JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 104, NO. B2, PAGES 2895–2908, FEBRUARY 10, 1999

Perturbation to the lithosphere along the hotspot track of La Réunion from an offshore-onshore seismic transect

Josep Gallart,[1] Lynda Driad,[2] Phillippe Charvis,[3] Martine Sapin,[2] Alfred Hirn,[2] Jordi Diaz,[1] Béatrice de Voogd,[4] and Maria Sachpazi[5]

Abstract. A 250 km long NE-SW lithospheric transect spanning the 40 km wide island of La Réunion and its submarine edifice is derived from lines of air gun shots at sea on either side, along the assumed hotspot trace. Seismic records were obtained from an array spanning the whole transect and including sea bottom and land receivers, providing a system of reversed and overlapping observations. Low seismic velocity, and hence density, is found on average for the whole edifice above the oceanic plate. We attribute high-velocity anomalies within the edifice to an intrusive core confined under the central northern quarter of the island-crossing segment. Unexpectedly, the main seismic interfaces, top and bottom of the prevolcanic crust, do not show significant flexural downwarping under the island. In addition, clear multipathing in the recorded wave field requires the presence of a body with a seismic velocity intermediate between the prevolcanic crustal material and the normal mantle. This lithospheric structure provides the first example where underplating occurs beneath an active volcanic island, suggesting a genetic relationship. The underplated body could represent residues of the evolution of primary picritic melts that yield erupted basalts. Evidence for reflectors deeper in the lithosphere may indicate further related heterogeneity. In the plate/hotspot model commonly assumed, the structural variation along the transect could be interpreted as a variation with time of the amount and physical state of underplated material.

1. Introduction

The oceanic island of La Réunion is the site of active basaltic volcanism at Piton de la Fournaise which occupies its southeastern third. It is located 800 km east of Madagascar (Figure 1) on 67 m.y. old oceanic crust [*Schlich*, 1982; *Schlich et al.*, 1990]. It is classically considered [e.g., *Duncan*, 1981] as the most recent expression of the activity of the hotspot which formed the Deccan Trapps at the Cretaceous-Tertiary boundary and subsequently the line of submarine topographic highs of the Mascarene Plateau and of Mauritius Island. The islands of La Réunion and Hawaii are both at the end of a chain of islands or submarine topographic highs and have similar volcanism. Oceanic island volcanism is broadly understood in the hotspot frame of deep mantle plume upwelling. However, the modification to the physical state of the lithosphere, the amount of plume production, and the proportion that is erupted with respect to that possibly intruded at depth are far from well documented.

Artificial source seismology (refraction and reflection seismics) provides an appropriate tool for a quantitative resolution of both horizontal and vertical variations in the structure of the crust and uppermost mantle. Previous marine surveys comprising seismics have traversed other hotspot archipelagos, providing information on the long-wavelength modification of the structure and topography of the lithosphere associated with the volcanic production and its loading of the plate. For the Hawaiian islands, a marine survey comprising multichannel reflection seismic (MCS) profiles and a number of expanding spread profiles crossed the island chain near the island of Oahu, extinct since 2.6 m.y. [*Watts et al.*, 1985]. The structure of the islands themselves was approached by separate surveys which centered on the presently active island of Hawaii [*Hill and Zucca*, 1987]. For the Marquesas, MCS [*Wolfe et al.*, 1994] and coincident ocean bottom seismometers (OBS) refraction observations [*Caress et al.*, 1995] sampled the submarine part of a now inactive hotspot archipelago.

In the survey described here, the main target was more specifically to obtain data on the structural framework of active volcanism. Hence the survey was focused around the recent island of La Réunion and its active volcano Piton de la Fournaise. In this paper a particular objective was to sound the structure of the lithosphere at depth along a transect, for the signature of present magma genesis and transport in relation to eruptivity. A more regional approach from various transects is developed by *Charvis et al.* [this issue]. Another aim was to tie the structure beneath the island itself with its broad submarine apron, which is more specifically discussed by *de Voogd et al.* [this issue]. With these basic targets, the "Fournaseis" project was sponsored by the European Union, Natural Hazards Program and the Programme National Risques Naturels (PNRN) of the French Institut National des Sciences de l'Univers–Centre National de la Recherche Scientifique

[1]Institut de Ciencies de la Terra, Consejo Superior de Investigaciones Cientificas, Barcelona, Spain.
[2]Laboratoire de Sismologie Expérimentale, UA 195 CNRS and Observatoires Volcanologiques, Paris.
[3]Unité Mixte de Recherche Géosciences Azur, Institut Français de Recherche Scientifique pour le Développement en Coopération, Villefranche-sur-Mer, France.
[4]Laboratoire de Géophysique, Université de Pau et des Pays de l'Adour, Pau, France.
[5]National Observatory Athens, Seismological Institute, Athens, Greece.

Paper number 98JB02840.
0148-0227/99/98JB-02840$09.00

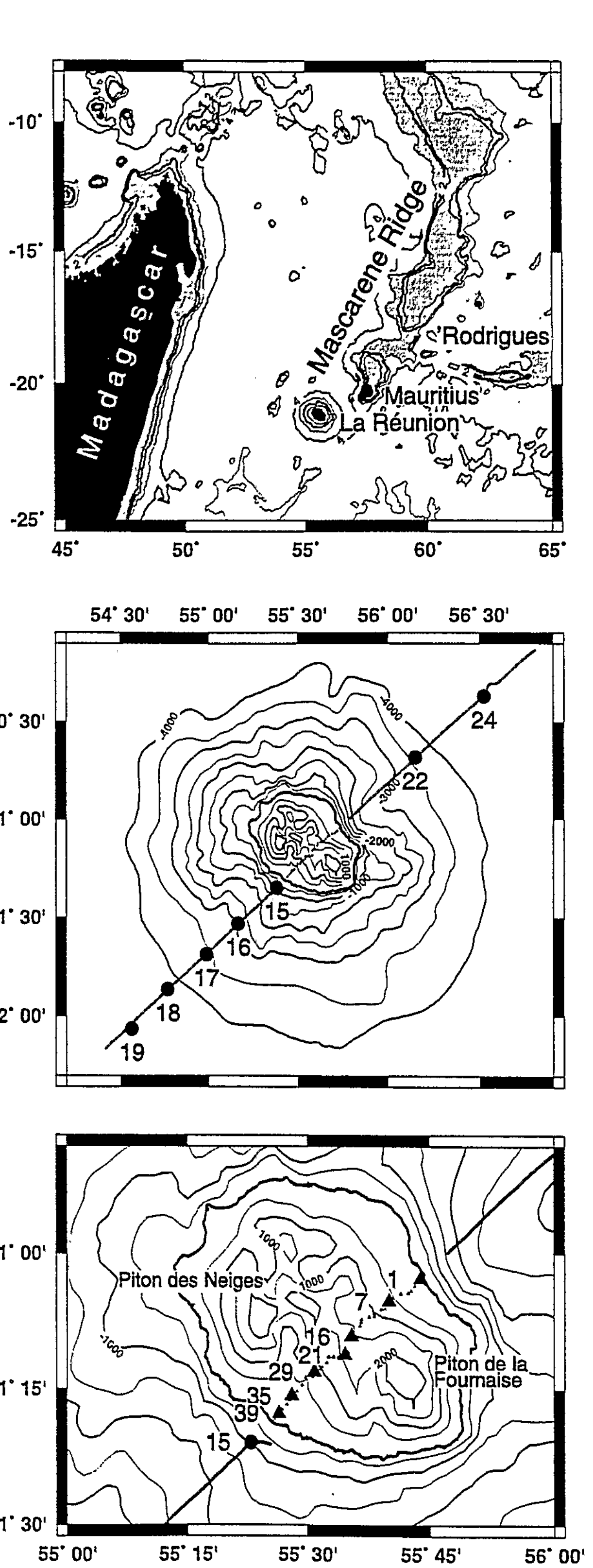


Figure 1. Location of La Réunion in the Indian Ocean, sea bottom shallower than 2 km in gray. (top) Topographic highs to the NE of La Réunion, related to past hotspot activity. (middle) The line of shots at sea of the NE-SW seismic transect across La Réunion with position of OBS recorders at sea. (bottom) Detail of land station array across the island, with identification of the discussed stations.

(INSU–CNRS). Taking advantage of the location of the volcano within a marine environment, we could use the R/V *Marion-Dufresne* of the Institut Français pour la Recherche et la Technologie Polaires (IFRTP) (MD-76 cruise "Reusis") and the marine seismic source of eight air guns of 16 L each, shot at 200 m intervals as well as the 96-channel streamer of Institut Français pour la Recherche et l'Exploitation de la Mer (IFREMER) and OBS and land seismometers (LS) from our institutions.

Active volcanism rather than plate flexure being the prime target, the long transect presented here is directed SW-NE along the supposed hotspot trace in order to sense the variation in structure of the lithospheric plate from its intact state before, during, and after having been submitted to the magmatic discharge of the mantle plume. The main refraction profile described here was recorded at sea by five OBS on the SW segment of the line of shots and two OBS on the NE segment (Figure 1), as well as on land in between by 32 portable, autonomous LS operated at about 1 km spacing along the main road crossing the entire island between the active volcano Piton de la Fournaise and the older volcano Piton des Neiges. Specific data retrieval and processing for OBS are described by *Nakamura et al.* [1987] and *Charvis et al.* [this issue].

2. Highlights of Structural Features Along the Transect: Layering and Heterogeneities

2.1. Overview

The multiple-receiver recording of seismic shot lines achieved is fundamentally important in wide-angle reflection-refraction seismics in order to obtain wave propagation along a system of reversed and overlapping profiles. Only such a cross sampling may resolve the variation of velocity with depth (layering) and the lateral variation of velocity (dip of interfaces or lateral heterogeneity within layers) from each other. Trial-and-error forward modeling then results in a two-dimensional (2-D) seismic velocity structure consistently but not uniquely accounting for the observations. In such a model, those elements required by data that are critical by their quality or sampling may not stand out from other ones that are less constrained. Therefore the present section is devoted to displaying crucial data and commenting on particular aspects of the structural model they constrain before the derivation of the 2-D velocity model in section 3.

The most relevant features, illustrated in Figure 2 by subsets of data recorded by the LS and OBS receiver array, can be summarized as follows:

1. At up to 20–25 km shot-receiver offset, first arrivals can be correlated with apparent velocities in the range 3–4.3 km/s and correspond to waves refracted within the volcanic pile (phases labeled 2a and 2b in Figure 2). At larger offsets, beyond 30–40 km, first arrivals define apparent velocities over 5 km/s, characteristic of refractions in the oceanic basement (phases 2d and 3 in Figure 2).

2. The deep part of the crust provides secondary, reflected arrivals with conspicuous energy at offsets in excess of 30 km (phases labeled *PiP*, *PmP* in Figure 2), as well as refracted phases (*Pi*, *Pn*). The upper mantle refraction *Pn* can be followed in many cases to more than 100 km offset. The structural meaning of these deep crustal phases will be analyzed in detail in later sections.

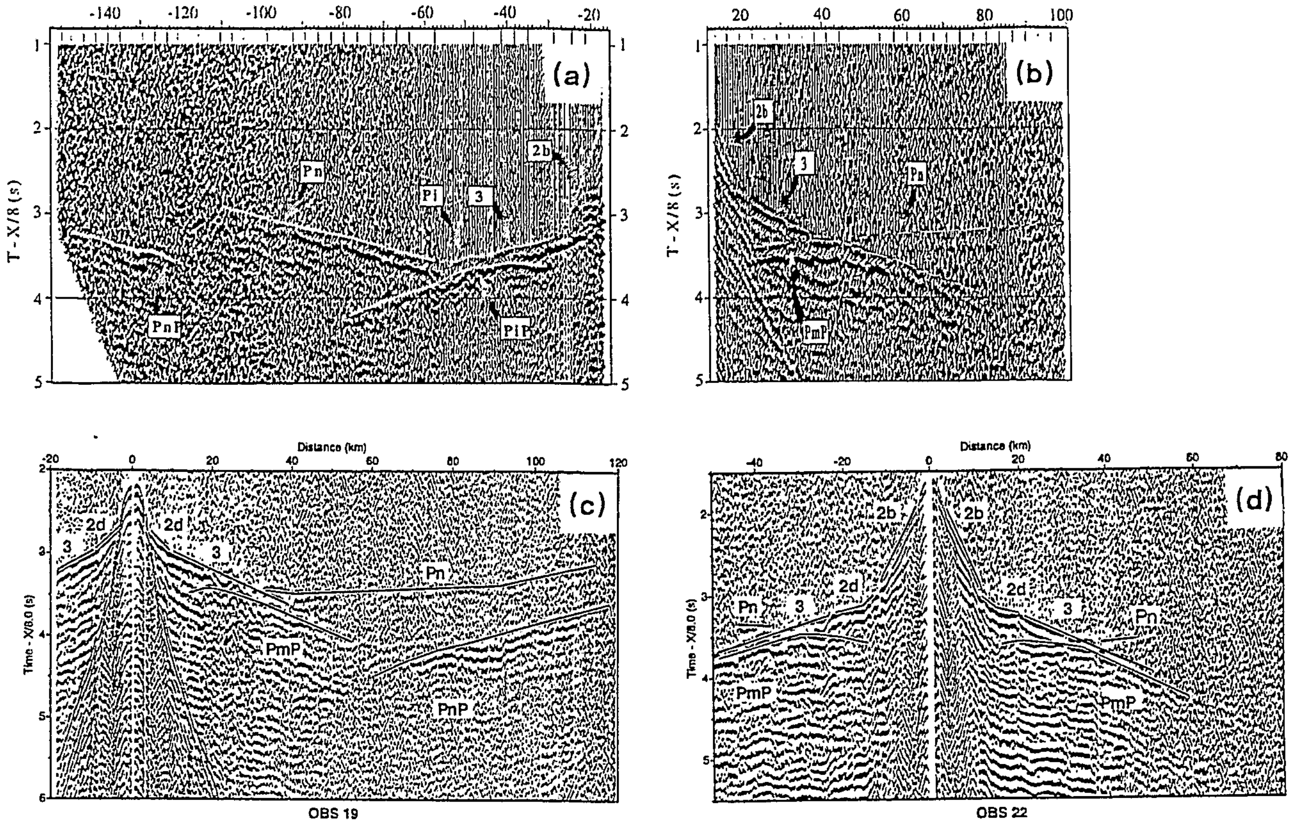


Figure 2. Record sections, with reduction velocity of 8 km/s and times corrected for bathymetry under shots. (a) Land seismograph station LS 29 on the SW coast for the shot line to SW, to the open sea. (b) Station LS 7 on the NE coast for the shot line toward Mauritius Island in the NE. Note difference of slopes of *Pn*, straight travel time curves of mantle-refracted waves, which are discussed in text in terms of asymmetry of structure behind and ahead of the hotspot trace. Note on the left side, toward SW, that the *Pn* straight travel time curve is not tangent to the hyperbola of the deepest crustal reflection (*PiP*), suggesting the presence of a layer with intermediate velocity between the normal base of the crust and top of the mantle. (c) Ocean bottom seismograph (OBS 19) providing reversed observations to LS 29 with smaller velocity slope of *Pn* travel time curve, confirming that its high apparent velocity on LS 29 records is due to NE, islandward dip of the top of the mantle. (d) OBS 22, showing slight asymmetry indicating further dip of crustal layers from La Réunion toward Mauritius. Note on the SW profile recordings the existence of a late phase *PnP*, 0.5 to 1 s after *Pn*, indicative of further heterogeneities in the lithospheric mantle.

3. Finally, some strong, late arrivals labeled *PnP* (Figure 2) can be observed, 0.5–1.5 s after the *Pn* phase along the SW segment of the transect. This deep energy appears as several successive arrivals (the signal length of 1 s is clearly in excess of the 0.3 s characteristic of the seismic source) which can be attributed to a reflective zone within the mantle.

The lateral continuity observed for the main seismic phases indicates crustal layering on a regional scale, although some marked variations in arrival times between neighboring LS also suggest the presence of lateral heterogeneities on a smaller scale beneath the island.

2.2. Crustal Layering

The most detailed resolution of the edifice of recent volcanic deposits over the preexisting oceanic crust is achieved near the coast, at the OBS located over its thicker part and the LS nearest to the line of shots. Figure 3 displays examples of recordings on LS 39 and 1, placed just onshore of the SW and NE profiles of shots. The first arrivals between 5 and 15 km offset define two short refracted branches, 2a and 2b, with apparent velocities of 3.4 and 4.1 km/s, indicating two levels within the upper volcanic products. The 2b arrivals vanish abruptly at 15 km offset, and first arrivals visible afterward (phase 2d in Figure 3) are clearly delayed with respect to their extrapolation, by 0.5 s in the NE profile and by more than 1 s in the SW line. Such delays strongly suggest the existence of a low-velocity layer (a "2c layer" not directly detectable in refraction seismics since no turning wave is associated with it) within the volcanic and sedimentary pile above the oceanic basement. According to the time delays, the low-velocity layer is thicker in the SW profile. It appears to be restricted to the island and its flanks, since no such delays are observed on more distant OBS (see OBS 19 and 24 in Figure 2).

At an offset around 18–20 km, the phase 2d, corresponding to apparent velocities well in excess of 5 km/s, can be associated with the oceanic basement, oceanic layer 2 [*de Voogd et al.*, this issue]. In cases where first arrivals are clear to larger ranges, a clear bending in the correlation indicates another seismic phase, labeled phase 3. This indicates the existence of a strong velocity gradient and a deeper layer. This refracted phase with higher apparent velocity and a more linear correlation can be attributed to a refraction in the lower oceanic crust (oceanic layer 3).

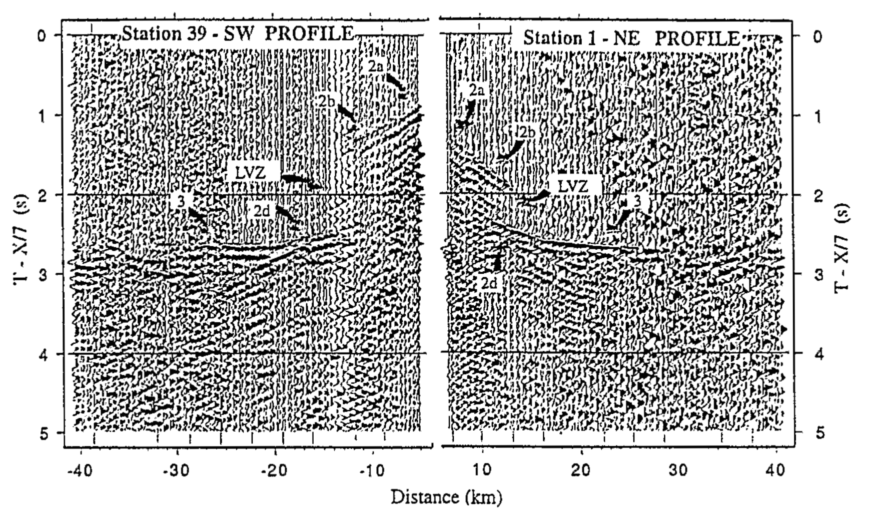


Figure 3. Record section details to document peculiarities of the uppermost crustal structure under the island's shores, from the nearest land stations. Reduction velocity is 7 km/s. Note clear arrivals with 4 km/s velocity at shortest offsets, defining layers 2a and 2b and a clearly delayed later wave with similar velocity, 2d. A low-velocity zone (LVZ) does not give rise to a travel time curve 2c in between. The delay between waves bottoming above and beneath the LVZ is larger to the SW than to the NE; hence the LVZ is thicker under the SW shoreline. Strong refractions labeled 3 are differentiated from those of 2d, with which they form a system of travel time curves for which the reflection on the interface between the two layers is seen forming a cusp with each of the two refractions. Velocities well in excess of 5 km/s characterize this layer 3, which may hence correspond to the oceanic basement. Layers 2a and 2b can be related to the products of the volcano, forming the upper part of the edifice erected above sea bottom. The LVZ and refractor 2d encompass the lower part of the volcanic edifice and preexisting oceanic sediments. The interface between them is presumably inside the LVZ, refraction 2d building up at the level where the velocity increases again with depth to its value in the lid of the LVZ.

2.3. Crust-Mantle Transition

The structure of the deep crust and the transition to the mantle are constrained by observations of a number of energetic reflected and refracted phases. Clear first arrivals beyond offsets of 38 km (OBS 19 and 22 and LS 7) and of 60 km (LS 29) in Figure 2 line up to large offsets with a mantle velocity. However, the signature of the *Pn* mantle refractions shows marked differences along the two segments of the transect. In the SW one, *Pn* is clearly observed in all the LS and OBS record sections, and the apparent velocities measured are about 8.7 km/s seaward and 8.0 km/s landward (see LS 29 and OBS 19, Figure 2). This documents both a rather high value of the true velocity beneath the refractor, which ascertains that these are *Pn* mantle refractions, and a significant deepening of the interface toward the island. In the NE profile, *Pn* has a much lower signal-to-noise ratio and is only observed in a few LS and OBS record sections for limited distance ranges. This indicates that there is not as strongly positive a velocity-depth gradient in the mantle beneath the Moho as beneath the SW segment. Here the apparent velocities in both senses of propagation are similar, 8.1–8.2 km/s (see LS 7 and OBS 22 in Figure 2), suggesting a rather constant Moho depth downstream of the island along the inferred hotspot trace.

Significant differences in the structure of the deep part of the crust under the two flanks of the island are also obvious in the data. Along the NE profile (LS 1 in Figure 2), the phases observed form the typical system of reflected-refracted branches of a simple crust-mantle boundary. The correlation of *Pn* refraction arrivals is tangent to that of the *PmP* reflection, which is asymptotic to large offsets with the continuation of refracted phase 3 (LS 7, OBS 22, Figure 2). A similar pattern is observed at the far end of the SW profile, on the presumably undisturbed oceanic crust, far ahead of present hotspot volcanism (OBS 19, Figure 2). However, closer to the island in this SW profile, a more complicated seismic pattern is resolved (LS 29, Figure 2). Here the *Pn* correlation clearly cuts across the prominent deep reflected phase. This phase therefore cannot be the regular *PmP* expected in association with the *Pn* refractor, and it is labeled *PiP*. It is reflected from a different, significantly shallower level. Its strength indicates a significant impedance contrast, suggesting that material of normal crustal velocity may be bounded at depth by this reflector, shallower than the *Pn* refractor. Associated with this reflecting level, a refracted *Pi* phase can be observed in LS 29 around 50 km offset, but only in a very short distance range.

The existence and lateral extent of these phases and of the corresponding peculiar crust-mantle transition zone can be documented from independent observations at different offsets and locations. Figure 4 displays the section from shots on the SW line obtained at a more distant station on the NE side of the island (LS 7, Figure 4a). The *Pi* refraction can be correlated between 45 and 70 km with an apparent velocity of 7.7 km/s, being tangent to the *PiP* reflection and well distinguishable from the *Pn* refraction defined after 75 km offsets. An alternative explanation might be that the change of wave pattern is due to horizontal variation in structure instead of velocity-depth variation. The various phases observed would then come from a single horizon with some local changes in topography. This can be ruled out by the overlapping observations provided by the line of stations on land, since on another station LS 35 (Figure 4b), closer to the shoreline no significant local perturbation occurs at this shot position in the correlation

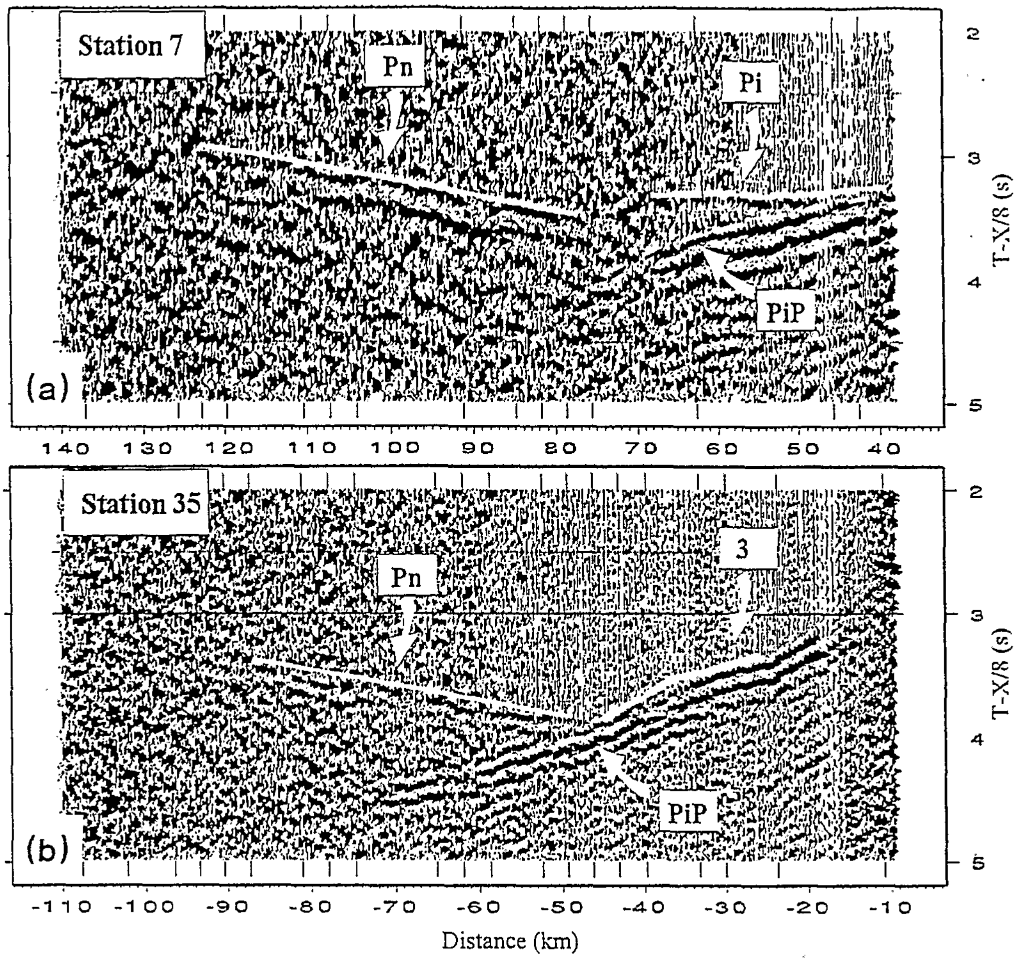


Figure 4. Lower crust-mantle structure complexity. The record sections of (b) LS 35 and (a) 7 are plotted and aligned for the same set of shots along the SW profile. At LS 35 the travel time curve of first arrival, here phase 3, is smooth. In contrast, for this same set of shots the first wave field arrivals at LS 7 show a complexity for which a local or superficial origin can hence be excluded as it should affect similarly LS 35. Multipathing is hence implied at the depth reached for waves propagated from these same shots towards LS 7, i.e., at crust-mantle depth. Accordingly, a layer of velocity intermediate between the lower crustal layer, which gives rise to phase 3, and the upper mantle layer, which gives *Pn*, has must be inferred, and this layer must be thicker at the bottoming point of rays to LS 7 than to LS 35.

of the two visible phases, *PiP* and *Pn* (the *Pi* phase has not emerged yet for that recording distance range). Therefore the seismic phases observed and the velocities interpreted beneath the SW flank of La Réunion suggest that the bottom of the normal crust giving rise to phases *PiP* and *Pi* does not coincide with the top of the normal mantle, with associated phases *PmP* and *Pn*. An additional layer, with a position and a velocity intermediate between lower crust and upper mantle, must be introduced to account for the observations.

2.4. Lateral Heterogeneities in the Volcanic Edifice and Upper Crust Beneath Réunion island

When comparing the record sections of the same profile from one land station to the other, the wave field appears to be displaced to shorter propagation times when considering data of stations going northward from station 22. Either this advance could be acquired by shallowing topography of the interfaces where the rays bottom or could be due to lateral variation due to a higher value of the velocity in the overburden. Since the experiment was designed with a line of shots on the two sides of the island, it provides reversed and overlapping observations to resolve which one of these two types of structure is responsible for the anomaly observed.

The test between topography of a deep interface and lateral velocity heterogeneity in the overburden can be visualized in the form used in marine two-ship constant offset profiling. In Figure 5, for each station, we represent the data in the range of 30–40 km, with a reduction velocity or linear move-out display. The onset of waves in this reduced arrival time plot is then a delay time to the interface as a function of the horizontal position. This delay time gets shorter northward of station 22 (toward the center of the island) for the two opposite azimuths. Corrections for topography effects (higher elevations in the middle part), not considered in Figure 5, should still increase the differences. Since the horizontal variations on the two sections coincides for common receivers, the structure causing the delay anomaly is most likely in the overburden near the receivers, not at the level where rays bottom, since the shot-receiver midpoints for the two opposite azimuths of incoming waves are located far from each other. The section from the northern azimuth senses the variation far enough northward to

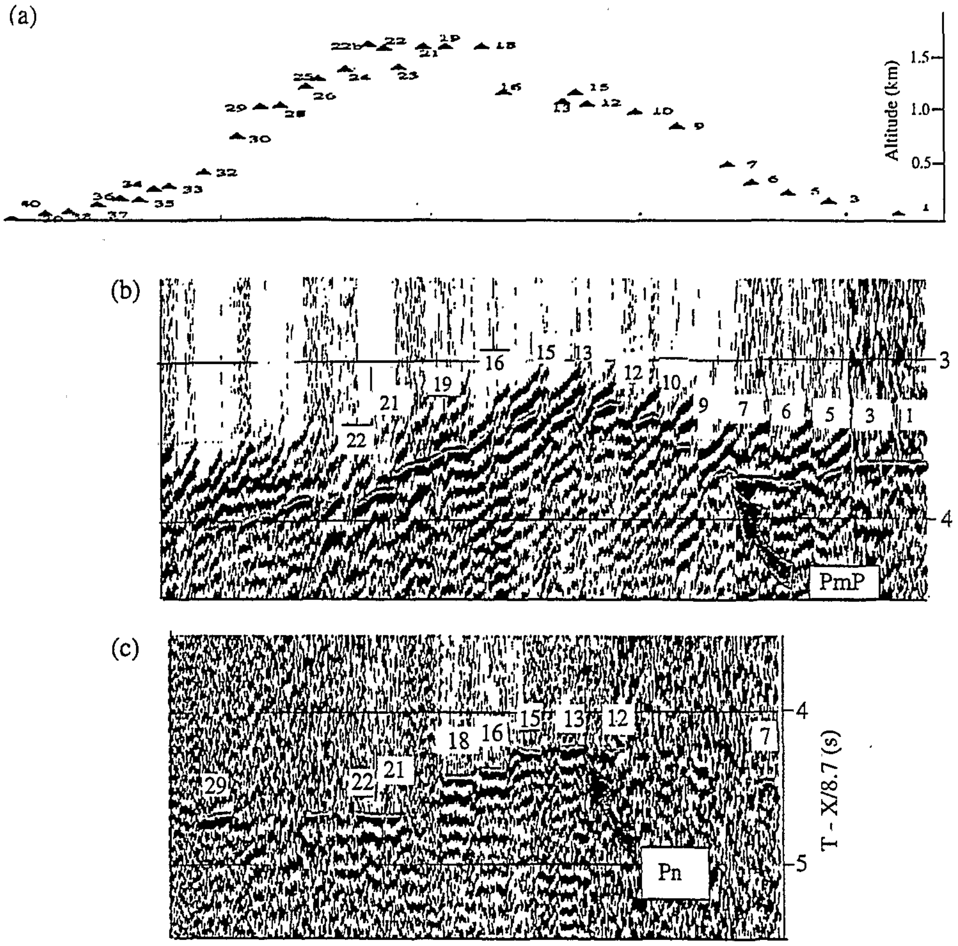


Figure 5. Illustration of strong lateral velocity contrast in the crust across the island. (a) Station numbers and their elevation. For Figures 5b and 5c, synthetic aperture profiles are constructed, taking for each station the data in the same offset, corresponding to 50 shots spanning a 10 km range. (b) Display of the section from shots on the NE profile. The *PmP* reflections show an early arrival in the middle of the section. This can be attributed either to topography of the reflector at the midpoint or to a velocity pull-up, i.e., higher velocity in the overburden on the downgoing or upgoing path from the particular shots to the particular stations. (c) Same as Figure 5b but taking data from the line of shots to the SW to the same stations. This also shows a spatial change of arrival times. This time, the *Pn* refraction is plotted. There is no common part of the midpoints for this data set (to the SW of the island), with the previous ones (to the NE). The same holds for the downgoing paths on the shot sides. Since the shapes of the anomalies from the two opposite azimuths grossly correlate when plotted here as receiver gathers, the lateral velocity variation is occurring on the upgoing part of the rays under the receivers, i.e., across the island. The synthetic aperture data are obtained by compensating times for variable water depth on the shot side, but in the plot, no elevation correction to the receiver side has been applied. Any such correction would enhance the size of the anomaly since it is the highest receivers, to which the propagation path is longest, which have the earlier arrivals.

see times getting longer again at LS 5. The high velocity in the shallow part of the propagation path is hence confined in space between stations 22 and 5. These high seismic velocities can be associated with the existence of plutonic bodies intruded into, or surrounded by, extruded material of lower bulk density.

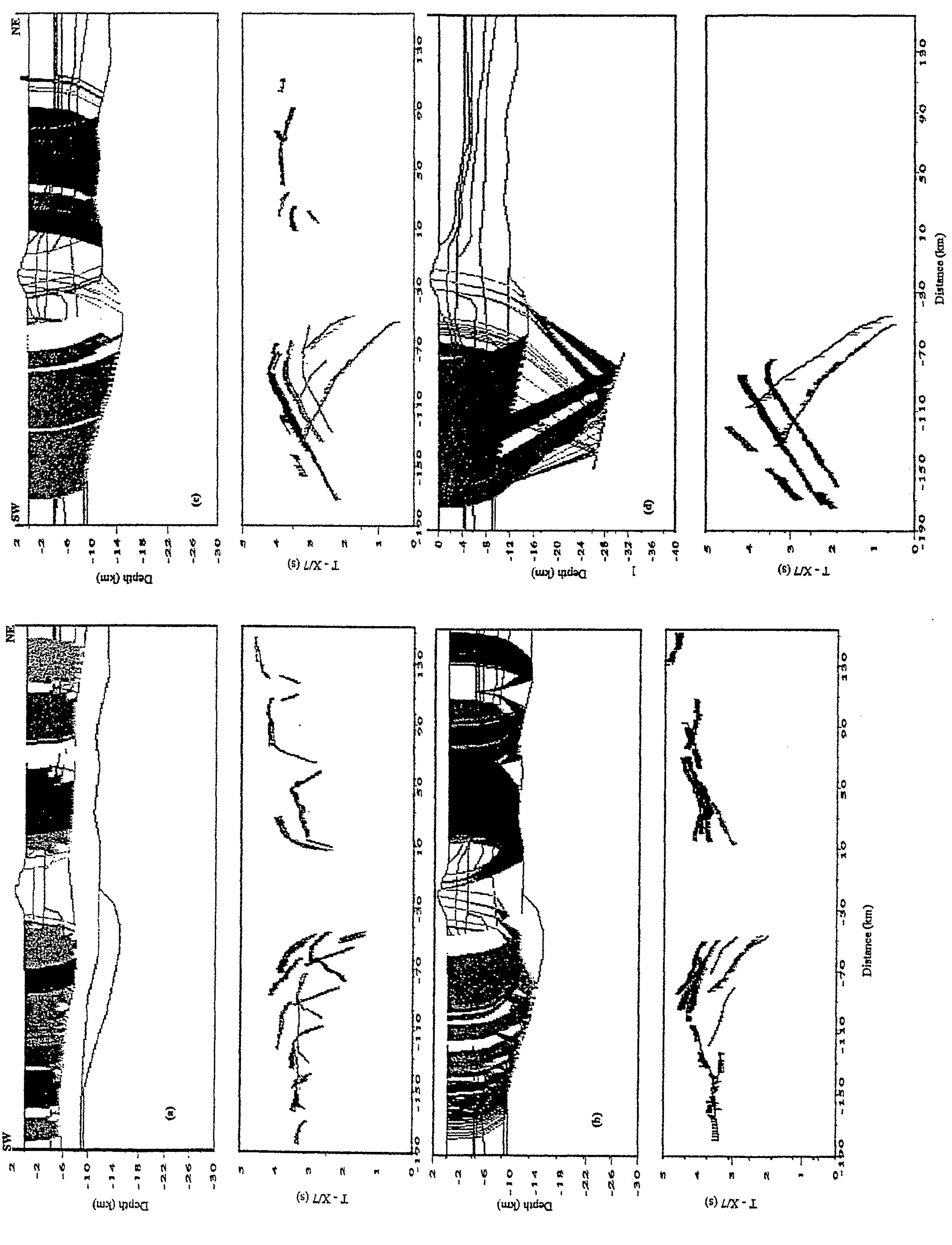
3. Interpretation: A 2-D Lithospheric Model

3.1. Forward Modeling, Data Inversion, and Amplitude Control

The interpretation of the reversed and overlapping observations along the seismic transect has been undertaken with the procedure of *Zelt and Smith* [1992] which, in addition to classical forward modeling of arrival times and amplitudes, allows inversion for parts of the model having appropriate coverage.

An initial 1-D model has been derived after travel time identifications on selected receiver gathers displayed as record sections corrected for water depth variation. A 2-D model was then built, taking into account the main differences in wave

Plate 1. (opposite) Modeling and inversion of the arrival times of the system of reversed and overlapping observations with the procedure of *Zelt and Smith* [1992] used to constrain the 2-D model of Plate 2. Examples of ray coverage and fit of the observed (black) and computed (color) arrivals for various phases. From top to bottom, (a) Upper crustal and midcrustal refractions 2a, 2b, 2d, and 3. (b) Reflections at the base of the crust and top of the mantle, *PiP* and *PmP*. (c) Refractions beneath the base of the crust, *Pi* and *Pn*. (d) Mantle phase *PnP*, modeled as a wide-angle reflection.


(a)
(b)
(c)
(d)
SW
NE
Depth (km)
T - X/7 (s)
Distance (km)

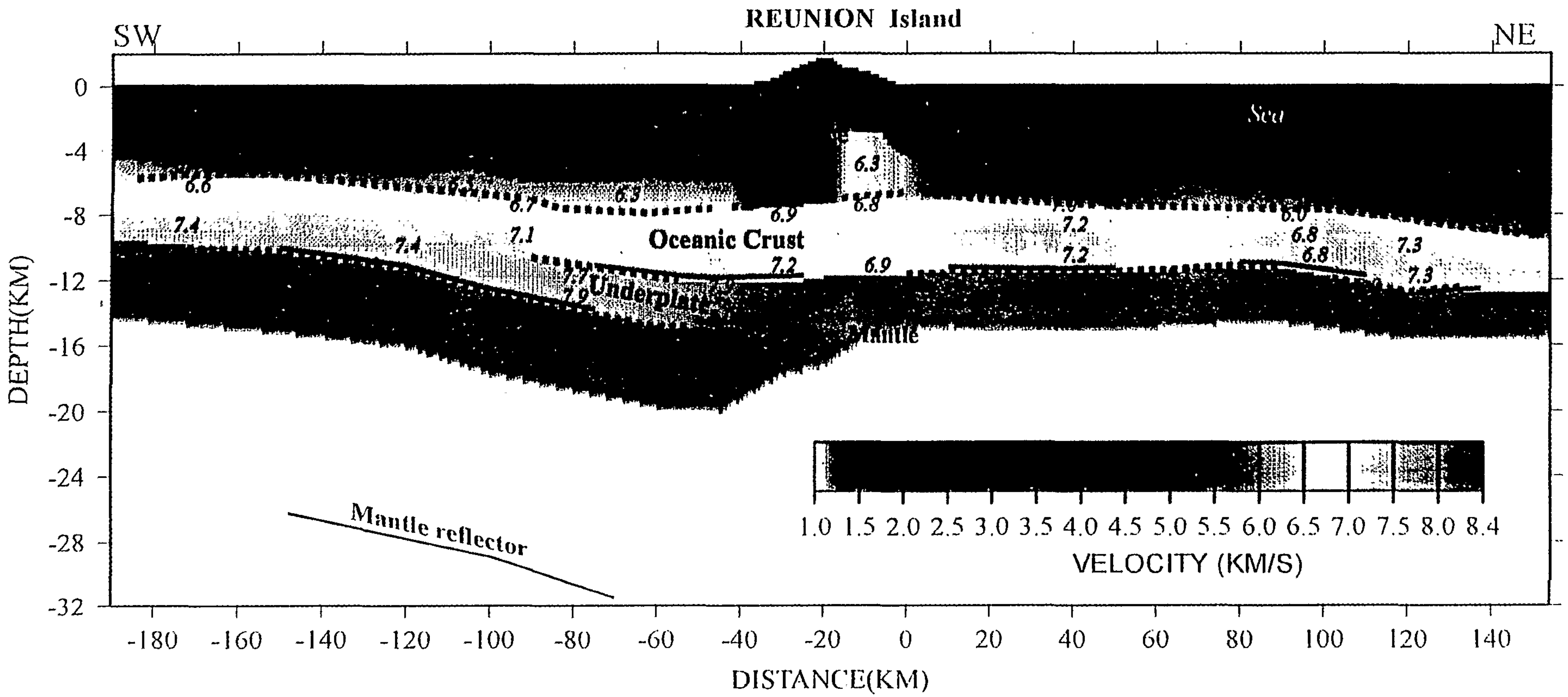


Plate 2. Two-dimensional P velocity model on a SW-NE transect through La Réunion from 36 land stations and 6 OBS recording 200 m spaced air gun shots along the marine part of the transect. Results from travel time modeling and inversion and modifications of gradients from comparison with synthetic seismograms are shown. Bold lines are interfaces sampled at reflection turning points by observations; dotted lines are interfaces where hit by downgoing or upgoing refraction at corresponding critical incidence. Note (1) lateral variation in green-blue upper crustal structure under the emerged part, (2) its greater thickness toward Mauritius in the NE along the hotspot trace, (3) variation in gradient in yellow-orange deeper crust, (4) intermediate velocity layer under the SW part of the volcanic edifice, orange-red above deep red mantle (possibly grading into the mantle color of the NE part of the transect), which represents crustal underplating, and (5) only limited deepening of interfaces of the top of the prevolcanic crust and of the mantle toward the island, over the 160 km from the open ocean in the SW.

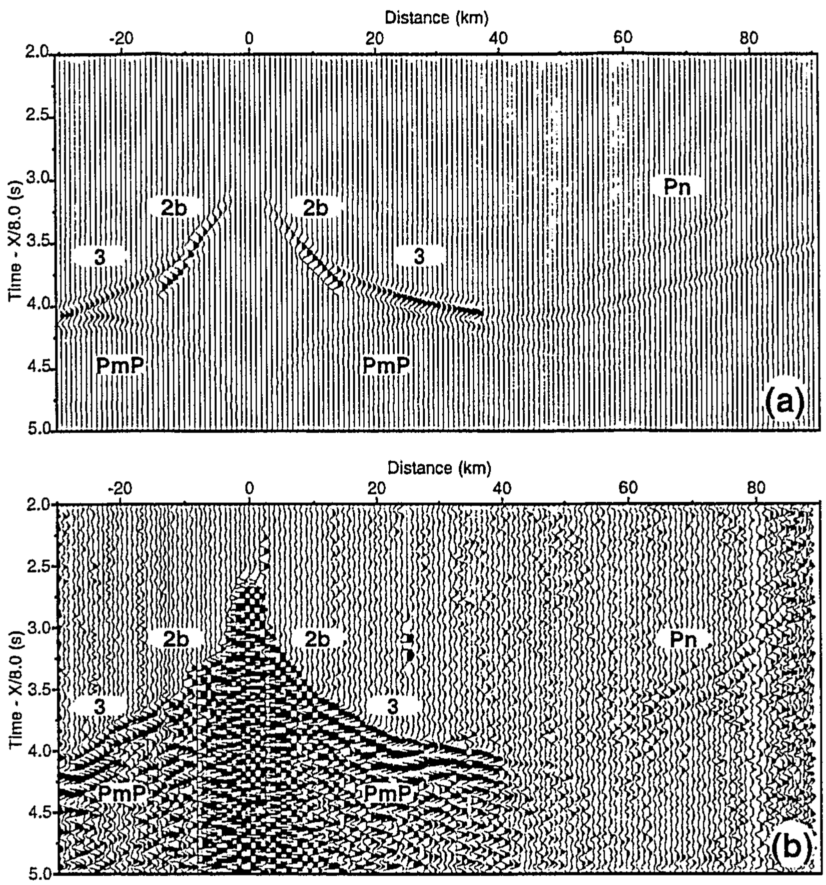


Figure 6. Amplitude constraints on velocity model. Examples of observed and synthetic seismograms sections displayed with same linear increase of scale factor with offset. (a) Synthetic seismograms corresponding to OBS 18. (b) OBS 18 record section for SW line of shots. Note sharp falloff of amplitude around 40 km offset, reproduced in the synthetics computed in the 2-D model inverted from travel times, which introduces a strong gradient in layer 3 bringing to shorter ranges the far-offset cusp of the refraction within this layer and the reflection *PmP* at its bottom. (c) Synthetic seismograms computed for OBS 22 in the final model. (d) Synthetic seismograms computed for OBS 22 in a model with a constant vertical velocity gradient in layer 3. (e) OBS 22 record section for the NE line of shots. The *PmP* amplitude corresponding to the model resulting from inversion of travel times remains high at larger offsets than in the observed data (compare Figures 6d and 6e). Introducing a higher gradient in the upper part of the layer, followed by a constant velocity in the lower part. reproduces more accurately the observations (compare Figures 6c and 6e).

field adjustment by forward modeling of the 6 OBS and 32 LS considered. In a further step, the inversion procedure is applied, layer after layer, considering always a trade-off between number and resolution of model parameters [*Driad*, 1997]. This procedure decreased the root mean square (RMS) of the residual times by 60% on average with respect to the initial model. The resulting accuracy in travel times is 0.05 s on average. Plate 1 shows the ray-tracing coverage and travel time fit for observations associated with the different crustal layers defined in the final model (see Plate 2), i.e., refracted waves in the volcanic upper layers (Plate 1a); *PiP*, *PmP* reflections at the base of the crust and top of the mantle (Plate 1b); the corresponding *Pi*, *Pn* refractions (Plate 1c); and the mantle wave *PnP* (Plate 1d).

In addition to travel time modeling, amplitudes have been checked in a detailed synthetic seismogram analysis, which discriminates between different models with similar travel times. Examples of how amplitudes may constrain the velocity-depth distribution in oceanic layer 3 are shown in Figure 6. In the SW profile, the model resulting after the inversion has a steady velocity gradient in the lower crust, from 6.6 to 7.4 km/s. The corresponding synthetic amplitude pattern (Figures 6a and 6b) fits the observed amplitudes, especially the abrupt vanishing of the *PmP* phase around 40 km offset. On the other hand, in the NE segment the velocity gradient in this layer 3 produces high-amplitude *PmP* waves to larger offsets than observed (Figures 6d and 6e). Improved amplitude fitting is achieved by introducing a higher gradient in the first 2 km, followed by a constant value to the bottom of the layer (Figure 6c). The corresponding amplitudes at the land stations are also consistent with this two-level subdivision of layer 3 along the NE profile.

3.2. Velocity-Depth Model Along the Transect

The final 2-D velocity-depth model along the La Réunion transect is shown on a color display (Plate 2). Levels directly sampled by turning points of reflected and refracted waves are indicated by bold and dotted lines, respectively. The model documents the general crustal layering: volcano-sedimentary sequences, basaltic basement, oceanic lower crust, and the crust-mantle transition. Structural variations, at different scales, ahead and behind the hotspot trace have also been

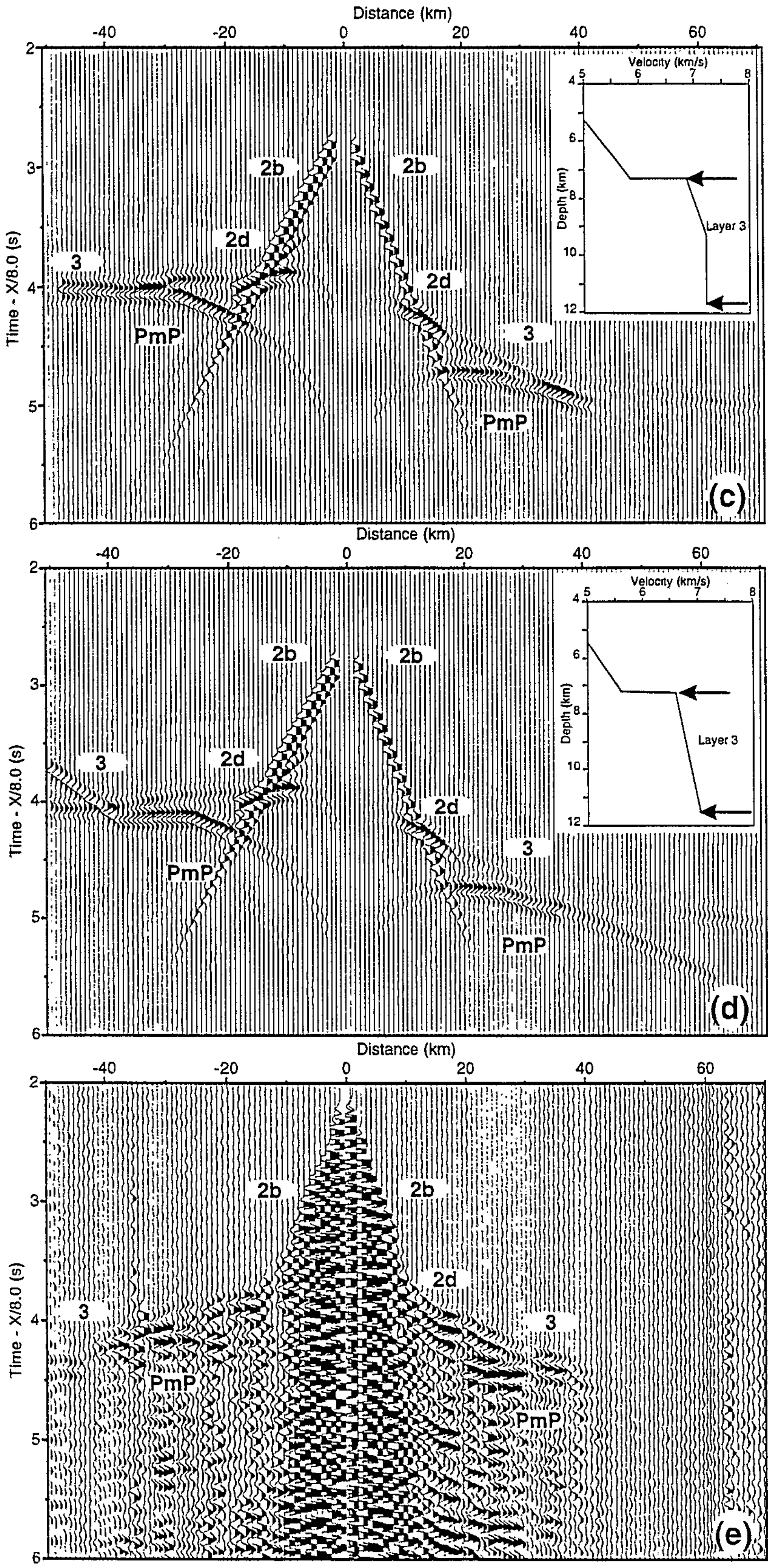


Figure 6. (continued)

inferred and are marked at the crust-mantle transition and sub-Moho levels. The main features illustrated in this model are as follows: (1) short-wavelength lateral velocity variations of the volcanic cover under the onshore part of the transect, and greater thickness of the upper crust including the basaltic basement towards Mauritius in the NE than to the SW, (2) beneath the island flanks, the superposition of volcanic products on preexisting oceanic sediments marked by a seismic low-velocity layer, which is thicker on the SW side of the island and thins out oceanward on both sides, (3) the presence of a high-velocity body embedded in the upper crust, well confined beneath the central northeastern quarter of the onshore transect, (4) a steady velocity gradient within the lower crust (oceanic layer 3) ahead of the hotspot trace contrasting with a division into two levels of this layer toward Mauritius (differences in velocity gradients are also observed beneath the Moho, with higher values in the SW segment), (5) the existence under the SW part of the transect of a layer between the bottom of the crust and the top of the mantle, with intermediate velocities, ranging from 7.4 to 7.9 km/s, well constrained by the interpretation of the dense, overlapping recordings of the complex wave field pattern of reflection and refraction groups (this layer of crustal underplating has a maximum thickness of almost 3 km beneath the SW flank of the island, wedging out some tens of kilometers offshore and it has not been observed in the NE segment), (6) very limited islandward dip of the different crustal levels, apart from the Moho deepening in the SW segment due to the underplated body, and (7) the existence of another velocity-depth variation in the mantle beneath the SW segment, represented in the model by an interface with a minimum velocity contrast of 0.2 km/s around 30 km depth.

4. Implications and Discussion

4.1. Heterogeneity of the Edifice Under the Island: Implication for Plate Loading and Volcanic Style

Under the island itself, a part of the northern half is underlain by a high-velocity body which reaches through the volcanic edifice with rather sharp sides. A Bouguer gravity anomaly high which connects the Piton des Neiges and Piton de la Fournaise at this latitude [e.g., *Lesquer*, 1990] could be caused by the high density consistently associated with such a high-velocity body. It has a N120°E directed axis, which is also marked by Strombolian cinder cones, some of which are recent, between the two large volcanoes. According to early interpretations [*Chevallier and Bachèlery*, 1981], such a direction could be regarded as magmatically controlled, with the axis interpreted to be a rift zone of the older Piton des Neiges volcano on which the first edifice of the Piton de la Fournaise volcano would have been built. However, since seafloor spreading magnetic anomalies mapped in the Mascarene Basin [*Schlich*, 1975] show a similar N120°E direction, the axis between the two volcanoes has also tentatively been related to the inherited structural grain of the oceanic plate or even considered to correspond to a fossil spreading center, providing a direction or structure which could be preferentially reactivated by the hotspot [*Schlich*, 1982].

The velocity structure resolved for the edifice has implications in terms of mass balance and plate flexure. Lithosphere loading models commonly consider the whole volume deposited on top of the preexisting plate as having a crustal density [e.g., *Wolfe et al.*, 1994]. For La Réunion, the seismic transect documents lateral heterogeneity under the island, with three-quarters of the section being of much smaller velocity (~4 km/s), and density down to the oceanic plate, than that of the bulk oceanic crust underneath (>6 km/s). In the common assumption of high density, the total mass of the surface load expected is much larger than observed, and the same should apply to the amount of flexure the estimate of which should be too large. The present seismic evidence shows, at best, a limited deepening of the top of the plate toward the island on the SW side. The high-velocity material in the edifice does not dominate in volume over the lower-velocity extruded material even in the part of the transect above the island, which has implications with respect to the regime of volcanic discharge.

With this transect, we have not attempted to sample the several hundred kilometer wavelength of the structure expected for plate flexure. However, the depths to markers of the oceanic plate can be constrained by this survey in the vicinity of the island, which is the center of the applied load and hence the site of maximum plate deflection. The top of the oceanic sediments is not a clear refractor under the volcanic edifice, but the top of the basaltic basement can be used as such a marker, since it is reliably documented by its specific refracted and reflected waves. With respect to the regional depth of 4200 m to the sea bottom, the top of the oceanic basement along this transect is found at 5 km depth 160 km to the SW of the island, and not more than 6 km depth beneath the La Réunion coastlines and 7 km toward Mauritius Island. The top of the oceanic plate or the basement is at most slightly depressed on the transect of Plate 2, and the areal coverage by MCS establishes that even this dip is not present elsewhere around the island [*de Voogd et al.*, this issue]. The measured depth to the preexisting plate can be contrasted with the much larger depths which have been proposed from indirect estimates of the plate deflection: 4–6 km from geoid anomalies [*Bonneville et al.*, 1988], and gravity modeling [*Lesquer*, 1990; *Malengreau*, 1995]. Clearly, the expected depth of 8–10 km to the base of the volcanic material is not documented by our direct seismic measurements, which place the oceanic basement much shallower.

The topography of the crust-mantle boundary can also be considered in order to estimate the plate deflection. Far south, it is at a normal oceanic depth of 10 km and only a kilometer or two deeper between La Réunion and Mauritius Island, showing thus only a limited variation over 250 km from the oceanic domain into the hotspot trace. Over the 60 km centered on the southern coast, the crust-mantle boundary could be either at 10–11 km depth or, depending on its definition, locally 2–3 km deeper as discussed in section 2.3. The base of the layer of normal crustal velocity, which may best represent the base of the original crust, is not significantly depressed, a feature confirmed on profiles outside the present traverse [*Charvis et al.*, this issue]. The reflector at the base of the original crust would probably have been missed had coarser spatial sampling on the shot or on the recording side restricted the interpretation to first-arrival refraction analysis. The Moho would then have been interpreted to flex downward by 3 km toward the island, more similar to the structural model of the island of Hawaii [*Hill and Zucca*, 1987] based on such an early, less tightly sampled experiment.

4.2. An Intermediate Layer at the Crust-Mantle Level: Relation With Models of Underplating

Data along the profile to the southwest of the island establish that strong reflections at the base of the crust and clear refractions in the mantle come from two interfaces at different depths, which define between them a layer of material with intermediate velocity between 7 and 8 km/s. This layer, almost 3 km thick under the south coast of La Réunion, wedges out and cannot be distinguished farther offshore than 80 km SW of the island. Layers or bodies with seismic velocities intermediate between those of normal crust and mantle have been previously reported beneath old volcanic islands and plateaus and at passive continental margins. They have been interpreted as representing mineralogical or compositional heterogeneities modifying the normal lithosphere. The specific features of La Réunion may first be discussed in the perspective of the corresponding evolutionary models.

In Hawaii a broad and thick level with intermediate velocity between the crust and mantle [*Watts et al.*, 1985] has been considered as an underplated body resulting from addition of hotspot material. *ten Brink and Brocher* [1987] suggested that underplating is mainly controlled by the increase with time of the compressive stresses under the volcano as a function of the increases in its size and in the resulting plate flexure, favoring subcrustal magma intrusions at a late stage of evolution. The intermediate value of seismic velocity in the underplate can in this case be due to an average between the basaltic intrusions and mantle. The situation is different in the Réunion transect, since the underplate is observed here under the presently active volcanic island and lithospheric flexure is absent or small. It therefore does not seem possible to attribute underplating to flexural stress fields.

Another mechanism for magmatic crustal underplating may consider the density contrast between the upwelling magma and the overlying crust. The seismic velocities determined in underplated bodies (7.2–7.8 km/s) are intermediate between velocities attributed to gabbros (6.7–7.2 km/s) and peridotites (7.8–8.3 km/s) [*Holbrook*, 1995]. These bodies may be the result of a mixing of intrusions and crystallization with part of mantle material. On the other hand, if they are massive homogeneous additions, their high seismic velocity suggests that their composition can be interpreted as marking the magnesian enrichment of rocks associated with hotspots [*White and McKenzie*, 1989]. Despite the fact that erupted lavas are essentially alkalic to tholeitic basalts, it is likely that for a hotspot, most of the primary melt generated at depth has a large MgO content (MgO > 16%). trending toward picritic composition [*Farnetani and Richards*, 1994; *McKenzie and Bickle*, 1988]. However, picrites appear rarely at shallow crustal levels since their density is higher than the average density of the crust and the melt can only rise to the surface if its density has been lowered by fractionation of olivine, augitic pyroxene, and variable amounts of plagioclase [*Farnetani and Richards*, 1994; *Farnetani et al.*, 1996; *White*, 1993]. Cumulates of these minerals in various proportions lead to seismic velocities of 7.5–7.9 km/s, consistent with those determined by seismic methods under provinces suggested to have been formed by hotspot activity [*Farnetani et al.*, 1996]. At Piton de la Fournaise, high-MgO lavas, locally known as oceanites, occur in rare eruptions of large volume among more frequent smaller eruptions. *Albarède et al.* [1997] conclude from the geochemical analysis of compatible elements that the lavas last equilibrated in solid-dominant conditions, precluding the evolution in a large superficial magma chamber and that this differentiation may have occurred during the magma ascent through dikes filled with olivine and clinopyroxene crystals in the upper mantle. These products may provide heterogeneity in the upper mantle, at the level of the underplate, but also possibly of the deeper level described later.

At La Réunion, we documented the presence of an intermediate velocity body at crust-mantle level in the vicinity of a recent, still active volcano island, the first observation of such a case to our knowledge. The formation of the underplate hence does not postdate the main volcanic activity at the surface but may be considered as coeval and hence genetically related. Such a relation is assumed in the model of hotspot melt transport and fractionation of *Farnetani et al.* [1996]; thus this model may be regarded as generally supported. However, the model of *Farnetani et al.* [1996] attempts to fit the seismic velocities measured now in structures over hotspots which are long extinct, so that the seismic structure of La Réunion may not readily be interpreted in such a model. Here seismic velocities are measured while the processes are active, and since the material is intruded as melt, the velocity values will depend on the proportion of the still-molten fraction and not only on the composition.

4.3. Perturbation to the Mantle in Relation to Active Volcanism

The presence of a layer with a velocity intermediate between normal crustal and mantle materials at La Réunion, above an active hotspot, may not necessarily imply inclusion of new material. The value of the velocity beneath the previous Moho could have been decreased with respect to that of normal mantle by the effect of an increase in temperature on mantle material, without magmatic transfer. However, since the heat provided by conduction presumably would come from depth, the temperature would be higher in the mantle material beneath the anomalous layer. Since we record a clear high-velocity *Pn* wave propagating at these larger depths, such an explanation is not satisfactory. A more efficient cause for lowering velocity beneath Moho is the presence of a partial melt fraction, but in situ melting is unlikely just at the top of the mantle by conductive heat transfer for the same reason as before. This implies that the inclusion of a melt fraction has to occur by upward migration of melt. The anomalous layer can then be seen as being due to preferential ponding of melt, most conveniently at a depth where there is a density barrier, such as below the crust. Such an interpretation accounts not only for the succession of layer velocities with depth but also for the nature of the boundaries of the anomalous layer constrained by seismics as described earlier. The clear reflector at its top is the base of the previous crust, and the variation of the velocity gradient into the normal mantle may mark the depth distribution of ponded material. In this interpretation for the origin of the anomalous layer, its average velocity depends on the relative proportions of preexisting mantle affected, of intruded material, and of its still-molten fraction, all materials with different velocities. From the single value of the velocity measured, the composition, amount, and present physical state of the material thus intruded cannot uniquely be derived.

The complexity of the observed phases indicates even deeper heterogeneity. A striking feature on many record sections, on land or at sea, is the strong phase arriving after *Pn* and with a larger apparent velocity (e.g., phase labeled *PnP* in Figure 2a). This phase does not indicate the same depth in all record sections, and if it were caused by a single continuous interface, this would need to have very strong local topography. The phase merely documents that localized strong velocity heterogeneity exists in the 25–45 km depth range where it has its turning point, well beneath the crustal layers. Models of the deep structure under a hotspot have generally been derived from considerations of the flexure of the lithosphere under the load, the fit to observations being achieved for a flexural rigidity supposedly reduced due to thermal rejuvenation, or thinning of the lithosphere [*Detrick and Crough*, 1978]. However, reheating can hardly provide the localized impedance contrasts observed by seismics in the upper mantle under La Réunion, which rather suggest sharp local contrasts in composition or amount of melt. In their alternative model accounting for the geoid and bathymetry over the Hawaiian region, *Phipps Morgan et al.* [1995] propose that melt extraction and transport induce a double compensation, by density alterations at two levels. The upper level where hotspot swell material accumulates is the uppermost mantle-crust level. A deeper one is located under the preexisting lithosphere where density is lowered by the depletion due to melts having migrated upward.

4.4. Variation of Structural Perturbation in Relation to Plate Drift

If complexity at the crust-mantle level to the southwest is indicated by the clear observation of the presence of additional phases at several stations and offsets, this is not so to the northeast. A spatial variation of structure at the crust-mantle level is accordingly evident in the section of Plate 2 along the hotspot trace.

In the common view that the plate drifted NNE with respect to the hotspot, the evolution of the seismic structure in space along the transect could be taken as corresponding to its evolution with time, provided, however, that the plume discharge can be regarded as continuous in time and confined to a point in space. The increase of thickness of the underplate from a distance of 100 km south of the southern coast toward the island could here be viewed as corresponding to the longer time the downstream parts of the drifting plate have been subjected to material supplied from the hotspot, La Réunion being now to the NNE of it. However, there is no underplated layer identified farther downstream than La Réunion. The underplate is thickest and most clearly defined between the southern coast and 50 km offshore southward, by wide-angle reflections from its top and refractions from beneath it. The base of the underplate, or the top of the mantle, must be deeper than normal to at least midway through the island to satisfy *Pn* observed there from the SW. However, the *Pn* observed there from the NE propagates right along the base of the oceanic crust, indicating that the underplate does not continue as a layer farther downstream than La Réunion. The ray coverage at these depths samples the average velocity but does not allow detection of lateral discontinuities. On the basis of velocity values, the underplate appears to grade horizontally into the mantle northward.

The *Pn* phases from the SW are very strong and their amplitude-distance behavior can be modeled accordingly by waves refracted by a strong gradient, rather than by head waves at an interface. Such a gradient would be consistent with the model discussed earlier, in which mantle material has been altered to lower velocity in the intermediate-velocity layer and the more so at its top. In such a model involving melt advection, the increase of velocity with time, by cooling or crystallization of molten intrusions, would restore the lower velocity of the underplate to that of mantle, with, however, a possible variation due to the composition of the material advected. Such an evolution with time would be mirrored in space along the hotspot track and could thus account for the variation observed between SW and NE of La Réunion from a positive velocity gradient with depth under the crust on one side to no such gradient on the other. In such an interpretation, the underplate could be regarded as closely related to active volcanism and vanishing with time, so that no underplate would be distinguishable under the older parts of the hotspot track. If the underplating were absent on the transect of the old Hawaiian chain near Oahu, as advocated by *Lindwall* [1988] in his reassessment of seismic data, which challenges the interpretation of a large volume of underplating [*Watts et al.*, 1985], this would agree with the model of a transient structure.

In a model in which the cooling of the underplate brings it back to mantle velocity values, it may be difficult to admit as a chance occurrence that this transition is located right under the island. The previous tentative interpretation is based on the two common assumptions of a significant drift of the plate and of a continuous discharge of the plume, which lead us to view the variation in space along the section as a variation with time. From sampling the region with several seismic profiles, *Charvis et al.* [this issue] constrain the size and shape of structures and discuss alternative plate/hotspot models.

5. Conclusions

A section through the edifice topped by La Réunion volcanic island documents that it is made of material in which seismic velocity, and hence likely density, is significantly smaller than in the oceanic crust, except for a region confined beneath a quarter of its emerged part. The load of the edifice on the preexisting oceanic crust is therefore smaller than expected in the current assumption that their densities are similar. A related observation is that the oceanic plate is only slightly depressed, if at all.

From its high velocity, the limited size body within the edifice above the regional sea bottom depth may represent its intrusive core surrounded by eruptive products. The small proportion of intrusives with respect to extrusives then appears in marked contrast with the island of Hawaii, under which intrusives dominate and have a large vertical extent and beneath which is a correspondingly large flexural depression of the plate [*Hill and Zucca*, 1987].

Clear multipathing calls for the existence of an intermediate-velocity layer between crust and mantle under the part of the transect southwest of the island. It is located beneath the base of the crust, which is preserved with approximately the same thickness and velocity structure along the whole transect. The intermediate velocity in this layer can be interpreted in terms of in situ modification due to pressure or temperature change but is more likely related to intrusion of material from depth.

The occurrence of a subcrustal layer of intruded material spatially related to presently active island volcanism has not been reported before to our knowledge. The situation may differ from the Hawaiian one, where a very large volume of underplate is inferred under older parts of the island chain by *Watts et al.* [1985], but disputed by *Lindwall* [1987]. Under the active volcano of Hawaii, no underplate is suggested by the seismic data [*Zucca and Hill*, 1987] or assumed in the model of *ten Brink and Brocher* [1987]. Because of the lack of plate flexure observed beneath La Réunion, the trapping of material under the crust is not likely to be controlled by flexural stress. Intrusion of material in the form of melt migrating from depth may then principally be controlled by density. The ponding of these melts under the crust suggests that their composition may differ from that of the tholeiitic basalts erupted. The crust-mantle intermediate layer could hence be the site where primary picritic melts are stored, and crystal fractionation will extract the tholeiitic basalts to be extruded on top of the plate [*Albarède et al.*, 1997].

Such a model of hotspot volcanism has been discussed by

Farnetani et al. [1996] in order to explain intermediate-velocity underplated bodies now seismically seen to underlie the crust over ancient hotspot provinces. However, the composition of the material advected cannot be simply derived since the seismic velocity observed would then depend on the proportion of the new material intruded into the mantle to form the underplate. Furthermore, since La Réunion is volcanically active, part of the material in the underplate may still be molten, and the observed seismic velocity depends on its proportion. In the currently assumed plate drift over the hotspot to the NE, the observation that the intermediate-velocity layer is no longer identified northeast of the island suggests that the underplate is a transient structure with its seismic velocity controlled by the melt fraction. Cooling could raise velocity with time, downstream of the hotspot.

Acknowledgments. Support by the authors' institutions and by the European Community Volcanic Risk programme under contract EV5V-CT92-0188 is acknowledged. IFRTP and IFREMER are thanked for the technical and financial support to the cruise MD-76 REUSIS. We also acknowledge the participation on land, in addition to the contractors, of the personnel and equipment of University of Copenhagen, Denmark (H. Thybo); Universita Trieste (R. Nicolich), CNR Milano, Italy (M. Demartin), ING Roma, Italy (L. Selvaggi): and Servei Geologic de Catalunya (P. Martinez, T. Teixidó). Michel Semet kindly read an early version and identified shortcomings that we tried to improve. We benefitted from reviews by A. Levander, J. McBride, and anonymous reviewers. UMR 6526 (Géosciences Azur) contribution 214.

References

Albarède, F., B. Luais, G. Fitton, M. Semet, E. Kaminski, B. G. J. Upton, P. Bachèlery, and J.-L. Cheminée, The geochemical regime of Piton de la Fournaise volcano (Réunion Island) during the last 530,000 years, *J. Petrol.*, *38*(2), 171–201, 1997.

Bonneville, A., J. P. Barriot, and R. Bayer, Evidence from geoid data of a hot spot origin for the southern Mascarene Plateau and Mascarene Islands (Indian Ocean), *J. Geophys. Res.*, *93*, 4199–4212, 1988.

Caress, D. W., M. McNutt, R. S. Detrick, and J. C. Mutter, Seismic imaging of hotspot-related crustal underplating beneath the Marquesas Islands, *Nature*, *373*, 600–603, 1995.

Charvis, P., Y. Hello, W. O'Brien, S. Operto, B. Pontoise, B. Toussaint, B. de Voogd, A. Hirn, J. Dañobeitia, and the MD76 Scientific Party, Deep structure of a hot spot volcano from seismic refraction data (La Réunion Island): European Geophysical Society General Assembly, Grenoble, *Ann. Geophys.*, *12*, suppl., C36, 1994.

Charvis, P., A. Laesanpura, J. Gallart, A. Hirn, J.-C. Lépine, B. de Voogd, T. A. Minshull, Y. Hello, and B. Pontoise, Spatial distribution of hotspot material added into the lithosphere under La Réunion from wide-angle seismic data, *J. Geophys. Res.*, this issue.

Chevallier, L., and P. Bachèlery, Evolution structurale du volcan actif du Piton de la Fournaise, Ile de la Réunion, Océan Indien Occidental, *Bull. Volcanol.*, *44*, 723–741, 1981.

Detrick, R. S., and S. T. Crough, Island subsidence, hot spots, and the lithosphere, *J. Geophys. Res.*, *83*, 1236–1244, 1978.

de Voogd, B., S. Pou Palomé, A. Hirn, P. Charvis, J. Gallart, D. Rousset, J. Danobeitia, and H. Perroud, Vertical movements and material transport during hotspot activity: Seismic reflection profiling offshore La Réunion, *J. Geophys. Res.*, this issue.

Driad, L., Structure profonde de l'Édifice volcanique de la Réunion (océan Indien) par sismique réfraction grand-angle, thèse, 179 pp., Univ. de Paris 7, Paris, 1997.

Duncan, R. A., Hotspots in the southern ocean: An absolute frame of reference for motion of the Gondwana continents, *Tectonophysics*, *74*, 29–42, 1981.

Farnetani, C. G., and M. A. Richards, Numerical investigations of the mantle plume initiation model for flood basalt events, *J. Geophys. Res.*, *99*, 13813–13833, 1994.

Farnetani, C. G., M. A. Richards, and M. S. Ghiorso, Petrological models of magma evolution and deep crustal structure beneath hotspots and flood basalt provinces, *Earth Planet. Sci. Lett.*, *143*(1–4), 81–94, 1996.

Hill, D. P., and J. J. Zucca, Geophysical constraints on the structure of Kilauea and Mauna Loa volcanoes and some implications for seismomagmatic processes, *U.S. Geol. Surv. Prof. Pap.*, *1350*, 903–917, 1987.

Holbrook, W. S., Underplating over hotspots, *Nature*, *373*, 559, 1995.

Lesquer, A., Structure profonde de l'île de la Réunion d'après l'étude des anomalies gravimétriques, in *Le Volcanisme de la Réunion*, edited by J. F. Lénat, pp. 19–27, Cent. de Rec. Volcanol., France, 1990.

Lindwall, D. A., A two-dimensional seismic investigation of crustal structure under the Hawaiian islands near Oahu and Kauai, *J. Geophys. Res.*, *93*, 12107–12122, 1988.

Malengreau, B., Structure profonde de La Réunion d'après les donnés magnétiques et gravimétriques, thèse, 366 pp., Univ. Blaise Pascal, Clermont-Ferrand, France, 1995.

McKenzie, D., and M. J. Bickle, The volume and composition of melt generated by extension of lithosphere, *J. Petrol.*, *29*, 625–679, 1988.

Nakamura, Y., P. L. Donoho, P. H. Roper, and P. McPherson, Large-offset seismic surveying ocean-bottom seismographs and air guns: Instrumentation and field technique, *Geophysics*, *52*, 1601–1611, 1987.

Phipps Morgan, J., W. J. Morgan, and E. Price, Hotspot melting generates both hotspot volcanism and a hotspot swell, *J. Geophys. Res.*, *100*, 8045–8062, 1995.

Schlich, R., Structure et age de l'océan Indien occidental, *Mem. Hors Ser. Soc. Geol. Fr.*, *6*, 102 pp., 1975.

Schlich, R., The Indian Ocean: Aseismic ridges, spreading centers, and oceanic basins, in *The Ocean Basins and Margins*, vol. 6, *The Indian Ocean*, edited by A. E. M. Nairn and F. G. Stehli, pp. 51–147, Plenum, New York, 1982.

Schlich, R., J. Dyment, and M. Munschy, Structure and age of the Mascarene and Madagascar basins, paper presented at Colloque International Volcanisme Intraplaque: Le Point Chaud de la Réunion, Inst. de Phys. du Globe, Paris, 1990.

ten Brink, U. S., and T. M. Brocher, Multichannel seismic evidence for a subcrustal intrusive complex under Oahu and a model for Hawaiian volcanism, *J. Geophys. Res.*, *92*, 13687–13707, 1987.

Watts, A. B., and U. S. ten Brink, Crustal structure, flexure, and subsidence history of the Hawaiian Islands, *J. Geophys. Res.*, *94*, 10473–10500, 1989.

Watts, A. B., U. S. ten Brink, P. Buhl, and T. M. Brocher, A multichannel seismic study of lithosphere flexure across the Hawaiian-Emperor seamount chain, *Nature*, *315*, 105–111, 1985.

White, R. S., Melt production rates in mantle plumes, *Philos. Trans. R. Soc. London, Ser. A*, *342*, 137–153, 1993.

White, R., and D. McKenzie, Magmatism at rift zones: The generation of volcanic continental margins and flood basalts, *J. Geophys. Res.*, *94*, 7685–7729, 1989.

Wolfe, C. J., M. K. McNutt, and R. S. Detrick, The Marquesas archipelagic apron: Seismic stratigraphy and implications for volcano growth, mass wasting, and crustal underplating, *J. Geophys. Res.*, *99*, 13591–13608, 1994.

Zelt, C. A., and R. B. Smith, Seismic traveltime inversion for 2D crustal velocity structure, *Geophys. J. Int.*, *108*, 16–34, 1992.

P. Charvis, UMR Géosciences Azur (6526), ORSTOM, BP 48, 06235 Villefranche-sur-Mer cedex, France. (charvis@obs-vlfr.fr)

B. de Voogd, Laboratoire de Géophysique, Université de Pau et des Pays de l'Adour, F-64000 Pau, France. (devoogd@univ-pau.fr)

J. Diaz and J. Gallart, Institut de Ciencies de la Terra, CSIC, Soli i Sabaris s/n, E-08028 Barcelona, Spain. (jgallart@ija.csic.es)

L. Driad, A. Hirn, and M. Sapin, IPG, Laboratoire de Sismologie Expérimentale, UA 195 CNRS and Observatories Volcanologiques, 4 Place Jussieu Boite 89, F-75252 Paris cedex 05, France (hirn@ipgp.jussieu.fr)

M. Sachpazi, National Observatory Athens, Seismological Institute, POB 20048, Athens 11810, Greece.

(Received October 13, 1997; revised July 13, 1998; accepted August 25, 1998.)