous) conditions, velocities in excess of 8.4 $\text{km}\cdot\text{s}^{-1}$ are excluded to a depth of 60 km at this location and azimuth. The hypothesis of unusual mantle petrologies to explain high mantle velocities appears to be misplaced in the case of the oceanic lithosphere although a line perpendicular to this might reveal higher velocities due to an anisotropic upper mantle (Forsyth, 1975; Raitt et al., 1971).

During the coming year the data density to 600 km will be further increased and the length of the line will be extended to 1600 km by Scripps Institution of Oceanography in cooperation with the University of Washington and Hawaii Institute of Geophysics. Continued oceanic and continental long range profiles will help to reveal the details of the lithosphere-tectosphere-asthenosphere relationship.

The analysis thus far completed has been restricted to seismograms from a single instrument. We plan to use the array to make estimates of the apparent velocities (or ray parameter) for use in the inversion schemes and to assist in arrival identification. In addition, the phase and amplitude information will be used to further constrain the models through the mechanism of seismogram synthesis (Orcutt et al., 1976).

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