IL NUOVO CIMENTO VOL. 21 C, N. 5 Settembre-Ottobre 1998

The ScS precursors for the study of the lowermost mantle(*)

M. OLIVIERI and N. A. PINO
Istituto Nazionale di Geofisica - via di Vigna Murata 605, 00143 Roma, Italy

(ricevuto il 20 Agosto 1997; approvato il 5 Ottobre 1998)

Summary. — The exploration of the lowermost-mantle structures by means of body waveform modeling allows the small-scale detection of heterogeneity and anomalous layers. In some regions the D″ layer presents a discontinuity at its top that seems to be a local feature. This anomalous reflector may be recognized by the detection of a small core-reflected phases precursor. These studies may present different order of problems. The main difficulties, are connected to the identification of the precursor and its association to the D″ region. Misunderstandings often result because of phases produced by heterogeneity and anisotropy along and in the vicinity of the ray paths, in the crust and mantle structures. These complexities are increased when large dataset and recording arrays, which may facilitate the waveform analysis, are not available. In this paper we discuss the body waveform modeling of lower-mantle phases for the study of the D″ with particular focus on the case of sparse data with only few events and stations available.

PACS 91.35.Ed – Structure of the Earth's interior below the upper mantle.

1. – Introduction

The large-scale inner Earth dynamical processes are commonly considered to be strictly connected to the structure of the lowermost mantle. In the last decade an increasing number of scientists dedicated their attention to the study of the deep mantle and the core-mantle boundary (CMB), in terms of seismic, thermal and chemical structure. The seismic methods still represent the most powerful techniques to get information on the inner Earth. Most of the current thermo-mechanical models are based on the results obtained by seismic investigations.

In general, the tomographic inversion of large sets of travel time data is able to provide global structural models, but also have a relatively low accuracy (thousands of kilometers) at the CMB (*e.g.* [1]). On the other hand, waveform analysis methods, especially for body waveform modeling, have much higher resolution (Weber [2] suggests that the study of

(*) The authors of this paper have agreed to not receive the proofs for correction.

seismic records observed at seismic arrays and networks allows a resolution which can be about 10 times better than that of tomographic methods) but only give information on small areas of the lowermost mantle.

While the knowledge of the gross structure of the Earth appears to be achieved, an increasing interest in a more detailed structure of the lowermost mantle and the CMB has recently developed. This finer-scale analysis has favored the recognition of small-scale heterogeneity embedded within the large-scale variations provided by tomography (for a detailed review see Loper and Lay [3]).

A particular attention has been addressed to the D″, the 150–300 km thick layer above the CMB firstly suggested by Bullen [4]. Its nature and its origin, a thermal or a chemical boundary layer between the convecting mantle and the liquid outer core, have not been yet clearly explained. The results achieved by using both travel-time tomography and waveform modeling show strong large- and small-scale lateral heterogeneity for this layer. Moreover, a discontinuity in both P and S wave velocity profile at its top has been detected in some regions (*e.g.* [2, 5, 6]) and denied in others [7]. This discontinuity, that commonly reaches 2-3%, has been mainly detected in regions were tomographic models show higher velocity in the lower mantle with respect to the global average [10]. Besides, large-scale low velocities seem to be related to regions in which discontinuities at the top of D″ are not visible but in such regions, as beneath the Central Pacific, a thin (5–40 km) ultra-low velocity layer with a 10% reduction in P-velocity [8, 9] is present above the CMB [10].

Owing to the high lateral resolution of body waveform modeling, some authors, like Schimmel and Paulssen [11] and Garnero and Lay [10], also succeeded in modeling the topography of D″. In particular, their efforts were devoted to overcome the traditional 1D Earth modeling. This result has been recently achieved also by Scherbaum *et al.* [12] that implemented a 3D imaging technique, called Double Beaming Imaging, but this requires source and receiver arrays which are not always available.

Many authors (*e.g.* [2, 3, 10]) suggest that the anomalous D″ is a local feature and its distribution seems to be somehow correlated to subduction zones. The present state of knowledge does not allow the construction of unambiguous dynamical model for this layer,

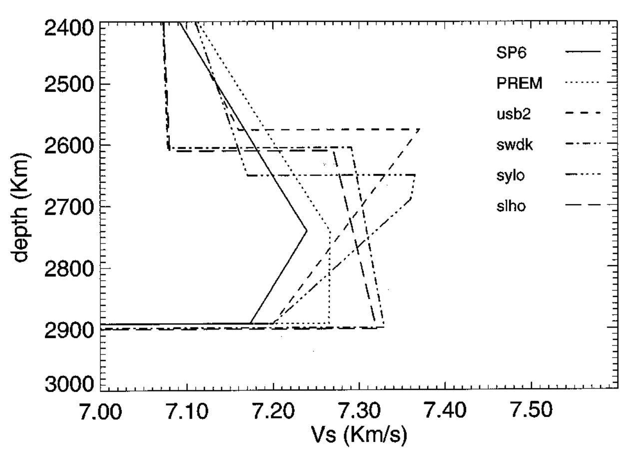

Fig. 1. – Different shear velocity models proposed for anomalous D″ in different regions compared to PREM and SP6: usb2 [20] for the region beneath the South East Pacific basin, swdk [2] for Northern Siberia, sylo [10] for North Pacific and Alaska and slho [5] for Alaska.

which would require extensive punctual studies. Several theoretical hypotheses have been developed about the nature in D″: a chemical [4] or thermal [13] boundary layer, subduction and accumulation at the CMB of oceanic lithosphere [14], phase change due to anomalous lateral temperature gradients [15], boundary layer instabilities and formation of thermal plumes [16], and others. In fig. 1 different shear velocity models proposed for the D″ structure in different regions are presented. The thickness of this layer varies within a range of about 100 km, and the velocity jump is in a range of about 1%. These discrepancies may be consequence of the heterogeneity of the methods used for the waveform modeling and the available datasets, that usually are broad band or WWSSN. But also regional different models for the D″ could be hypothesized and, if verified, this occurrence should by taken into account by the theoretical models.

The aim of this paper is the discussion of the lowermost-mantle investigation by body waveform modeling, in particular the core-reflected phases and their precursors, and to point out problems and difficulties connected to this analysis.

2. – Phases and data

Two are the favorite phases deputed to investigate the lowermost mantle: the core-reflected (PcP and ScS) and the core-diffracted (P_{diff} and S_{diff}). These follow different ray paths and are best visible at different epicentral distances, giving different information about the CMB. PcP and ScS are recognizable on the seismogram at epicentral distance smaller than about 90°, before the superposition with the direct phase. The majority of the studies based on core-reflected precursors uses near-grazing incident phases, which are observed at distances ranging from about 65° to 85°. Near-grazing precursors have higher signal-noise ratio due to the critical reflection and constructive interference of reflected and refracted waves that occurs at the discontinuity because of the increasing velocity through the interface. Schimmel and Paulssen [11] observed a precursor at epicentral distance smaller than 30° in a region in which the D″ reflector was already observed by previous studies, but they pointed out that, for these epicentral distances, the ScS precursor has not an appreciable amplitude on single data.

Core-reflected phases interact with very small patches of the CMB, giving punctual information on the lowermost mantle. In fact, as computed by Weber and Davis [17] and Weber [2], the effective Fresnel zone at the core-mantle boundary for an S wave with dominant period of about 6 seconds observed at a distance of about 80° is an ellipse with axes $3.5° \times 7°$ (about 200×400 km at the CMB) that may be assumed small if compared to the lateral resolution of tomographic studies that usually is not less than thousands of kilometers. On the contrary, the vertical resolution may be limited, because of the uncertainties in the velocity depth function due to the apparently decreasing velocity within the D″ layer (fig. 1), which does not allow energy bottoming at those depths.

Instead P_{diff} and S_{diff} are visible for epicentral distances greater than 90° and have poor lateral resolution being affected by the average structure along large portion of the CMB. Moreover, as shown by Okal and Geller [18], the amplitude of diffracted waves exponentially decreases by increasing the distance from the diffraction surface and this becomes "neglectable" at distances of about one wavelength. Then these phases only sample the last 35–45 km above the CMB and give detailed information on this very thin region.

The small-distance range of observation for these phases strongly limits the areas where the core-mantle boundary can be studied, depending on the available couple source/stations. In fact, this yields important outcomes for the global modeling of D″.

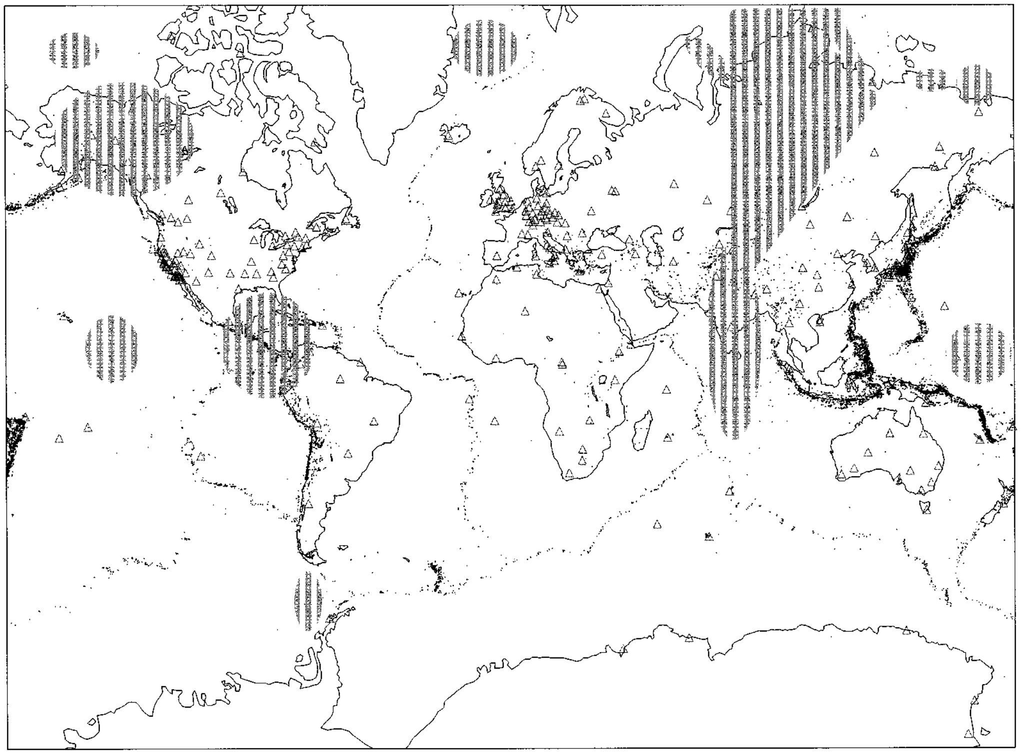

Fig. 2. – Worldwide map including both the location of all seismographic stations affiliated to FDSN [19] and the global seismicity since 1977 with magnitude greater than 5.5 (from the Harvard CMT catalog). The hatched areas represent the regions in which the D'' anomalous reflector has been detected as reported by Loper and Lay [3] plus the area detected by Olivieri *et al.* [20]. All these evidences have been obtained investigating PcP or ScS, and their precursors, or $P_{\rm diff}$ and $S_{\rm diff}$.

This is evident by looking at the distribution of areas in which the anomalous reflector at the top of D″ has been detected. Figure 2 is a worldwide map including both the location of all seismographic stations affiliated to FDSN [19] and the global seismicity since 1977 with magnitude greater than 5.5 (from the Harvard CMT catalog). The hatched areas represent the regions in which the reflector has been detected as reported by Loper and Lay [3] plus the area detected by Olivieri *et al.* [20], all obtained by investigating PcP or ScS and their precursors, or $P_{\rm diff}$ and $S_{\rm diff}$. The distribution of areas where an anomalous reflector has been detected mainly concentrates in the northern hemisphere and this result appears to be affected by the inhomogeneous coverage of seismic stations. However, the inhomogeneous seismicity distribution implies that not all the CMB may be investigated by using core-reflected phases. Wysession [21] has recently analyzed the global seismicity to point out which are the preferred regions to deploy seismic stations on the Earth surface depending on the phases, and then the region within the Earth, that one wants to investigate. Even if areas with a lack of useful PcP and ScS exist, there are others, like Antarctica, where a large amount of core-reflected phases, coming from different seismic regions, is present but the stations coverage is exiguous.

3. – SdS and the possible misunderstanding

The discontinuity at the top of D″ introduces in the data a precursor of the CMB-reflected phases commonly named as PdP or SdS and firstly detected by Lay and Helmberger [5]. The best modeling may be performed when large datasets, as a large number of events recorded by array or regional network, are available. This permits the usage of stacking techniques that allow the discrimination of different phases and the evaluation of their slowness. Then the possibility of misunderstanding is strongly reduced. However, even when arrays are not available, studies of D″ anomalies by a single-station approach are still possible, but these require some more attention in order to avoid possible errors and phase misinterpretations.

As suggested by Weber [2], the best approach to the D″ modeling is to select data with epicentral distances between 65° and 83°, where PcP and ScS are best visible, with source deeper than 70–80 km and magnitude greater than 5.8. These selection criteria would avoid the superposition of direct and depth phases and give a better signal-noise ratio. But great-magnitude earthquakes may have complexities in the source function, as multiple events, especially events with magnitude much greater than 6.0. Moreover, in deep events coming from subduction zones, anomalous arrivals produced at the slab surface may be present. These complications may usually be identified by analyzing teleseismic records of the same event at smaller distance. This kind of events usually has to be discarded and, especially in regions with worse station coverage or low seismicity, the dataset may become too poor. In this case, when only single-station data are available, different event-station couples sampling the same CMB region, *i.e.* when their Fresnel zones intersect, may be grouped in the same dataset.

We will concentrate our analysis on SH wave field, but the following considerations also

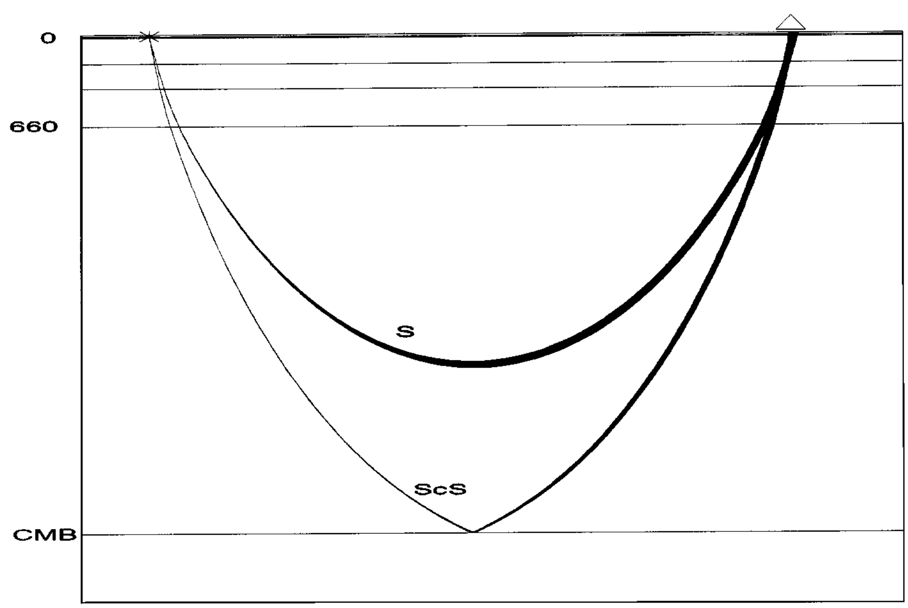

Fig. 3. – S and ScS ray paths traced for PREM at epicentral distance of 78°. The Earth Flattening approximation is applied.

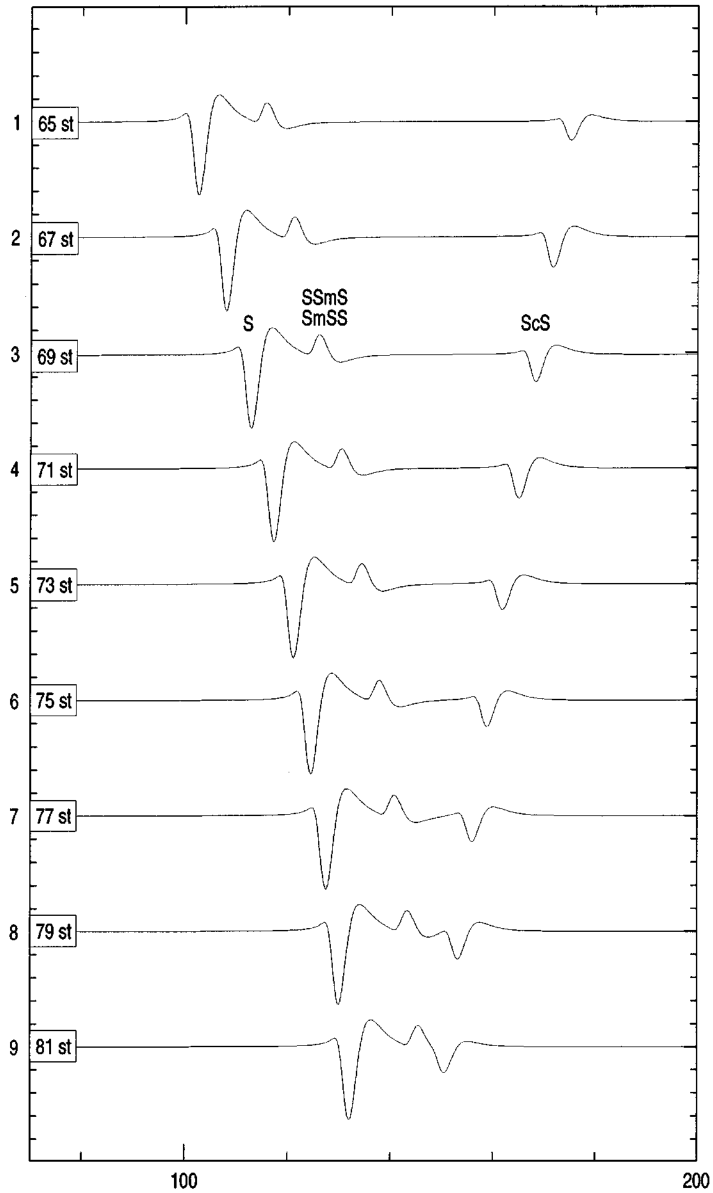

Fig. 4. – SH synthetic profile section computed for PREM fixing the source 10 km deep in the crust.

hold for P and SV wave fields. Even if SH has a worse resolution than P (its lower dominant frequency implies a Fresnel zone two times larger), on the transverse component no P-SV conversions are present then the seismograms are expected to be simpler, and the wider ScS Fresnel zone increases the possibility of grouping different events recorded by different stations.

Most of the difficulties result from the interaction between lower-mantle phases and the structure of crust and upper mantle. In fact, the main criticisms against the results

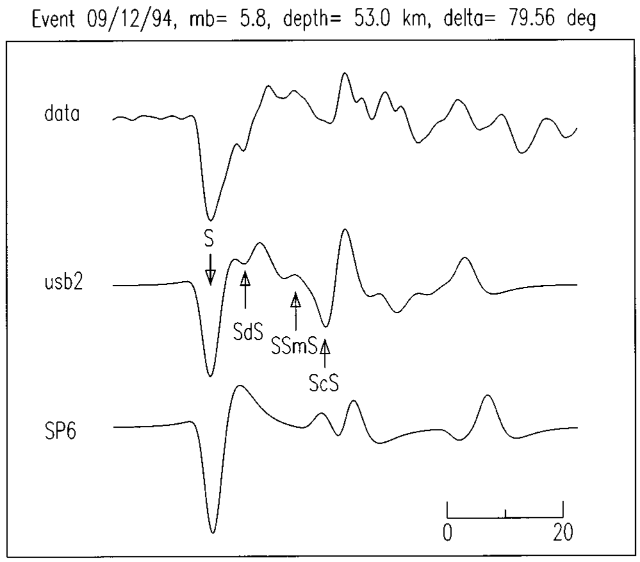

Fig. 5. – Comparison of the transverse component recorded by DRV and synthetic seismograms computed for usb2, and SP6.

obtained by body waveform modeling, pointed out, for instance, by Schlittenhardt *et al.* [7], refer to the distortions caused by source and receiver regions that may perturbate the seismograms and introduce anomalous phases. Indeed, several effects may occur that add strong perturbations to the recorded phases with respect to the theoretical Earth described by the spherically symmetric reference velocity models.

For near-grazing rays, the path of core-reflected phases slightly differs from the direct-phases one whose turning point deepens approaching the shadow zone, approximately starting at 103°. As shown in fig. 3, these phases have nearly identical crust and mantle paths, then possible heterogeneity in the crust and upper mantle would equally affect both phases. Then, if anomalies are detected, these may only be addressed to the region in which the ray paths differ, *i.e.* the lowermost mantle.

The usage of body waveform modeling permits the recognition of anomalies in the travel times and in the amplitude ratio of the main phases with respect to those predicted by a global reference model and it is also helpful in the recognition and modeling of possible unexpected phases. 1D or 2D approaches, that only consider the vertical source-station plane, simplify and speed up the computation of seismograms, but they do not permit the modeling of not-great-circle path energies due to heterogeneity close but out of the source-station plane. As pointed out by Cormier [22], SKS may contaminate the transverse component because heterogeneity out of the source-station plane may deflect energy toward the station. In this case the rotation of the two horizontal components to the transverse and radial ones does not allow the separation of P-SV from SH wave field. In such a case, the apparent azimuth may be determined by an evaluation of the vector slowness, when possible, or by the orientation of the ellipsoid of particle motion.

For deep and intermediate events, in the range 65°–82° only crustal reverberations are expected to be present between S and ScS and, among these, only SSmS, the top-side Moho reflection at the receiver, may reach an appreciable amplitude in the S-coda. For

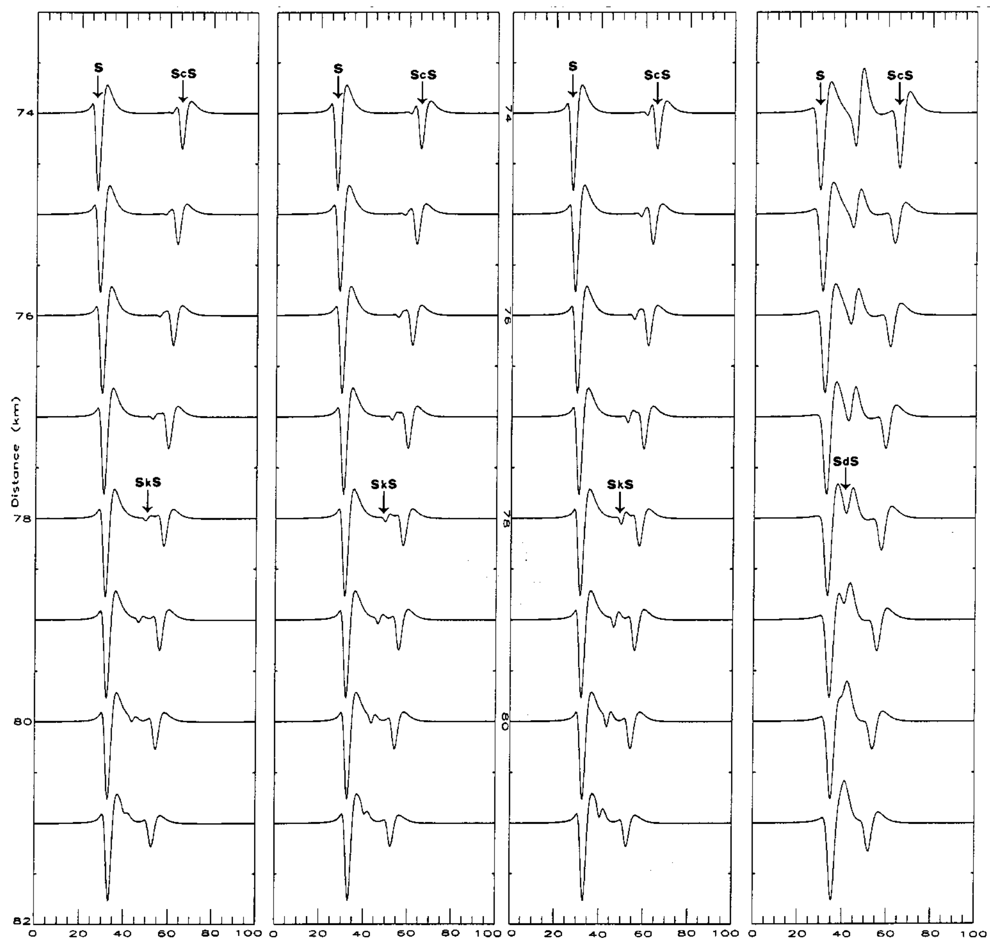

Fig. 6. – SH synthetic profile sections are calculated to display the influence on the seismograms of a possible anisotropy in the upper mantle. From left to right, 20%, 30% and 50% of the SKS amplitude is added to the transverse component. The last section, showing the SH waveforms predicted by usb2 (fig. 1), is reported for comparison.

crustal events, the symmetric phase SmSS is also present and, in particular conditions as when the crust in the source and receiver region has the same thickness and the same average velocity, these two phases have the same travel time and the same polarity then they constructively interfere. The resulting phase displays an amplitude comparable with ScS as shown in fig. 4, where a profile section for the range of distances between 65° and 81° and for a source 10 km deep in the crust is plotted. The synthetic seismograms are generated by the Gaussian beam method [23] for PREM [24]. A misunderstanding is therefore possible, especially in the range 75°–81° in which S and ScS are closer and the eventual precursor travel time would be similar to the Moho multiples. SSmS-S travel time, and SmSS-S when present, is nearly constant as a function of the source depth and of the epicentral distance, then a profile section with all data may be useful to highlight energy with different move-out. When the database contains events with different source

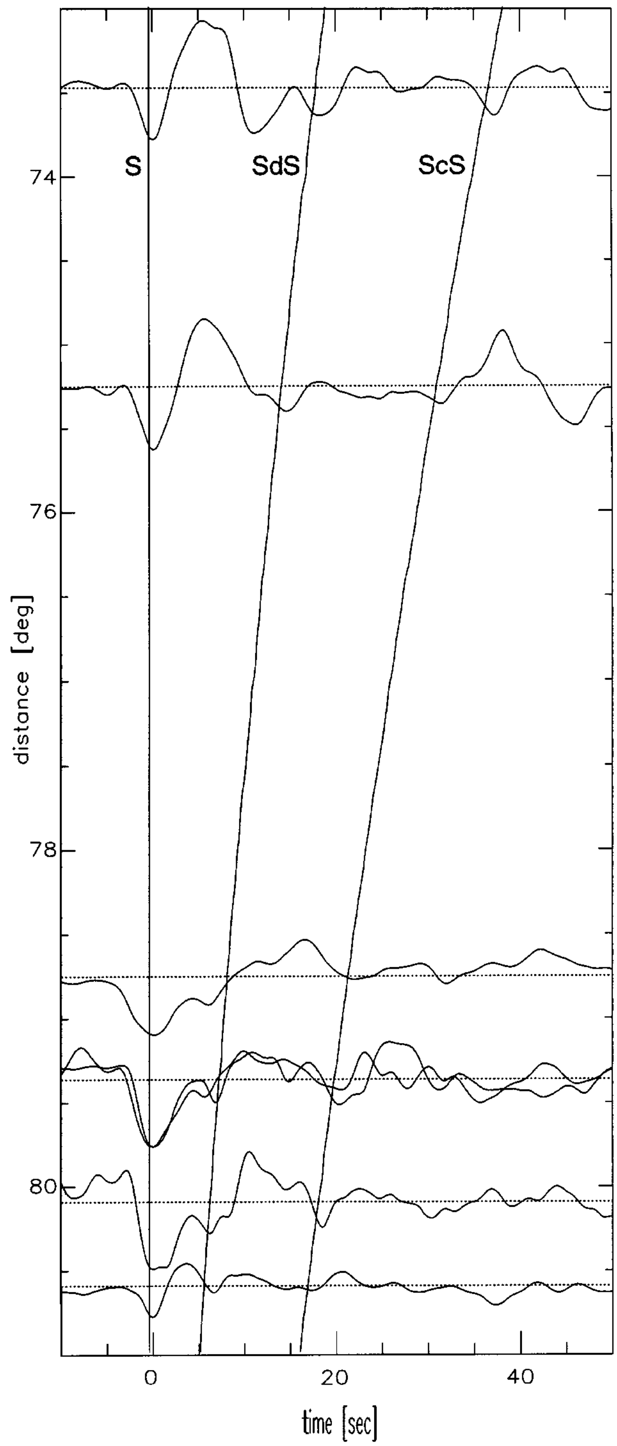

Fig. 7. – Profile section of the transverse component for Chilean Cordillera earthquakes recorded at DRV (Antarctica), data are plotted at the equivalent epicentral distance (see text). Reduced travel time curves are relative to the model usb2.

depth, the record section cannot be constructed and a different strategy must be applied. Only an accurate modeling of both these phases for each event is possible. Especially when the scarcity of deep-source events imposes to use also the crustal ones, different crustal thickness and velocity profile in the source and receiver region may introduce further perturbations.

Figure 5 is a good example of the modeling of CMB anomalies by using single-station data. The transverse component of the event 09/12/94 (Chilean Cordillera) recorded at DRV (Antarctica) is compared with the synthetics computed by usb2 and SP6 [25] where S, ScS, SSmS and SdS energy is included. usb2, the proposed model for the lowermost-mantle region beneath the South East Pacific basin, differs from SP6 by the addition of a 3% discontinuity 310 km above the CMB. It is worthwhile to notice that in this record SSmS is clearly recognizable (the source is 53 km deep so SmSS is not present). The velocity model that best fits data is found by applying a trial and error procedure.

The anisotropy of the upper mantle in the receiver regions may also be a source of problems, because in the range of distance 74°–82° the SKS travel time is enclosed between S and ScS. If anisotropy is present, a part of the SV energy may migrate on the SH and appear in the segment of the seismogram where SdS is usually searched. The amplitude of SKS on the transversal component depends on the energy transferred, this may be as high as 50% of the original SKS amplitude. In fig. 6, three profile sections are generated for PREM together with a profile generated for usb2 [20] in which the reflector at the top of D$''$ is present. In the first three profiles SKS is added to S and ScS on the transversal component and its amplitude is fixed, from left to right, as 20%, 30% and 50% of SKS. It is evident that SKS slowness strongly differs from the slowness of deeply reflected phases, also plotted in this figure, but the possible anisotropy beneath the receiver region should not be neglected when single-station analysis is performed, especially for epicentral distance of about 80°, when its amplitude may be sizable.

As discussed above, a single-station approach does not permit the evaluation of the slowness of different phases, that usually requires stacking techniques. However, when the dataset is made by events whose ray paths insist on the same portion of CMB, a profile section could be arranged overcoming the dependence of the relative travel time, of different phases, on the source depth. Choosing a reference depth for the source and then calculating the theoretical travel time of S and ScS for the proposed model, an equivalent epicentral distance can be retrieved for each event, such as to make the observed ScS-S time coincide with the theoretical one.

Figure 7 shows the displacement proportional profile, obtained by this method to plot profile sections, realized for Chilean Cordillera earthquakes recorded by an Antarctica station (DRV). The SdS phase is clearly recognizable very close to the predicted travel time branch. From this plot SdS slowness can also be evaluated, confirming that this is the product of a deep reflection.

4. – Conclusions

The body waveform modeling is a powerful tool for the study of the lower mantle. But the risk of misunderstanding is always present because difficulties may arise from different situations. In particular, waveform complexities would result from crustal reverberations, source complexities, not-great-circle path energies and anisotropy in the upper mantle. These problems may be easily overcome when arrays or regional networks composed by a large number of stations are available. Even when arrays are not available, a fruitful single-station analysis may be accomplished. This single-station approach is especially suitable in those regions where CMB phases in the useful range of distance exist, but the stations coverage is exiguous. Waiting for more stations and arrays to be deployed, especially in the southern hemisphere, single-station approaches may enlarge and refine the present knowledge of the Core Mantle boundary.

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This work was supported by PNRA (Programma Nazionale di Ricerche in Antartide). We would like to thank M. WEBER, T. LAY and J. SCHWEITZER whose suggestions have favored the realization of this paper.

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