Bulletin of the Seismological Society of America, Vol. 75, No. 1, pp. 211-221, February 1985

CRUSTAL REFRACTION PROFILE OF THE LONG VALLEY CALDERA, CALIFORNIA, FROM THE JANUARY 1983 MAMMOTH LAKES EARTHQUAKE SWARM

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ABSTRACT

Seismic-refraction profiles recorded north of Mammoth Lakes, California, using earthquake sources from the January 1983 swarm complement earlier explosion refraction profiles and provide velocity information from deeper in the crust in the area of the Long Valley caldera. Eight earthquakes from a depth range of 4.9 to 8.0 km confirm the observation of basement rocks with seismic velocities ranging from 5.8 to 6.4 km/sec extending at least to depths of 20 km. The data provide further evidence for the existence of a partial melt zone beneath Long Valley caldera and constrain its geometry. This evidence is in the form of pronounced secondary arrivals, which we interpret as waves that have propagated through a volume with low *Q* (the magma chamber) and reflected from the lower boundary of the southern edge of this volume at depths of 18 to 20 km.

INTRODUCTION

The Long Valley caldera, California, is one of the most well-studied calderas in the United States. The shallow structure of the caldera and its history of eruption are known from geological and geophysical investigations (Bailey *et al.*, 1976; Hermance, 1983). Knowledge of the deeper structure comes mainly from long-range, seismic-refraction profiles (Johnson, 1965; Eaton, 1966; Prodehl, 1979) and teleseismic delay-time studies (Steeples and Iyer, 1976). Most investigators agree that one or more magma bodies exist beneath Long Valley caldera. The geophysical evidence for these magma bodies includes teleseismic delay-times (Steeples and Iyer, 1976), seismic-refraction profiles showing reflections from the top of a magma body (Hill, 1976), inflation of the resurgent dome by more than 30 cm since 1979 (Savage and Clark, 1982), and analysis of regional records of local earthquakes showing systematic attenuation of S waves for propagation paths through the caldera (Ryall and Ryall, 1981; Sanders and Ryall, 1983; Sanders, 1984). In this study, we further examine the seismic velocity structure of Long Valley caldera using local earthquakes. In particular, we attempt to constrain the lower boundary of the southern edge of the large magma body in the south-central caldera.

On 7 January 1983, an intense earthquake swarm began in southwestern Long Valley and the adjacent Sierra Nevadan block. The swarm included two events with magnitudes of 5.3 and 5.6, respectively. Over 400 events with a magnitude greater than $M = 1.5$ were recorded by the regional network during the first 3 hr (Pitt and Cockerham, 1983). During the earthquake swarm, the U.S. Geological Survey deployed 120 portable seismographs along two lines in the vicinity of Long Valley and Mono Craters of east-central California. Figure 1 shows the location of the north-northwest–south-southeast (NNW-SSE) recording line within Mono Craters, the one discussed here. A total of 25 earthquakes of magnitude greater than $M_L = 0.5$ were recorded on the line during 1 hr of recording time. Explosive sources were previously recorded along the line during a crustal structure investigation of the area in August 1982 (Meador and Hill, 1983).

This study is unusual, possibly unique, in that both earthquake and explosion sources have been recorded along the same line by a fixed, dense array of seismographs. Clear first arrivals (a high signal-to-noise ratio) were recorded for both the earthquake and seismic-refraction profiles. In addition, the earthquake data contain S-wave arrivals and a clear secondary arrival not found in the explosion data. We, therefore, have the opportunity to compare seismograms from instruments at common recording locations along a dense, linear array for surface and buried sources.

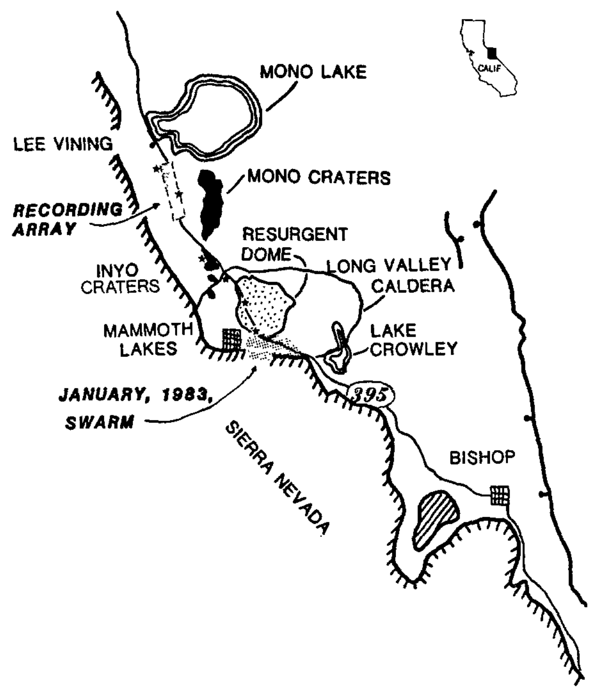

FIG. 1. Location map showing the Long Valley caldera region. The epicentral region of the January 1983 swarm and the recording array used in this study are shown. Stars indicate shotpoints for a previous explosion refraction profile recorded, in part, at the same locations.

Geologic Setting

The Long Valley caldera is an elliptical depression at the eastern margin of the Sierra Nevada 30 km south of Mono Lake. The caldera formed from the collapse of the Long Valley magma chamber following the explosive eruption of the Bishop Tuff 0.7 m.y. ago (Bailey *et al.*, 1976). The caldera walls are well defined on all but the southeast side, with relief as great as 1200 m. The caldera floor is flat in the east, but broken by a group of faulted and dissected hills in the west-central section that forms a resurgent dome (Figure 2). The regional morphology is dominated by the eastern escarpment of the Sierra Nevada range on the west, the White Mountains on the east, the Long Valley caldera, and the Mono-Inyo Craters chain of eruptive centers that trends north-south from the western edge of Long Valley to the southern shore of Mono Lake (Figure 1).

The basement rocks in the immediate area of Long Valley are Jurassic and Cretaceous granodiorites and granites of the Sierra Nevada batholith, and Paleozoic and Mesozoic metamorphic rocks of the Mount Morrison and Ritter Range roof pendants. Overlying the basement rocks are late Tertiary and Quaternary volcanic rocks, mainly basalt, andesite and rhyodacite (Bailey *et al.*, 1976). These volcanic rocks include the Bishop Tuff, a voluminous rhyolite ash flow with a thickness in Long Valley of 1000 to 1500 m (Hill, 1976), which provides a low-density fill for the caldera. Although it is identifiable over about 1100 km^2 in eastern California (Gilbert, 1938), the Bishop Tuff is not exposed in Long Valley as subsequent eruptions have buried it within the caldera.

Regional seismic-refraction studies provide a coarse view of the crustal structure in the area of eastern California and western Nevada (Johnson, 1965; Eaton, 1966; Prodehl, 1979). Since the Long Valley caldera is situated on the boundary between the Sierra Nevada, with a crustal thickness of 50 km, and the Basin and Range Province where the crust is 30 to 35 km thick, it occupies a region of pronounced

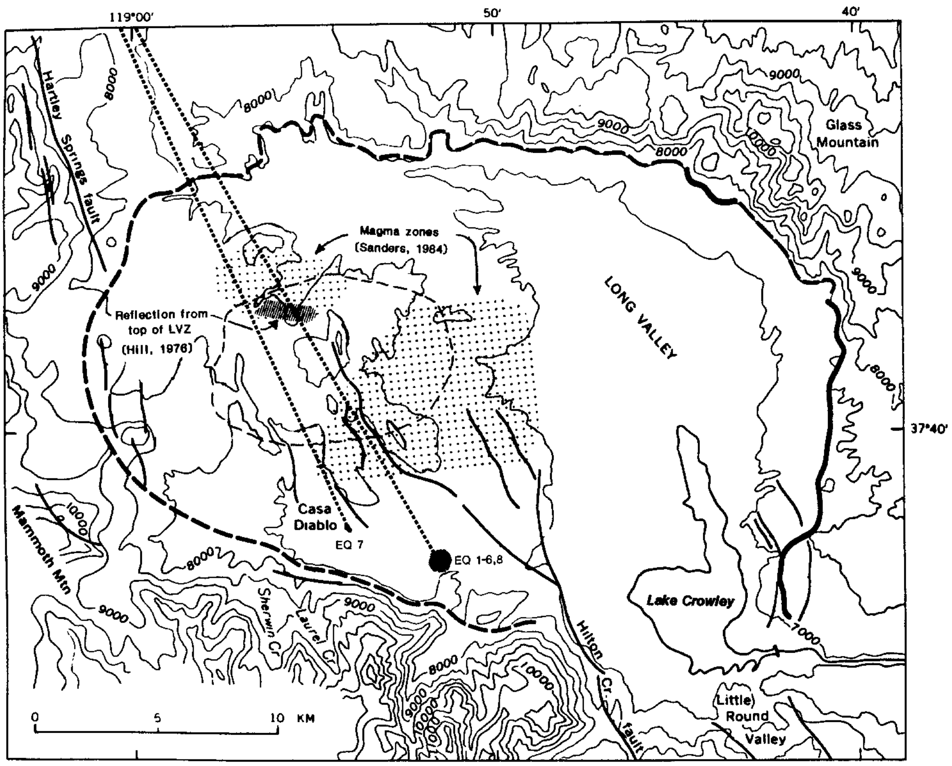

FIG. 2. Reconnaissance map of Long Valley caldera. The approximate area of the resurgent dome is indicated by a small dashed ellipse. Dotted areas are those which exhibit *S*-wave attenuation for events recorded at regional stations (Sanders, 1984). The *hatchured region* represents the location of an observed reflection from the top of a partial melt zone (Hill, 1976). Dotted lines are the surface projection of ray paths between the events recorded and the recording array. LVZ, low-velocity zone; EQ, earthquake.

lateral change in crustal thickness. Seismic-refraction studies show that the crust in the Long Valley area consists of three layers. Below a layer of surficial volcanic and fractured metamorphic rocks is a layer in which the *P*-wave velocity increases gradually from 6.0 to 6.2 km/sec (at a depth of 2 to 4 km) to 6.4 km/sec at a depth of 25 to 30 km. Velocities of 6.8 to 7.2 km/sec are measured in the lower 10 to 20-km-thick layer of the crust. Upper-mantle velocities beneath the crust are in the range 7.8 to 7.9 km/sec, typical of those beneath the Basin and Range.

In addition to the regional seismic-refraction studies, three high-resolution seismic experiments concentrated on the Long Valley-Mono Craters area and bear on the question of the existence of a magma chamber. Hill's (1976) two perpendicular refraction profiles crossing west-central Long Valley caldera established the thickness of the Bishop Tuff within the caldera, defined formation velocities, and identified a secondary arrival as a possible reflection from the top of a low-velocity zone (presumably a magma chamber) at 7 to 8-km depth.

In a study of relative teleseismic *P*-wave arrival times, Steeples and Iyer (1976) observed systematic travel-time delays for ray paths traversing the west-central part of the Long Valley caldera. These delays were explained by a spherical region of low-velocity material (15 per cent velocity decrease) with a 7 km radius extending over a depth of 8 to 22 km beneath the caldera.

Ryall and Ryall (1981), Sanders and Ryall (1983), and Sanders (1984) interpreted data from a regional seismic network operated by the University of Nevada to indicate an attenuation of *S* waves for shallow earthquakes with epicenters in the area of the southeast boundary of the Long Valley caldera when observed at stations to the northwest, north, northeast and east of the caldera. They conclude that this attenuation pattern is the result of propagation through magma chambers beneath the western part of Long Valley at depths greater than 7 to 8 km. Thus, three previous seismological studies support the existence of a zone of partial melting in the mid-crust beneath the Long Valley caldera.

Field Procedure and Data Reduction

Our strategy in this experiment was to take advantage of the high seismicity level during the January 1983 swarm to record earthquake profiles in Long Valley to augment the explosion-source profiles recorded in the summer of 1982. Data recording began on 9 January using 120 portable seismic recorders. These recorders, which normally are used for explosion refraction profiles, do not have an "event-trigger" capability. They were programmed to simultaneously record for 30-min recording windows. By turning the cassette tapes over between recording windows, a total of 1 hr of recording was obtained on a specific deployment line. The sensor is a single vertical-component velocity transducer with a natural resonance frequency of 2 Hz. The data are recorded on three channels, each with a different, preset amplifier gain (Healy *et al.*, 1982).

Recorder locations were determined using orthophotographic quadrangle maps, topographic maps, and by distance measurements using precision odometers. Recorder spacing was 160 m, with an estimated location accuracy of ±5 m.

Seismic record sections were prepared by digitizing the analog tapes at 200 samples/sec. Subzero nighttime temperatures resulted in a greater than normal number of instrument malfunctions, and only 68 of the recorders provided final digital records on the Mono Craters deployment used in this study.

The eight earthquakes used in this study (Table 1 and Figure 2) were selected solely on the basis of being the largest aftershocks recorded during the two 30-min recording periods. All were located beneath the south moat of the caldera at depths between 5 and 8 km. The epicenters of events 1 through 6 and 8 are located in a tight cluster; event 7 is located 4 km northwest of the others.

Critical to interpreting velocity structure of the deeper crust using data at relatively short source-receiver ranges (less than 50 km) is the certainty with which secondary arrival branches may be identified. To dependably identify secondary arrival phases, the sampling interval should be less than a wavelength. To enhance the chances of success, we used close spatial sampling (160 m spacing) both to provide a measure of appropriate sampling density for future work and to compensate for records lost due to expected cultural noise and instrument failures caused by the freezing temperatures. Also, although we were using only vertical geophones, we hoped to be able to identify the S-wave arrivals as a basis for gaining some information about the shear-wave velocities of the crust.

Observations and Interpretations

A seismic-refraction profile recorded by the U.S. Geological Survey in the summer of 1982 in the Long Valley-Mono Craters-Inyo Craters area is of particular relevance to our study (Meador and Hill, 1983). The profile, with multiple shotpoints extending 40 km from west-central Long Valley to Mono Lake, provides a detailed look at the transition across the northern caldera wall and at the upper crust in the Mono Craters area. Many of its recording sites were reoccupied during the earthquake profiling.

Preliminary interpretation of the 1982 seismic-refraction data led to the upper crustal velocity structure used in modeling the earthquake profiles shown in Figure 5 (D. P. Hill, written communication, 1983; Hill *et al.*, 1985). Under this interpretation, the seismic velocity of basement rocks is approximately 5.8 km/sec at a depth of 2.5 km. The absence of strong secondary reflected arrivals in the refraction profile from shotpoint 1 (Figure 3) indicates a lack of strong (horizontally continuous) velocity discontinuities within the upper 10 to 15 km of the basement beneath the central section of the profile.

TABLE 1

Events Recorded on 11 January 1983

Event	Time	Latitude	Longitude	Depth (km)
1	06:00:14.22	37°36.62′	118°51.78′	6.3
2	06:04:14.79	37°36.78′	118°52.57′	5.5
3	06:21:43.57	37°36.42′	118°52.27′	4.9
4	06:23:23.21	37°36.63′	118°52.42′	5.7
5	06:26:12.43	37°36.96′	118°52.96′	6.1
6	10:01:36.54	37°37.30′	118°51.62′	4.9
7	10:11:16.25	37°37.64′	118°54.79′	6.7
8	10:26:08.02	37°36.75′	118°52.26′	8.0

Ray trace modeling of the explosion data from the profile of the summer 1982 experiment has shown that the horizontally refracted arrivals reached a maximum depth of penetration of about 4 km beneath and north of the northern caldera wall. The earthquake sources, which occurred at depths of 5 to 8 km, provide depths of penetration within the caldera which could only be obtained with longer profile lines and larger surface sources. Thus, the earthquake sources provide an opportunity to extend to greater depth the crustal velocity model determined from the explosion profile.

Figure 3 is a comparison of record sections recorded from shotpoint 1 in the summer of 1982 with that from earthquake 4 recorded 11 January 1983. The traces are aligned for direct comparison; reoccupied sites line up. The recording sites, shotpoint 1, and earthquake 4 are roughly colinear, with the earthquake being at approximately 5 km greater range. The first arrival branch, identified as P_g in both cases, is easily picked on both record sections and has a slightly higher apparent velocity on the earthquake record section, as is expected for a buried source in the presence of a small positive velocity gradient. The lack of identifiable secondary arrivals on all the explosion record sections stands in contrast to the clear S arrival, and a clear secondary arrival falling 2.8 to 3.0 sec later than the first arrival on the earthquake record sections (Figures 3 and 4). This secondary arrival will be the focus of our interpretation below.

The P_g and S_g arrivals were analyzed using ray tracing techniques based on a simple model with the shallow structure (0 to 4 km) constrained by the 1982 explosion data (Figure 5). The result was a model in which P-wave velocities of 5.8 km/sec at the top of the basement at 1. to 2.5 km depth increase with a vertical gradient of 0.05 sec^{-1} to a velocity of 6.1 km/sec at 6 km, the median depth of the earthquakes. This model is, of course, nonunique because these single-ended earthquake profiles do not provide resolution of lateral variation in basement velocity at depth.

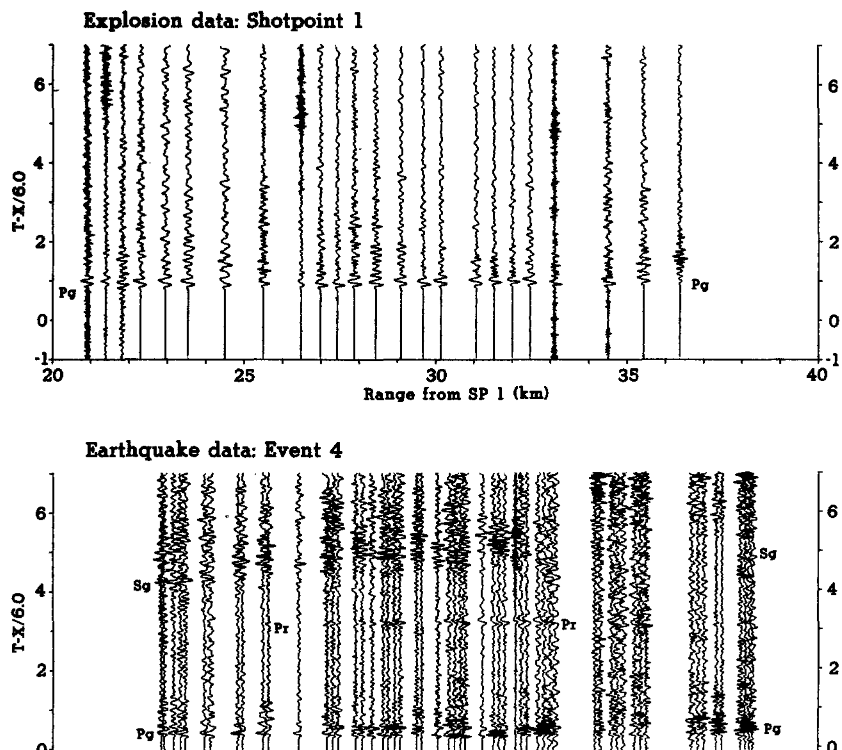

FIG. 3. Profile data recorded from shotpoint 1 (August 1982) and earthquake 4 (January 1983; source depth = 5.7 km). Data are recorded along the same profile line with many stations reoccupied. Record sections are positioned with traces from reoccupied sites aligned although the earthquake is at 5 km greater range. P_g is readily identified on both sections, but S_g and P_r appear only on the earthquake section.

The S_g arrivals were modeled by simply converting the P-velocity model into an S-velocity model using a uniform V_p/V_s ratio. We found that the best fit was obtained for a "normal" V_p/V_s ratio of 1.7. This implies that for the region of the upper crust sampled by the direct arrivals (depths less than 6 km), the major portion of each ray path passes through rocks with normal P and S velocities, and that there is little or no partial melt shallower than about 6 km in the area of western Long Valley sampled. The fact that we have recorded S_g at all argues against partial melt along the P_g/S_g ray path, a result consistent with that of Ryall and Ryall

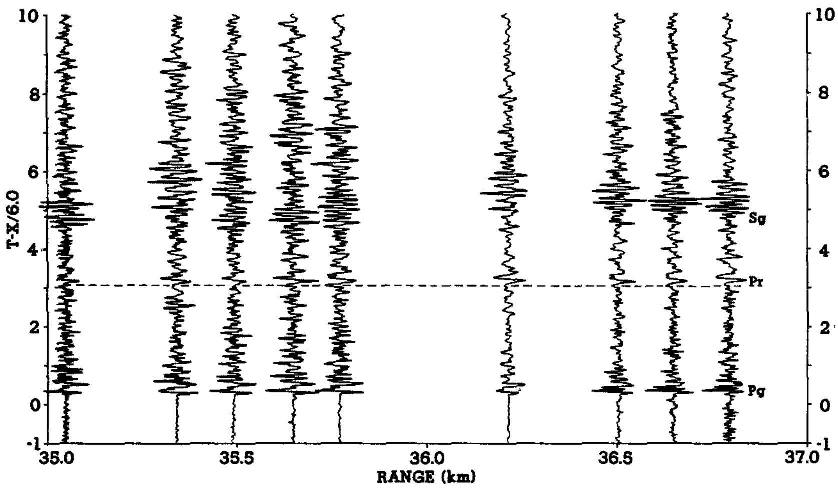

FIG. 4. Detail of record section from earthquake 4 showing arrivals in the distance range 35 to 37 km. The P_r phase has amplitude comparable to that observed for the P_g and S_g phases, but the frequency content of the arrivals is quite different.

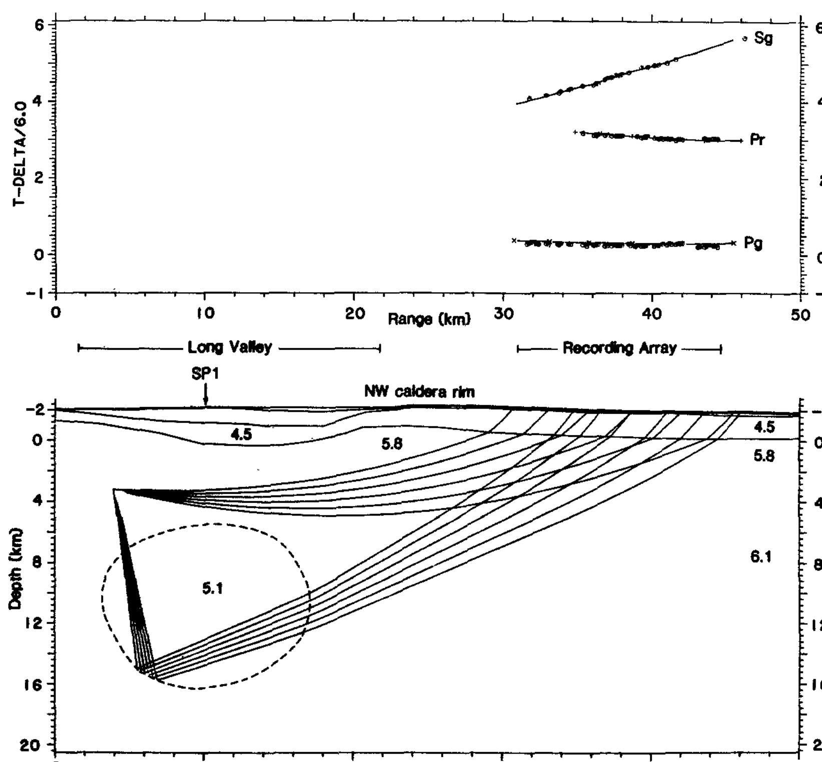

FIG. 5. Ray trace model used to model the arrival times observed from earthquake 4. The velocity structure above the basement is from a preliminary interpretation of explosion refraction profiles and should not be considered definitive (D. P. Hill, written communication, 1983; c.f., Hill *et al.*, 1985). The observed P_g and S_g arrival branches are modeled by the rays with near-horizontal take-off angles. The P_r branch is attributed to rays traveling through and reflecting from beneath a low-velocity zone of limited extent beneath Long Valley caldera. The dimensions and velocity contrast of the low-velocity zone are in agreement with the model of Steeples and Iyer (1976) from teleseismic delays.

(1981) and Sanders (1984), who studied seismograms from the 1980 Mammoth Lakes earthquakes recorded at regional stations.

Perhaps the most intriguing feature observed in this data set is the clear secondary arrival following the P_g arrivals by about 2.8 sec on most of the record sections (Figures 3 and 4). Seven of the eight earthquakes used in this analysis cluster within a small epicentral area in south-central Long Valley. Record sections from these seven events all show this phase; event 4, used as an example in Figures 3 and 4, is representative of these record sections. The eighth event, whose epicenter is 4 km northwest of the others, lacks the later arrival. For many seismograms, the secondary arrival (henceforth referred to as P_r) has an amplitude approximately equal to that of the P_g arrival. A number of observations must be taken into account in efforts to model this phase.

1. The P_r arrivals are not observed over the full range of the recording array. This implies that they result from either a focusing effect due to curved boundaries or velocity gradients or that they are cut off at either end of the recording array by a shadowing effect due to, say, limited extent of a reflector.
2. The arrival times of the P_r phase fall 2.8 to 3.0 sec later than the P_g arrivals, but well ahead of the S_g arrivals. This implies the path for the P_r phase must be a P-wave path for a large part, if not all of its length.
3. The apparent velocity of the P_r branch is 6.3 km/sec, slightly greater than that of the P_g arrival branch (6.1 km/sec). This implies that the P_r rays travel the latter part of their paths with P velocities and along ray paths similar to, but somewhat steeper than, those of the direct P arrivals.
4. The dominant frequency of the P_r arrivals is significantly lower than that of the P_g arrivals. This implies that the rays pass through a low-Q region not sampled by the direct arrivals.
5. The failure to observe the P_r phase on seismograms from earthquake 7, which is displaced 4 km from the cluster of other events that show the P_r phase, suggests that the body responsible for the P_r phase has a spatially limited geometry.

A succession of crustal velocity models were tested and rejected because they contradicted one or more of the above observations. For example, modeling these secondary arrivals with a simple horizontal reflector at depths sufficient to give the observed time delay yields precritical reflections at these ranges which have apparent velocities substantially greater than those of the observed P_r phase. An attempt to match both the apparent velocity and the time delay by using a simple dipping reflector at depth within a basement having no large lateral velocity variations was unsuccessful. Positioning the reflector at depths shallow enough to provide the observed apparent velocity does not provide an adequate time delay. Several models were attempted in which we sought to increase the travel time by initially propagating the P_r phase as S waves. By using $S - P$ conversions at depth, we found that although the P_r branch could be modeled, additional reflected phases ($P - P$, $P - S$) were predicted for times where none are observed. A model that might be expected to satisfy the amplitudes and high-frequency attenuation observed in the data is one in which the P_r rays start as upward propagating S waves, undergo $S - P$ conversion at one of the high impedance contrast near-surface interfaces, reflect from the surface, and complete their ray paths with P velocities. Although large-amplitude secondary phases are predicted by such a model, the second reflected legs of the ray paths are similar to those taken by the explosion sources, and the apparent velocities, like those from the explosion profile data, are always less than those of the observed earthquake P_g and P_r arrival branches. In addition, if this were an appropriate model, a difference in source location of 4 km is unlikely to account for the disappearance of the P_r phase as observed for earthquake 7.

In view of the strong evidence for a partially molten zone beneath Long Valley, our preferred model is one in which the P_r phase is the result of propagation through and reflection from below a low-velocity zone (Figure 5). We obtained this model by first selecting a depth and dip for the reflector that would produce the observed apparent velocity for the P_r phase. We then adjusted the *P*-wave velocity of the volume between the earthquakes and the reflector to give the proper delay. The resultant velocity, 5.1 km/sec is 16 per cent less than the velocity of unaltered basement rocks here. This model accounts for both the time delay and apparent velocity of the P_r phase; it also provides a low-*Q* zone to account for the difference in dominant frequency observed. Ryall and Ryall (1981) noted the attenuation of frequencies higher than 2 to 3 Hz in *P*-waves recorded at regional stations and attributed this to ray paths through a magma body beneath Long Valley. The existence of a low-velocity zone of limited extent also helps to explain both the apparently limited range over which the P_r arrival branch is observed and the correlation between location of earthquakes and observation of the P_r phase. A reflecting boundary of limited extent associated with the lower boundary of the southern edge of the magma zone requires a particular source-receiver geometry to be observed. Comparison of the surface projection of the raypaths taken with the inferred surface expression of the magma zones (Figure 2) suggests that energy from events 1 to 6, and 8 traverses a greater portion of the magma zones than that of event 7.

As a test of the feasibility of this model, simple two-dimensional ray method synthetic seismograms were calculated for the P_g and P_r phases. We found that, although comparable amplitudes could be obtained, they relied upon numerous factors over which we have no constraints (i.e., source radiation pattern, focussing/defocussing effects of the reflecting interface and the low velocity region). Thus, we can not use amplitude considerations as proof of the model and emphasize that this model is largely based on the modeling of travel times.

The observed data require only that the reflecting boundary lie beneath the low-velocity zone. This model does not require the reflecting boundary to represent the lower edge of the low-velocity zone, although that seems to be a likely candidate for a velocity contrast sufficient to give a strong reflection. Nor does the reflection point as modeled necessarily represent the deepest extent of the magma body. It is likely that we are observing a lower boundary of the southern edge of the magma body. Nonetheless, the dimensions and velocity contrast of the modeled low-velocity zone are concordant with the preferred model of Steeples and Iyer (1976) from observations of teleseismic delays, and the apparent attenuation of high frequencies along the P_r ray path is strong evidence that the rays have traversed a partial melt zone beneath the Long Valley caldera.

Summary and Discussion

Explosion seismic investigations in the Long Valley-Mono Craters area define the velocity structure to a depth of 4 km using refracted arrivals. Later reflected arrivals have been used to identify a localized reflector at 7 to 8-km depth in west-central Long Valley (Hill, 1976). The failure to identify any further reflected phases in the explosion data implies that there are no other velocity discontinuities in the basement to depths of at least 10 to 15 km. The use of earthquakes as sources for refraction profiling allows us to confirm this implication to depths of 6 to 8 km and to suggest that, due to the lack of observed secondary arrivals other than the P_r phase, there are no major horizontal velocity discontinuities in the crust to a depth of at least 20 km. This is in accord with the regional crustal structure earlier determined by much broader surveys (Prodehl, 1979).

Previous geophysical studies have established that magma bodies exist beneath the Long Valley caldera. The depth to the top of the magma bodies is estimated variously at 4.5 to 7 km based on local earthquake data (Sanders and Ryall, 1983; Sanders, 1984) and seismic-refraction data (Hill, 1976; Hill *et al.*, 1985). Teleseismic earthquake travel-time delays indicate that the magma bodies occupy a depth range of approximately 8 to 22 km in the upper crust, but these measurements have not provided a precise estimate of the vertical extent of the bodies. In the present study, earthquake profiles are interpreted to record a deep crustal reflection from beneath a magma body in the south-central caldera. The reflection point is at 18 to 20-km depth. If we assume that this corresponds to the lower boundary of the southern edge of the magma body, we may roughly estimate its volume. The geophysical data for the magma body in the south-central caldera indicate that it has dimensions of 10 km by 5 km in the horizontal plane (Sanders and Ryall, 1983; Sanders, 1984) and a height of about 12 km (this study). Thus, the total volume of the partial melt zone is on the order of 600 km^3.

This experiment emphasizes the valuable information that can be gained by the use of earthquake aftershock sequences as sources for refraction profiling. Two obvious advantages are the enhanced penetration depth for a given recording range provided by buried sources and the shear-wave velocity information. The success of this particular experiment is enhanced by the detailed near-surface velocity and structural control provided by the previous explosion profile and the accurate timing and location of earthquakes provided by the U.S. Geological Survey seismic network stations in the area.

It is also enhanced by our ability to record arrivals on many stations at close intervals. This capability is crucial to the proper identification of phases and the determination of mean apparent velocity over distances of several wavelengths.

Acknowledgments

Field data were collected by J. M. Murphy, V. D. Sutton, P. J. Meador, and R. H. Colburn; R. W. McClearn and S. K. Gallanthine provided technical support in the field. Support from the U.S. Geological Survey Volcano Hazards Program and the Department of Energy is gratefully acknowledged. Our ideas benefited from discussion with H. M. Iyer and D. P. Hill. Comments on the manuscript by J. Mills greatly improved the text.

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Manuscript received 29 March 1984