COMBINED TRAVEL-TIME AND AMPLITUDE INTERPRETATION
OF TWO SEISMIC REFRACTION STUDIES IN EUROPE

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ABSTRACT

Refraction seismology is one of the most effective methods to obtain information about the structure of the earth's crust. However, interpretations based on travel-time data alone suffer from a certain degree of non-uniqueness. Synthetic seismogram calculations for a number of theoretical models and the combined travel-time and amplitude evaluation of digitized refraction profiles in southwestern Germany and in the Southern Alps illustrate how information about the amplitude-distance behaviour of the recorded waves can contribute significantly to constraining the range of possible interpretations.

The theoretical model calculations were performed, mainly using the reflectivity method, to examine the effect of various physical parameters on the amplitude-distance behaviour of the wave refracted in the upper-crustal basement (P
g) and of the reflection from the crust-mantle boundary (PMP). This gave a quantitative picture of how the peak in the amplitude-distance curve of the Pg-phase, caused by a positive velocity-depth gradient, is enhanced for higher frequency source signals, and how the end of the gradient zone or the existence of a velocity inversion below it accelerates the amplitude fall-off with distance. While the amplitude-distance behaviour of the PMP-phase can be used as a measure of the abruptness of the crust-mantle boundary itself, it is not sensitive to the structure of the transition zone just above it. However, under favourable circumstances, the signal character of the reflected waves, in particular the amplitude of possible precursors to the PMP, can help distinguish between transitions consisting of a smooth velocity gradient, of a stepwise increase in velocity or of a series of velocity inversions. The synthetic seismogram calculations demonstrated, furthermore, the effect of near-surface structure on the character of the record sections, causing multiply reflected and converted waves, which can mask later arrivals or be misinterpreted as phases from greater depth.

The data from southwestern Germany consists of a 113 km long seismic refraction profile, recorded from a quarry near Sulz a.Neckar, along the eastern margin of the Black Forest. The resulting upper crustal model is characterized by a strong velocity gradient reaching 6.0 to 6.1 km/s at a depth between 8 and 10 km. Sedimentary reverberations are found to mask possible evidence for a discontinuity or a low-velocity layer in the middle crust. However, profiles recorded from Urach, about 50 km to the northeast and crossing the line from Sulz, indicate the absence of a mid-crustal velocity inversion. The reflections from the crust-mantle boundary consist of higher-frequency precursors and a lower-frequency main phase, followed by irregular reverberations. While the latter are probably due to multiple reflections and conversions within the sediments, the amplitudes and the frequency selective character of the former is best explained by a transition zone, between 20 and 26 km depth, consisting of a lamina-like sequence of velocity inversions. This relatively thin crust is an effect of the proximity to the rift structure of the Rhinegraben in the west. The laminated transition zone above the mantle is not an isolated phenomenon, but matches evidence from reflection measurements in the Rhinegraben and below the Urach geothermal anomaly in the Swabian Jura.

Newly digitized and amplitude controlled record sections from the 1977 Southern Alps refraction campaign permitted a reinterpretation of the crustal structure in the area between western Lombardy and the Giudicaria
fault. The resulting model exhibits considerable lateral heterogeneity: in the west, below 7.5 km of sediments of the Lombardy Basin, the crust reaches a depth of only 31 km, while it thickens towards the more mountainous area in the east, reaching a depth of 46 km below the Adamello Massif. Although the signal character of the corresponding reflections is somewhat erratic, the data is satisfied best by models with a low-velocity zone in the upper crust. An additional small velocity discontinuity from 6.2 to 6.4 km/s was found in the middle crust at around 20 km. Earlier interpretations, based on travel-times alone, found a layer with high velocities of about 7 km/s at this depth. This was interpreted as lower-crustal material of the Adriatic-African plate, which had been overthrust onto the European one during Alpine orogeny, thus explaining the uplift of the Southern Alps. However, such a model of crustal doubling is questionable, since a high-velocity layer of that kind is not in agreement with the amplitude data. The unusually thin crystalline part of the crust under the Lombardy Basin is interpreted, in accordance with geological evidence, as a relic of a Late Hercynian rifting phase.
ZUSAMMENFASSUNG

Refraktionsseismik ist eine der erfolgreichsten Methoden zur Erforschung der Erdkrustenstruktur; doch in der Praxis sind Auswertungen, welche allein auf Laufzeituntersuchungen beruhen, mit einer unvermeidlichen Mehrdeutigkeit behaftet. Theoretische Modellrechnungen sowie die kombinierte Laufzeiten- und Amplituden-Auswertung von Refraktionsprofilen in Suedwestdeutschland und in den Suedalpen, zeigen wie quantitative Betrachtungen des Amplituden-Distanz-Verhaltens der seismischen Wellen einen entscheidenden Beitrag zur Eingrenzung des Bereichs möglicher Modelle liefern können.


1. INTRODUCTION

Over the last twenty years, the various disciplines of the earth sciences, in particular geology, petrology and geophysics, have very successfully joined forces in an effort to arrive at a unified picture of the structure and evolution of the earth. This has led to a very elegant theory under the well-known heading of "plate tectonics", providing a framework that links a countless number of diverse observations from different regions and tectonic settings into a global model. Having definitely abandoned everything resembling a static earth, this model attempts to explain geological phenomena in terms of a dynamic process by which continents break up, move about and collide with each other, while ocean basins open and close, and mountain ranges are formed and disappear. Though the driving forces of these motions are far from being understood, and while there remain numerous paradoxes and unexplained phenomena (see e.g. Anderson 1984), it is a remarkably successful theory, which has contributed significantly to a unified picture on a global scale and has given the earth sciences unprecedented impulses (see e.g. the Final Reports of the International Geodynamics Project, AGU/GSA 1980-1983).

Fig. 1.1 Epicenters for the period 1970-1980, outlining the plate boundaries between North America, Africa and Eurasia (from Mueller 1984, after Wanleik et al. 1982). The inset represents the contemporary seismotectonic scheme for Central and Northwestern Europe.
Earthquakes all over the world are a dramatic expression of the dynamic nature of the earth. The large majority of epicenters delineate the boundaries along which the rigid plates forming the earth's lithosphere move relative to each other. The epicenter map in Figure 1.1 illustrates this very clearly. The earthquakes aligned neatly along the Mid-Atlantic Ridge are associated with sea floor spreading, which has caused the Atlantic Ocean to open in Mesozoic times and which is still active today. The higher spreading rate of 41 mm/year in the South Atlantic, compared to 18 mm/year in the North Atlantic, induces a counterclockwise rotation of the African plate. As a consequence of this rotation, the boundary between the African and European continents, as delineated by the complex pattern of earthquakes in Figure 1.1, corresponds in part to shear motion, as in the region of Gibraltar and northern Africa, and in part to compression, as in the Alpine collision zone. Much of the seismic activity associated with this collision is concentrated around the so-called "Adriatic Promontory", which is the northernmost tip of Africa pushing against the European plate. The ensuing stresses are propagated across the interior of the plate, leading to intraplate seismicity in zones of crustal weakness. Typical zones of weakness within a continental plate are recently active rift zones, such as the Upper Rhinegraben and the Lower Rhine Embayment, which are also clearly delineated by epicenters in Figure 1.1.

Continental rift structures and the orogens associated with continental collisions are features of particular interest, since insight into their present structure can give valuable clues to the understanding of the evolution of the earth. The choice of the two areas investigated in this study was in part governed by this interest in tectonically active regions and they are thus not typical examples of an undisturbed continental crust. The first area lies in the transition zone between the northern Alpine foreland and the previously active but now dormant rift system of the Upper Rhinegraben. The second, situated in the Southern Alps, also shows evidence for the existence of an ancient rift system or continental margin, which, however, as a consequence of continental collision, has been overprinted by the effects of the Alpine mountain building process.

One of the key inputs to the plate tectonics model is, of course, detailed information regarding the physical structure of the earth's interior and its regional variations. Only a combination of all geophysical disciplines, such as seismology, gravity, geomagnetism and geothermics, can clarify the puzzling results provided by each discipline alone. In an effort to delineate changes in material property, both laterally and vertically, seismology, which measures the velocity of propagation of seismic waves and thus the material dependent elastic properties in the earth, has been extremely successful. Refraction seismology, in particular, still is one of the most efficient and economical methods to obtain information about the structure of the earth's crust (i.e. the upper 30 to 50 km) over a large area (see e.g. the monograph, Explosion Seismology in Central Europe, edited by Giese et al. 1976).

The first step in the interpretation of crustal seismic refraction experiments consists of modelling the travel times of the recorded phases.
Based on previous experience, the interpreter identifies and correlates the various refractions and reflections from the interior of the crust and from the crust-mantle boundary, and then either enters the measured arrival times into an inversion program which calculates a possible velocity-depth model, or, assuming a starting model, calculates travel time curves for a variety of models in a trial-and-error procedure until a sufficiently good match between calculated and observed travel times has been achieved. In tectonically more complex areas with strong lateral heterogeneities, where several overlapping and reversed profiles are available, this interpretation is refined by modelling the arrival times with two-dimensional ray-tracing techniques.

Given a sufficiently large amount of closely spaced and carefully recorded data, this procedure should lead to an accurate picture of the crustal structure. However, in practice, data are usually far from ideal, being marred by either man-made or natural disturbances. Due to logistical or economic considerations, experiments are often performed in areas where complex geological features near the surface produce secondary effects which mask the primary phases needed for the interpretation of the deeper crustal structure. In addition, the presence of several reflectors within the crust produces waves which interfere with each other in such a way that it is often difficult to identify and correlate them with certainty in the record section.

Seismologists usually agree fairly well on the gross structure revealed by a given seismic experiment. However, several joint workshops, during which both theoretical and experimental record sections were interpreted by different persons, showed that travel-time interpretations alone often do not produce unique results: a variety of models can be made to satisfy the data and the interpretational and methodological bias of the interpreter has a strong influence on the results (Mooney and Prodehl 1980, Mooney 1981, Ansorge et al. 1982, Finlayson and Ansorge 1984).

The structure of a medium traversed by seismic waves influences not only the travel times of the various phases in a seismogram (kinematics) but also the respective amplitudes (dynamics). It is therefore quite obvious that one should attempt to constrain the range of possible interpretations that satisfy a given data set by modelling the amplitudes as well. In order to accomplish this, certain conditions must be fulfilled by the recording procedure as well as by the interpretational techniques.

The dynamic behaviour of elastic waves has been the subject of theoretical investigations for a long time, but only recently, with the advent of high-speed computers, has it been possible to calculate and plot synthetic seismograms to model the effect of complex crustal structures. At first, these synthetic seismograms were used only for qualitative comparisons with the observed data, or, at most, for modelling the amplitude ratios of different phases within the same seismogram. Only through the use of large numbers of instruments with matching recording characteristics and by computer processing the signals after analog-to-digital conversion, was it possible to achieve proper control over the amplitudes, allowing a quantitative interpretation along an entire profile.
Thus the procedure to employ for the interpretation of crustal seismic refraction data is to first model the travel times and then to refine the model by matching the measured amplitude-distance behaviour of the various phases with synthetic seismograms (see e.g. Mueller 1977). Using such a procedure it is possible to tackle several problems of geophysical interest.

Two of the goals of refraction seismology are the identification of the material at depth and the determination of its physical state, that is pressure, temperature, degree of fracturing and possible fluid inclusions. Laboratory measurements indicate that these parameters strongly influence the velocity-depth gradient, to which the amplitudes of refracted and diving waves are very sensitive.

Of tectonic significance is furthermore the question of the existence of low-velocity layers, corresponding to magmatic intrusions or changes in the metamorphic grade, or of high-velocity inclusions, such as slivers of mantle material, within the crust. Given a sufficient velocity contrast relative to the material above and below, such structures strongly influence the amplitude-distance behaviour of reflected waves.

A further question which has been the subject of intense debate over the last two decades is that of the nature of the crust-mantle transition. Among the models considered are a single large velocity jump (first-order discontinuity), a smooth gradient, a series of step-like velocity increases or a lamina-like structure of velocity inversions. A careful analysis of the signal character of the waves reflected from the crust-mantle boundary allows one to distinguish between such models.

The first part of the present study is devoted to some theoretical considerations regarding the effect of various features of crustal models upon synthetic seismograms and to the possibility of distinguishing between different models on the basis of amplitude behaviour. The second part is a brief description of the recording, digitizing and signal processing procedures used. The third and fourth parts present the interpretation of a profile along the eastern margin of the Black Forest in southern Germany (Sulz-south) and the reinterpretation of a set of profiles recorded in the Southern Alps (SUDALP 77), using both travel time and amplitude information. The Sulz-south profile provides insight into the velocity-depth gradient of the upper crystalline crust and into the fine structure of the crust-mantle boundary in the vicinity of the Rhinegraben. Lateral variations in crustal structure under the Southern Alps profiles, are evidence for the existence of an ancient (pre-Jurassic) rift system under the Lombardy basin and amplitude data suggest a revised interpretation of the structure of the middle and lower crust in the area directly involved in the Alpine orogeny.

Parts of the theoretical considerations concerning the amplitude-distance behaviour of the Pg-phase (the first arrival travelling through the upper part of the crystalline basement) and parts of the results from the Sulz-south profile have been published in two papers by Banda, Deichmann, Braile and Ansorge (1982) and by Deichmann and Ansorge (1983).
2. METHODOLOGY

2.1 Definition of terms and brief historical review

In the context of modelling the dynamic features of seismic waves, one must distinguish between amplitude and complete waveform modelling. While in the first case one is concerned only with the amplitude level of individual phases without regard to the signal shape, in the second case one attempts to synthetically reproduce the complete seismogram including all secondary and converted phases. Both of these types of modelling can be either qualitative, being limited to a mere visual comparison between recorded and calculated seismograms, or quantitative, which involves numerical fitting of the measured amplitude levels or actually bringing recorded and synthetic signals to match as closely as possible. Quantitative waveform modelling in refraction work has so far only been attempted in studies of the oceanic crust, which is laterally more homogeneous and where the near-surface geology is less complex than on continents (see the review by Spudich and Orcutt 1980).

The amplitudes used for a quantitative interpretation can be either relative or absolute. In the first case, only the amplitude ratios between two phases of the same signal are used. This precludes the necessity for equalization of shot size or calibration of instruments. In the second case, the amplitude-distance behaviour of individual phases over the whole record section is considered, thus entailing the proper calibration of shots and instruments. Of course, this too is only absolute in a restricted sense since calibration and subsequent matching of the calculated amplitudes is always relative to some shot whose size is arbitrary. For a complete model, the absolute amplitudes of individual phases must also have the correct ratio to each other, but the advantage of this method over the first is that the amplitude values must not necessarily pertain to the same seismograms.

This study is concerned mainly with the quantitative evaluation of absolute amplitudes, making use of amplitude ratios where feasible, and to some qualitative discussions of the influence of structural features on signal form.

Among the first to point out the possibilities and to show examples of amplitude interpretations in refraction seismology were Russian researchers. Their work, performed in the middle and late sixties, is summarized in two monographs, translated into English (Davydova 1972, Galkin 1972) and in two papers by Davydova et al. (1970, 1972).

The theoretical foundations for a large part of the amplitude interpretation to follow were established by Červený (1966) from a ray-theoretical point of view and by Fuchs (1968, 1970) on the basis of wave theory. Early systematic evaluations of absolute amplitudes to constrain their crustal models better were published by Helmberger and Morris (1969)
and by Hill (1971). More recent investigations interpreting the absolute amplitude-distance behaviour in a manner similar to that employed in this study include work by Berckhemer et al. (1975), Müller and Fuchs (1976), Braile (1977), Müller and Mueller (1979), Banda and Ansorge (1980) and Braile et al. (1982). Publications by Olsen et al. (1979), McMechan and Mooney (1980) and Zucca (1984) are examples of the use of relative amplitude ratios.

Fig. 2.1 Example of synthetic record-section calculated with the reflectivity method with a dominant frequency of 4 Hz for model PG11K displayed in the left inset. The amplitude-distance curve (solid line) is shown in the right inset. Pg denotes the first arrival, which corresponds to the wave refracted in the crystalline basement. Note the prominent secondary phases: Pw - "whispering gallery" wave (see Červený et al. 1977), PgS - direct Pg converted to S at the base of the sediments, PgPg - Pg reflected once at the free surface, PgPgS - reflected Pg converted to S.
2.2 Synthetic seismograms

Synthetic seismograms can at present be calculated by a variety of methods. Chapman (1978) and Spudich and Orcutt (1980) published extensive reviews of the most common methods, discussing the merits and limitations of each. This study uses mainly the reflectivity method developed by Fuchs (1968, 1970) and Fuchs and Müller (1971) with the modifications by Kind (1978), and for some special considerations, the asymptotic ray method described by Červený et al. (1977). With the modifications by Kind (1978), which take into account the effect of the free surface and provide the possibility to place the source in the reflectivity zone, the reflectivity method automatically includes all the secondary effects, such as multiply reflected, refracted and/or converted waves, which are an important and often neglected part of a seismogram. This technique requires velocity models consisting of laterally homogeneous and isotropic layers. Velocity gradients are approximated by a stack of thin layers with small velocity contrasts and with thicknesses corresponding to less than one wavelength of the frequency range considered. A summary of the main points of the theory behind the reflectivity method and of the parameters used in the computation of the synthetic seismograms is given in Appendix C, and the calculated models are listed in Appendix D. A typical synthetic record section for a simple upper crustal model calculated with this method is shown in Figure 2.1.

All synthetic record sections in this study are plotted with a reduction velocity of 5.0 km/s, and, in order to compensate for the amplitude decrease due to geometrical spreading, the amplitudes of each trace are multiplied by the distance. The amplitude values plotted on a logarithmic scale as a function of distance correspond to the maximum (peak-to-peak) of each phase.

Effect of frequency content of the source signal

The frequency content of seismic refraction data varies from about 2 Hz to more than 20 Hz depending on the shooting technique, charge size, frequency response of the instrument and local geological environment. The comparison of the shape of the amplitude-distance curves for low-frequency source signals (2 Hz) with those for high frequency (8 Hz) shows significant differences when velocity gradients are truncated at shallow depth or low-velocity layers are present in the model.

Figure 2.2 shows an example of the results for the model in Figure 2.1 computed for 2.5, 4, 6 and 8 Hz dominant frequency. The fact that the velocity structure of the upper crust changes, from a positive gradient (0.077 km/s/km) to zero gradient at 7 km depth produces a different response of the medium depending on the source frequency. The variation of amplitude with distance is more pronounced at the higher frequencies. As discussed in more detail in the paper by Banda et al. (1982), this is due to the fact
Fig. 2.2 Amplitude-distance curves for model PG11K (upper right inset) for frequencies of 2.5, 4, 5 and 8 Hz (Fourier spectra in the lower inset) computed with reflectivity method (continuous lines). Dotted line corresponds to ray-method computation, with program SEIS4.
Fig. 2.3 Comparison of amplitude-distance curves of displacement (continuous line) and velocity (dotted line) seismograms for model PG11KK (upper right inset). The corresponding spectra show the frequency shift between displacement and velocity signals. The bottom insets show the signal (at 35 km) for displacement and velocity; the time intervals used to calculate the spectra are marked by vertical bars.
that shorter wavelengths are mainly affected by the focusing effect of the
gradient zone, thus producing strong relative amplitude maxima at a distance
of 45 km. At the same distance, longer wavelengths are already affected by
the homogeneous layer beneath the gradient zone.

For the same model, amplitudes were computed with program SEIS4, based
on asymptotic ray theory, and are also displayed in Figure 2.2. As was to
be expected, the ray method works reasonably well for high frequencies,
although the slope of the amplitude decay is somewhat enhanced. This is
because the asymptotic ray method represents a "high-frequency
approximation" to the wave equation (Červený et al. 1977; Chapman 1978).
Asymptotic ray theory accounts for the influence of different frequencies on
the amplitudes of waves reflected at first-order discontinuities (Červený et
al. 1977) but not for waves refracted from a gradient zone (Banda 1979).
The ray method is inexpensive to use and for much of the available data and
some crustal models, which can also include lateral inhomogeneities, this
approximation is accurate enough. For more detailed studies, in which the
models can be approximated by flat homogeneous layers, the reflectivity
method is more appropriate. Therefore, if we are dealing with good quality
data, suitable for amplitude studies, it is of fundamental importance to
compute the theoretical seismograms using the reflectivity method, or other
wave theory methods, with a source which has a dominant frequency similar to
that of the experimental data.

Effect of computing ground displacement instead of velocity

The synthetic seismograms in Figure 2.1 represent ground displacements
instead of ground velocities as measured in observed seismograms. As shown
in the example in Figure 2.3, for which both velocity and displacement were
calculated, the difference in the amplitude-distance behaviour is not
significant. However, for more complicated structures and in cases where
several phases interfere with each other, differences between displacement
and velocity can be more pronounced. Moreover, it should be noted that the
dominant frequency used in displacement computations is increased when the
displacement seismograms are differentiated to obtain velocity (see spectra
in Fig. 2.3). Thus, in this study, for direct comparison with recorded data,
synthetics corresponding to ground velocity were calculated.
2.3 Model calculations of Pg-amplitudes

Variation of gradients

It is well known that even a slight positive velocity gradient greatly influences the amplitude of refracted phases (see e.g. Červený 1966, Hill 1971). In order to determine systematically the effect of different gradients, amplitude-distance curves for models with gradients between -0.015 and 0.108 km/s/km were computed and are shown in Figure 2.4. From a comparison with a calculation for model PG11 at higher frequency and from the discussion in the previous section, it can be stated that, although these curves were calculated for 2.5 Hz, they are representative also for higher frequencies as long as the gradient zones extend sufficiently deep. The results shown in Figure 2.4 illustrate that small positive velocity gradients in the upper crust will be resolvable by amplitude measurements on reasonably good experimental data. For example, the amplitude-distance characteristics of Models PG1 and PG6 are significantly different although the velocity structure differs only by the presence of a small (0.038 km/s/km) gradient in Model PG6.

Thickness of the gradient zone and low-velocity layers

Significant differences in the amplitude-distance curves for models having the same gradient but different thickness of the gradient zone are seen in Figure 2.5. The models include structures with a continuous gradient between 2 and 15 km (model PG11), 2 and 10 km (model PG11KK) and between 2 and 7 km (model PG11K) on top of a half space. A decrease in the thickness of the gradient zone leads to a faster drop-off of the amplitudes with distance.

The introduction of a low-velocity layer below a gradient zone between 2 and 7 km and 2 and 10 km (model PG11L and PG11LL) shows another interesting effect. A significant shift of the maximum and a change in the slope of the amplitude decay is evident when a low-velocity layer is present at the same depth at which the gradient is terminated (compare PG11K and PG11L in Fig.2.5).

As discussed in more detail in the paper by Banda et al. (1982), this amplitude-distance behaviour is due to the wave nature of seismic signals and deviates from that predicted by asymptotic ray theory in two respects. From simple ray-theoretical considerations, one would expect amplitude levels to stay high as long as the diving waves turn above the bottom of the gradient zone and to abruptly go to zero as soon as the rays penetrate into the low-velocity zone. In reality, however, the waves behave as if they "sense" the low-velocity zone well before the turning point of the corresponding rays reaches that depth, thus causing the amplitudes to decay.
Fig.2.4 Amplitude-distance curves for models PG12 (negative gradient), PG1 (no gradient) and PG3, PG6, PG11 and PG22 (positive gradients) for a frequency of 2.5 Hz.
Fig. 2.5 Amplitude-distance curves for models PG11, PG11K, PG11KK with variable gradient zone thickness (continuous, dashed and dotted line, respectively, in the lower inset) and PG11L and PG11LL with a low-velocity layer at different depths (dashed and continuous line, respectively, in the upper inset) computed for a frequency of 4 Hz.
Fig. 2.6 Ray amplitudes (top) calculated with program RAYS1 for model PG11, shown in the inset, with three different interface shapes (bottom).
at shorter distances than expected. Conversely, energy continues to arrive at the surface for some distance even after the rays have penetrated the low-velocity zone. This effect is, of course, highly sensitive to the wavelengths of the signals used, and stresses once again the necessity of computing synthetic seismograms with a frequency similar to that of the recorded data.

Effect of layer curvature

As mentioned before, the calculations based on the reflectivity method require models with flat horizontal layers, including a correction for the earth's curvature (the so called "earth flattening approximation", see e.g. Müller 1977b). In order to estimate the error introduced by this simplification into the amplitude calculations in the case of curved interfaces, several test-models were calculated using program RAYS1, written by V. Červený and I. Pšenčík (Praha), which, based on asymptotic ray theory, allows for laterally inhomogeneous models (Červený et al. 1977; Červený 1979). Figure 2.6 shows the amplitude-distance curves for the Pg-phase, computed by this method for a flat-layered model (PG11) and for two models with a curved sediment-basement boundary. The discrepancies between these curves and those calculated with the reflectivity method in Figure 2.4 are due to the inability of asymptotic ray theory to account for wave effects occurring at the top and bottom of the gradient zone. However, asymptotic ray theory correctly accounts for geometrical spreading, so that the effects of different model geometries on the amplitudes can be compared with each other.

From this comparison, it can be concluded that, in the presence of a sufficiently strong gradient, the diving waves will be affected only insignificantly by such an upwarp of the basement.
2.4 Model calculations of PMP-amplitudes

Davydova (1972) classified possible models of the crust-mantle transition, known as the Mohorovičić discontinuity or just Moho, into three different types: first-order discontinuities, transition zones with smoothly or stepwise increasing velocities, and transition zones consisting of lamina-like velocity inversions (Fig. 2.7). Her conclusions indicate that a detailed analysis of the dynamic properties of the compressional waves reflected from the Moho, designated as PMP, can provide criteria for distinguishing between the different types (see also Davydoval et al. 1970, 1972).

Fuchs (1970) as well as Braile and Smith (1975) presented several synthetic record sections for various models of the Moho, illustrating qualitatively the effect of structure on amplitudes. Figure 2.8 shows the amplitude-distance curves for several models of the Moho, calculated in the course of this study to enable a direct and quantitative comparison between them.

![Diagram of crust-mantle transitions](image_url)

Fig. 2.7 The main types of crust-mantle transitions: I—first-order boundaries; II—step-like or smooth transition layers; III—laminated transition zones; h = thickness of transition zone; M = Mohorovičić discontinuity; (from Davydova et al. 1972).
The models consist of a 6.0 km/s crust with a transition zone starting at 20 km depth and reaching 8.0 km/s at 25 km, which corresponds to the Moho. Such a thin crust is not typical for a normal continental crust, but corresponds more closely to a continental rift structure. It was chosen for these calculations in view of modelling the crustal structure below the Sulz-south profile. In addition, a thinner crust allows one to reduce the distance over which seismograms need to be calculated and thus also to reduce computational costs. The results, however, are equally applicable to a thicker crust if one corrects for the appropriate time and distance shift of the arrivals.

Fig. 2.8 PMP amplitude-distance curves (vertical component, ground displacement, 8 Hz dominant frequency) for various models of crust-mantle transitions. The corresponding seismograms are shown in Fig. 2.9.
Fig. 2.9 Synthetic seismograms of the PMP-phase (vertical component, ground displacement, amplitudes multiplied by distance) for various models of the lower crust. The transition zone extends from a depth of 20 to 25 km, and the velocity increases from 6.0 to 8.0 km/s in all models. The phase with an apparent velocity of 5.5 km/s, visible in the lower part of each quadrant, is a numerical effect corresponding to the lower limit of the phase velocity window used in the reflectivity program.
The amplitude values always correspond to the maximum of the whole reflected wave group. The amplitude fall-off from the peak in the curves towards longer distances is practically the same for all five models. Even the small peak at 95 km in the curve of model 9, due to the reflection from the top velocity jump, is not significant when matching measured amplitudes, which usually exhibit a scatter of about a factor of 2 (see Section 3.4). This means that the amplitudes of the wide-angle reflection or diving wave from the transition zone above the Moho are not very sensitive to the detailed structure of that zone. The amplitude decay towards shorter distances, on the other hand, is significantly sharper for models 4 and 10 than for models 2 and 9. This is due to the larger velocity jump at the Moho itself in the latter two models, which produces stronger subcritical reflections. The laminations in model 11 have a similar effect of increasing the amplitudes as compared to the simple steps in model 10, but it is doubtful whether it is sufficient to be detectable in real data. In order to illustrate the effect of structure on the form of the reflected signal, the relevant portions of the synthetic record sections are shown in Figure 2.9.

Effect of a step-wise velocity increase

As shown by Fuchs (1968), the behaviour of a velocity gradient can be approximated by a stack of thin layers with stepwise increasing velocities. If, however, these layers are not thin enough relative to the wavelength of the incident signal, and if the velocity contrast between each layer is too strong, they will generate individual reflections. The faint precursors in the first seismograms of model 2 in Figure 2.9 are due to subcritical reflections from the individual layers, while the large amplitude phase corresponds to the reflection from the velocity jump at the bottom of the transition zone. The amplitude of the precursors relative to the main phase will increase as the number of steps in the gradient zone is decreased, but, at the same time, the reverberation-like character observed in the data is lost. This is illustrated by models 5 and 9 in Figure 2.9. By extending the gradient zone all the way to the mantle without a larger velocity jump at the Moho, as in model 10, the intermediate velocity jumps are more pronounced than in model 2, which increases the amplitude of the reverberations. The velocity contrast and consequently also the precursor amplitudes will be even larger if the individual steps include velocity inversions.

Effect of a laminated velocity transition

Reflections from a transition zone consisting of a series of lamellae with alternating high and low velocities have been investigated theoretically by Fuchs (1968, 1969, 1970), who demonstrated the dependence
Fig. 2.10 Synthetic seismograms (vertical component, ground displacement, amplitudes multiplied by distance) calculated for a model consisting of a 2 km thick sedimentary layer, with \( V_p = 4.5 \) km/s, over a basement with \( V_p = 6.0 \) km/s and a mantle with \( V_p = 8.0 \) km/s at 23 km depth. The first phase corresponds to PMP, while all subsequent arrivals are multiple reflections and conversions of the PMP within the sediments. Each letter of the phase identification denotes one leg of the ray paths in the sedimentary layer. Note the phase reversals of the reflections at the surface.
of the signal character both on frequency and on angle of incidence. A structure in which the velocity of the lamellae increases with depth will produce reverberations whose amplitudes increase gradually. This is illustrated by models 8 and 11 in Figure 2.9. The amplitude of the precursor relative to the main reflection depends on the size of the individual velocity jumps. The synthetic seismograms, corresponding to model 11, were also calculated for a signal with a dominant frequency of 4 Hz instead of 8 Hz (Fig. 2.9). At lower frequencies, the precursors, corresponding to subcritical reflections from the lamellae, are significantly attenuated, and the whole transition zone appears more like a smooth gradient, producing a single strong phase whose amplitude decreases rapidly towards shorter distances. A similar frequency-dependent behaviour will, of course, also be produced by a step model without inversions, such as model 10 in Figure 2.9 (see Fuchs 1968). However, as can be seen from a comparison between the seismograms of models 10 and 11, the amplitudes of the precursor reflections from a transition zone without velocity inversions are significantly smaller than from a laminated structure.

The reflections shown in the synthetic seismograms so far all end very abruptly right after the phase of maximum amplitude. In real data, however, these are often followed by some kind of a ringing signal. Fuchs (1970) showed how such a feature could be produced by a laminated upper mantle. In the presence of a surficial sedimentary layer, an alternative explanation for these reverberations is illustrated by the synthetic seismograms in Figure 2.10. In this case, the phases following the main PMP-reflection are entirely due to multiple reflections and conversions of the PMP within the sedimentary layer both beneath the shotpoint and the receivers. Gane et al. (1956) postulated this effect to explain the almost complete obliteration of S-wave arrivals by a group of irregular reverberations on earthquake seismograms that had travelled through the 8 km thick Witwatersrand sedimentary basin in South Africa. Assuming a sedimentary layer which is thinner than the 2 km chosen for this example and whose thickness is not constant over the entire profile, the individual arrivals will merge and interfere with each other, thus producing an irregular pattern of vibrations, similar to those observed in the data. This shows that reverberations from an upper mantle lamellation could be at least partially masked by this sedimentary effect. Additional crustal discontinuities will, of course, increase these interference phenomena even further.
3. DATA ACQUISITION AND PROCESSING

3.1 Recording procedure

All of the data from the Sulz-south and Southern Alps experiments discussed in this study were recorded on FM-magnetic tape instruments of the MARS type (Berckhemer 1970). The signal input was provided either by a single three-component Mark-L4 seismometer or by three single component seismometers of the FS-60 type, both with a natural frequency of 2 Hz. Except for the instruments of the University of Karlsruhe, the two seismometers were matched to give the same output of about 1 Volt/cm/s. Timing was accomplished by recording the coded DCF radio-transmitted time signal along with the shots. The three analog signals, corresponding to the three components of ground velocity, and the time signal are modulated onto separate carrier frequencies between 0.86 and 9.6 kHz, which are mixed together with a constant pilot frequency of 6.4 kHz, and recorded onto a single track of a magnetic tape reel or cassette.

Upon playback, the four carrier frequencies and the pilot are demultiplexed by narrow bandpass filters and the signals are demodulated to produce a maximum analog output of ±1 Volt. These can either be viewed on a paper chart recorder, or transferred via an analog-to-digital converter onto a computer for further processing and plotting.
3.2 Digitizing procedure

Until the advent of generally available medium sized computers a few years ago, record sections were constructed by playing back the seismograms on a paper chart recorder and then, after aligning each trace with the aid of the recorded time signal, the signals were redrawn by hand onto transparent paper. In addition to the tediousness of the procedure, the disadvantages were numerous. Since alignment was based on one or two arbitrarily chosen second marks close to the signal onset and since the time scale was subject to the fluctuations of both the tape and paper chart speed, timing accuracy was low. The sensitivity of the paper chart recorder was usually adjusted by eye to produce a clearly visible seismogram, without regard to the absolute amplitudes, thus discarding a priori a valuable part of the information. Moreover, once the record section was constructed, any changes of the time and distance scale or of the filter settings could only be accomplished by starting all over again with the playback.

Having each signal available on computer in digital form allows one to apply any desired filter algorithm and to plot any number of record sections with different time and distance scales, in order to enhance various features of the signals. At the same time, by noting the recording instrument's gain and keeping track of all processing steps, it is possible to assign an absolute amplitude value in terms of ground motion to each digital value. To accomplish this, each digitized seismogram must be preceded by a header record containing all the necessary information for its identification and of its processing history. In addition, the onset time as well as the sampling rate must be determined with sufficient accuracy. The details of the hard- and software in use at the Institute of Geophysics of the ETH-Zuerich are described in Appendix B. The following is intended as a general explanation of the motivation for the procedure used.

The process consists of eight steps:

1. transfer of data from A/D-converter to disc
2. insertion of seismogram information into header record
3. determination of exact sampling rate
4. decoding of time signal
5. control plot of digitized sequence
6. removal of possible spikes
7. anti-alias filter and reduction of sampling rate
8. transfer of data from disc to digital tape for final plotting of record sections.
In crustal refraction experiments the usable signal rarely contains much energy above 20 Hz, so that a sampling rate of 100 Hz, corresponding to a Nyquist frequency of 50 Hz is ample sufficient. Nevertheless, to obtain high-quality data, it is very important to digitize the seismograms at a somewhat higher rate, usually 400 Hz, and to reduce it to 100 Hz only after the onset time has been determined and all spikes have been removed. This is necessary for several reasons. The output from the demodulator usually contains additional noise, both seismic and electronic, with frequencies above 50 Hz, which would be a source of aliasing. In addition, the signals sometimes contain sharp spikes, caused by defects of the recording system or by disturbances of the carrier frequencies. The danger of aliasing could be avoided by inserting an analog low-pass filter between demodulator and converter. However, aside from introducing a further undesirable phase shift into the signal, any existing spikes would be broadened and smoothed, making them much more difficult to remove, or, in extreme cases, making them indistinguishable from the seismic signal and thus falsifying the record. Moreover the accuracy of the time determination is partly a function of the rate at which the signal is sampled, so that the error becomes larger when using lower sampling frequencies.

In order to avoid the influence of tape speed fluctuations during recording and play-back, the clock of the digitizer is provided by the recorded pilot signal. The corresponding sinusoidal wave of 6.4 kHz is reduced to 400 Hz and shaped to produce a sequence of sharp pulses, which trigger the digitizer. Possible interruptions of the pilot, which would lead to gaps in the digitized record, are bridged by an additional oscillator in a phase-locked loop circuit. Although the pilot frequency is quartz-controlled, and can therefore be regarded as constant over the length of a seismogram, it usually deviates by a few Hz from the given value of 6.4 kHz. To be able to assign an absolute time value to each sample, the exact sampling rate must be known. As explained in greater detail in Appendix B, this is calculated in step 5, by applying a stacking algorithm to the beginning and end of the digitized time signal. The fractional part of the second corresponding to the time of the 0-th digit, i.e. the onset time of the digitized sequence, is also determined from the stacked time signal. Through the stacking procedure, the signal-to-noise ratio of the time signal is considerably enhanced, and small fluctuations of the receiver are averaged out, thus increasing the accuracy of the time determination as compared to the manual procedure of using only one or two arbitrarily chosen second marks. After automatically decoding the time signal for date, hour, minute and whole second in step 4, the complete information is stored in the header record.

A control output of all the header information and a plot of the seismograms, time signal and the two time stacks allows one to check for mistakes and serves as a permanent record of the whole processing procedure, which can be returned to in case of uncertainties during the interpretation phase. This raw plot, including only that interval actually containing the signal of interest, is also necessary to check for possible spikes. These are then removed one by one in step 6 with an interactive program on the graphics terminal.
At this point the seismograms can be filtered in the time domain with a zero-phase anti-aliasing filter and the sampling rate is reduced to 100 Hz. Up to this stage, all four channels are stored in multiplex form on a single file. During the filtering process, the four channels are separated, the time signal is discarded, since all the necessary time information is contained in the header record, and the individual components are written onto three separate files.

Finally these three files are stored on magnetic tape, from which then the record sections are plotted. During plotting, an additional recursive, zero-phase band-pass filter can be applied to the data, to enhance various features of the signals.

The system as described here was developed on the basis of the one implemented by W. Kaminski at the Geophysical Institute of the University of Karlsruhe. There, however, steps 1 and 2 are performed by digitizing a whole sequence of recordings and copying them directly onto tape. Once a tape is full, the remaining steps are performed one by one, from one tape to another, for all the signals on the tape. In the system used here, on the other hand, each recording, consisting of the three components of ground motion plus the time signal, are subjected individually to the whole procedure and written onto tape only in the final form. Thus the next seismogram is digitized only after the previous one has been completely processed and stored on tape. This system takes better advantage of the available hardware (see Appendix B for details) and allows a more immediate control of the results, since preliminary record sections can be plotted after each digitizing session. For crustal refraction experiments, with seismograms usually less than a minute long, it is also very efficient, since the whole procedure does not take more time than what is needed to position the analog tape for the next recording (about 10 to 15 minutes per recording). For very long seismograms the filtering and reduction step (number 7) would be more efficient in batch mode, which, however, requires a very large amount of disc space or two tape drives.
Fig. 3.1 Histogram showing the deviations from the mean arrival time for the test shot in the Snake River Plain.
3.3 Accuracy of travel-time measurements

The accuracy of travel-time determinations is influenced by the following factors:

1. picking of the onsets
2. variations in the system characteristics, i.e. the electronic specifications, of recording instruments and receivers
3. determination of time of 0-th digit and of sampling rate during digitizing
4. station locations
5. shot time and shot location
6. elevation corrections.

In the summer of 1980, at the end of an extended seismic refraction campaign in the Snake River Plain, U.S.A., all the employed instruments were grouped in one spot to record the same test shot 1.4 km away. The 95 kg of explosive were fired in a 15 m deep borehole in dry basalt. The signals were processed in the standard way, and both travel-times and amplitudes of the first arrivals were measured from records of the vertical components. A total of 4 instruments from the University of Stuttgart, 10 from the University of Karlsruhe and 15 from the ETH-Zuerich, all equipped with Mark-L4-3D seismometers, were involved in the test.

Figure 3.1 shows a histogram of the deviations from the mean arrival time. No systematic deviation could be found to account for the apparent bimodal distribution around the mean, so that it was attributed to the incompleteness of the data set and a Gaussian distribution was assumed. The standard deviation computed from the 25 usable arrivals amounts to ±22 ms. Though the number of readings is too small for a meaningful statistical analysis, this can be regarded as a usable uncertainty estimate for the combined factors 1, 2 and 3. It must, however, be borne in mind that the onsets in this example were very sharp, so that the uncertainty of the onset determinations was lower than for normal data, which are recorded over larger distances and in the presence of higher background noise.

Errors in station locations, caused by inaccurate maps or by carelessness in marking the positions on the maps are equivalent to an additional timing error: on a record section with a time scale reduced with a velocity of 6.0 km/s, an error of 60 m in direction of the profile will be hardly detectable as a shift in distance, but will cause a deviation of 0.01 s in the arrival time. Since the actual error is dependent on the quality of the maps used and on the care taken by the operator in the field, it is difficult to give a generally valid estimate of its size.
Inaccuracy of shot time and location, causes more of a systematic error than the previous ones. Its effect is to add a constant time error to all arrivals from a particular shot. This causes a velocity error only for the layer immediately below the surface, and changes the depth of the interfaces, but does not affect the deeper layers' velocities. Nevertheless, it can be a source of error which must be taken into consideration when dealing with profiles recorded piecewise from several shots, since then, depending on the distribution of stations with respect to the explosions, it may be a cause of discrepancies between the different data sets. Usually, however, sufficient care is taken when locating the shotpoints on the map, and time breaks are measured by the same person with the same instrument, so that the actual errors are not significant.

The last factor in the list actually is one which only comes into play during the interpretational stage, but which, nevertheless, must be kept in mind. Along profiles with large topographic variations, it may often be necessary to correct the arrival times for elevation differences among the stations. Poor knowledge of the surface layer velocity can be a further source of significant timing errors when applying topographic corrections. The size of this error is, of course, proportional to the elevation difference itself, and thus impossible to estimate a priori. This problem is discussed in more detail in Appendix A.

In summary, the uncertainty of ±22 ms is to be considered as a lowermost bound, valid for good quality signals with a frequency content around 10 Hz, to which one must add the other factors, dependent on the actual circumstances valid for the data set under consideration. Thus, an arrival time accuracy better than 30 ms seems unrealistic with the present recording equipment and processing procedure.
3.4 Accuracy of amplitude measurements

The same data set used for the travel-time test was also used to estimate the experimentally caused uncertainty of the amplitude determinations. This was done by measuring the peak-to-peak amplitude of the first arrivals on a true amplitude plot. As mentioned in Section 3.1, the sensitivity of the instruments from Karlsruhe and Zuerich was obviously different, so that the measured amplitudes were examined separately. Since only two of the instruments of the University of Stuttgart had usable records, they were not included in this comparison. The results are summarised below:

<table>
<thead>
<tr>
<th>INSTRUMENTS</th>
<th>NUMBER USED</th>
<th>AVERAGE AMPLITUDE</th>
<th>STANDARD DEVIATION</th>
<th>NORMALIZED</th>
</tr>
</thead>
<tbody>
<tr>
<td>Karlsruhe</td>
<td>8</td>
<td>11.9 mm</td>
<td>1.0 mm</td>
<td>8%</td>
</tr>
<tr>
<td>Zuerich</td>
<td>15</td>
<td>4.5 mm</td>
<td>0.7 mm</td>
<td>16%</td>
</tr>
</tbody>
</table>

Ratio of sensitivity Karlsruhe/Zuerich = 11.9/4.5 = 2.64

The resulting errors include the influence from the following factors:

1. measurement of the amplitudes on the plot
2. differences in seismometer sensitivity
3. differences in recording amplifier gains
4. effects of digitizing and processing.

The last point can actually be neglected, since its influence is insignificant relative to that of the others.

The signal character of the different records was very similar except for the obvious polarity reversals (the instruments from Karlsruhe and Stuttgart which functioned properly had a downward first arrival, while all but one of those from Zuerich had an upward onset, in agreement with the standard conventions for vertical component seismometers).

The better relative accuracy of the instruments from Karlsruhe is probably not due to lower instrumental scatter, but rather to the smaller relative error incurred when measuring larger amplitudes, which were a consequence of the higher instrument sensitivity.

Thus one can say that amplitude scatter due to different instrument characteristics or to measurement errors amounts to no more than 20%. On the other hand, the additional scatter due to differences in geophone-ground coupling and in site geology from one station to another is very difficult to estimate. Until systematic investigations of this problem are available, it is thus only possible to argue that, based on theoretical seismogram calculations, the amplitude-distance behaviour, due to the deeper and nearly...
horizontally layered structures which one actually wants to model, is a smoothly varying function, and that consequently the large amplitude jumps from one station to the next, visible in the data, are due to near-surface station effects. The examination of amplitude data from clearly identifiable phases, such as Pg or PMP, indicates that, in general, amplitudes can jump by more than a factor of two from one station to the next, but that, given sufficient recording points, the overall shape of the amplitude-distance curve expected from the theoretical calculations is reproduced very well (see Figs. 4.5, 4.9 and 5.13).
4. PROFILE SULZ-SOUTH

4.1 Introduction

As mentioned in Chapter 1, the first data set under consideration in this study was obtained in southwestern Germany, along a transition zone between the northern Alpine foreland and the continental rift structure of the Upper Rhinegraben. Figure 4.1, after Ahorner (1975), shows schematically the tectonic evolution of this area. Rifting began about 45 to 40 million years ago, in a region weakened by the updoming of mantle material and subjected to a SW-NE oriented compressive stress, associated with the northward push of Africa against the European continent (Illies 1975). After a brief clockwise rotation of the stress field some 20 million years ago, the counterclockwise rotation of Africa has led to the presently dominant SE-NW direction of maximum compressive stress, which manifests itself in predominantly strike-slip mechanisms of the earthquakes beneath the Swabian Jura and Upper Rhinegraben (see e.g. the recent compilation by Bonjer et al. 1984).

Fig. 4.1 Two stages in the evolution of the Rhinegraben rift system under the influence of a rotation of the stress field caused by changes in the relative motion between Africa and Europe (after Ahorner 1975). URG = Upper Rhinegraben; LRG = Lower Rhinegraben (or Lower Rhine Embayment); HGZ = Hessian Graben Zone.
Fig. 4.2 Geological map with shotpoint (star) and station locations (dots) of profile Sulz-south. Stations are numbered consecutively from north to south. Information regarding the crust-mantle transition applies to the range between stations 10 and 20.
The updoming of mantle material that initiated the rifting process also
induced the uplift of the Vosges and the Black Forest, which form the
present day graben flanks. As a consequence, the surrounding area is
characterized by an anomalously thin crust, typical for continental rift
structures.

The southern Rhinegraben has been studied extensively in the past both
by seismic reflection methods (Dohr 1968) and refraction methods (Mueller et
these results was published by Prodehl et al. (1976). Recently, Zucca
(1984) reinterpreted the travel times along two of Edel's refraction
profiles with the aid of two-dimensional ray tracing techniques and modelled
amplitude ratios with synthetic seismograms calculated with the reflectivity
method. The crustal structure beneath the South German Molasse Basin is
quite well known on the basis of early reflection measurements by Liebacher
(1962, 1964) and from the interpretation of several refraction profiles by
Emter (1971, 1976). Using earthquake data from various azimuths,
Peterschmitt (1979) derived a generalized crustal structure for the Swabian
Jura, while the geothermal anomaly of Urach has recently been the object of
intensive research, which included structural studies with the aid of
reflection and refraction measurements (Bartelsen et al. 1982; Jentsch et

The refraction profile Sulz-south fills a gap between the Upper
Rhinegraben and the Black Forest to the west and the Swabian Jura to the
east; as such it provides information about the transiton structure in the
vicinity of a rift structure active in Cretaceous to Tertiary times
(Fig.4.2). Using both travel-time and amplitude information, it was
possible to investigate the velocity gradient of the upper crystalline
basement, the problem of the possible existence of a low-velocity layer in
the upper or middle crust and the nature of the crust-mantle transition.
Fig. 4.3 a) Trace-normalized, vertical component record section, with 32-Hz anti-aliasing filter. Numbers above each trace indicate station and shot.

4.2 Geological setting

As can be seen from the locations of the recording sites in Figure 4.2, the first 90 km of this 113 km long, north-south trending line are situated along the edge of the crystalline Black Forest, where it dips beneath the Triassic sediments. Beyond 90 km, the line enters the Swiss Molasse Basin and crosses the eastern end of the Swiss Jura. From borehole data at Sulz and at other locations along the profile compiled in the geological literature (Breyer 1956; Buechi et al. 1965; Lemcke et al. 1968; Boigk and Schöneich 1968; Schneider 1980), it is possible to construct a rough cross-section of the sedimentary structure and of the topography of the crystalline basement (Fig.4.3b). Thus the sediment-basement boundary is characterized by an up-dip over the first 40 km and a down-dip beyond about 60 km. In the north, the surficial upper Triassic sediments (Keuper and Muschelkalk) are underlain by a thick wedge of lower Triassic and Permian deposits (Buntsandstein and Rotliegendes). In the south, these Triassic sediments are covered by a layer of Jurassic limestones and by two Molasse basins, which are separated by the Jurassic outcrop of the Laegeren, the easternmost part of the folded Jura mountains.

4.3 Acquisition and presentation of the data

The data were obtained between 1974 and 1980, from ten explosions in a single quarry near the town of Sulz am Neckar. Because of poor recordings, some sites were occupied more than once, and two of the shots, which, unlike the others, were not instantaneous but fired with multiply delayed charges, were not evaluated at all. All shots were recorded on FM-magnetic tape instruments of the MARS type (see Section 3.1 and Berckhemer 1970). Except for three stations with FS-60 seismometers (nos.4.5, 6, 7 in Figs.4.2 and 4.3), all instruments were equipped with three-component MARK L-4 seismometers.

The records were digitized and processed as described in the previous chapter. Figure 4.3a presents the vertical component data, filtered only with the 32-Hz anti-aliasing filter. The amplitudes are trace-normalized, but gain factors, which are calculated for each seismogram by the plot program, allow the determination of absolute ground velocity.

The most striking features of these data are a clear Pg-phase as first arrival out to 90 km and, beyond 55 km, strong PMP-reflections from the crust-mantle boundary. Between these two arrivals, there seem to be additional phases in the seismogram, but they are difficult to correlate over more than two or three consecutive records (Fig.4.3). Clearly visible in the upper left corner of Figure 4.3 is also a strong Sg-wave, which is shown again together with the SMS in the transverse component record section in Figure 4.10.
Fig. 4.4  

a) Vertical component, trace-normalized record section, band-pass filtered 8-16 Hz.  
b) Elevation corrected P$_g$-arrivals with travel-time curves of models 9 (continuous line), 12 (dotted line) and 13 (dashed line). See the velocity-depth models in the insets of Fig. 4.5.  
c) Ray-trace of the P$_g$ phase for model 9.
4.4 Upper crust

Since the records presented in Figure 4.3 are trace-normalized, the amplitudes of the first arrivals beyond about 70 km appear unduely reduced, due to the larger PMP-phases. In addition, beyond about 90 km, the records are severely disturbed by low-frequency noise. For these reasons, the first two seconds (reduced with 6 km/s) of the vertical component record section are replotted with a band-pass filter of 8-16 Hz (Fig.4.4a). Since band-pass filtering tends to distort impulsive onsets like those of the Pg-phase at shorter distances, arrival times were picked from the unfiltered record section wherever possible, and are plotted with elevation corrections in Figure 4.4b. Station elevations range between 405 and 780 m above sea level (Fig.4.3b), so that the Pg-arrival-times displayed in Figure 4.4b were reduced to 600 m a.s.l., corresponding roughly to the elevation of the shots. The elevation corrections were calculated as discussed in Appendix A, and amount to 0.035 s or less.

Taking into account the uncertainty in picking the onsets, the errors in time determination based on the digital decoding of the recorded DCF time signal, the uncertainty of the elevation corrections, the effect of uncertainties in the station locations and in the shot times, most of the arrival times should be accurate to ±0.05 s. This is indicated by the length of the vertical bars plotted for each arrival in Figure 4.4b. At distances beyond 80 km, due to the poorer signal-to-noise ratio, the uncertainty is greater, so that in some cases onsets were not picked at all. The early arrival at 50 km is probably due to the fact that this station was placed to the east of the general trend of the profile, almost directly on the crystalline rocks of the Black Forest (station 19 in Fig.4.2).

From the information presented in the geological cross-section (Fig.4.3b) and from the decrease in apparent velocity beyond 70 km, it is evident that lateral variations in velocity and/or thickness of the sedimentary cover preclude the use of either the Wiechert-Herglotz or the tau-p method to model the basement structure. Instead, a ray-tracing program developed by Gebranede (1976) was employed.

Lacking detailed velocity measurements along the profile and considering the low station density, no attempt was made to model the local travel-time anomalies limited to a few consecutive records. The published sedimentary P-wave velocities from refraction and borehole measurements in southern Germany and northern Switzerland scatter over wide ranges, depending not only on sediment type but also on location and depth (Breyer 1956, Rybach 1962, Lohr 1967, 1969). Thus, in constructing the ray-trace models, the sedimentary structure was simplified to a homogeneous surface layer with a constant velocity between 4 and 5 km/s, over a low-velocity wedge below the shotpoint, corresponding to the Triassic and Permian sediments shown in Figure 4.3b (Breyer 1956; Emter 1971). The shape and depth of the sediment-basement boundary was kept fixed, while the velocities above and below were adjusted in a trial-and-error procedure. The best of the resulting ray-trace models is shown in Figure 4.4c, and the corresponding travel-time curve in Figure 4.4b. The large delays beyond 90
Fig. 4.5 a and b) Amplitude-distance data of the Pg-phase (crosses with shot number above and station number below), with the theoretical amplitude-distance curves, calculated with the reflectivity method, corresponding to the models in the insets. The top insets show only the enlarged portion of the gradient zones. The black dots correspond to the level of the background noise for each trace. The length of the vertical bar on the right side of each figure indicates the amount of scatter, equal to a factor of 2 in both directions.
km will be discussed in more detail further on.

The peak-to-peak amplitudes of the first cycle, measured from the record section in Figure 4.4a and multiplied by the corresponding scale factor to obtain ground velocity values, are plotted as a function of distance in Figure 4.5. As shown by the black dots corresponding to the background noise level for each trace, beyond 90 km where the profile crosses into the Swiss Molasse Basin, the signal-to-noise ratio decreases significantly. The vertical size of the crosses marking the data points corresponds roughly to the instrumental uncertainty, while the additional scatter, amounting to about a factor of 2, must be attributed to differences in local site response (see Section 3.4). The values from stations number 6 and 25 deviate by a factor of 3 to 4 from their neighbouring points, which is indicative of faulty gain settings. The charge size of shot number 1 was only 700 kg, while that of the other shots was about 2000 kg, so that a correction factor had to be applied. The investigation of several quarries by Burkhardt and Vees (1976) indicates that it is not possible to establish a generally valid scaling law between signal amplitudes and charge weight for quarry blasts. Since the stations recording shot number 1 were interspaced between those recording shot number 10, and thus covered the same distance range, the correction could be determined from the average amplitudes for each shot. The resulting factor of 1.4 turns out to be about equal to the cube root of the charge ratio. This corresponds roughly to the lower bound of the range of amplitude-charge relations found in the literature (see a compilation of data by Mueller at al. 1962). Though the remaining scatter is quite large, the data show a significant local amplitude maximum at a distance of about 40 km, which is characteristic for a strong velocity gradient in the upper crust.

Since, as discussed in Chapter 2, only a wave-theoretical method can account correctly for the amplitude decay of waves refracted near the bottom of a gradient zone, the synthetic seismograms were calculated with the reflectivity method. The curvature of the sediment-basement boundary, as depicted in Figures 4.3b and 4.4c, is almost identical to the weaker of the two examples in Figure 2.6. As shown in Section 2.3, this curvature has only a negligible effect on the amplitude-distance behaviour, so that the flat-layer approximation of the reflectivity method is valid for modelling the Sulz data. Since the absolute source energy does not enter into the synthetic seismogram calculation, the offset of the resulting amplitude-distance curves, which are plotted in Figure 4.5, is arbitrary and was adjusted to the data by a least-squares method. The logarithmic standard deviations of data points from these curves range between 0.23 and 0.27, which corresponds to amplitude factors of 1.7 and 1.9, respectively.

Model number 9, which, except for the sediment velocity, is the same as model number 6 in Figure 4.5a, fits the travel-time and amplitude data best. In this model, the crystalline basement is characterized by a zone with a strong velocity gradient (0.074 km/s/km) between 1 and about 6 km depth, followed by a second zone with a weaker gradient (0.02 km/s/km) down to a depth of about 8 km. Models number 12 and 13 were calculated in order to obtain a measure of the possible variations in basement velocity gradients which would be compatible with the data. Usually, several models can be
made to fit the travel-time data by compensating changes in basement velocities with changes in sediment structure. However, in this case, the overall topography of the sediment-basement boundary is fairly well known and for the first 70 km the sediments are less than 1 km thick, so that the travel-time data alone strongly constrain the range of possible velocity gradients in the upper crust. Reducing the upper gradient to less than 0.06 km/s/km (model 12), or increasing it beyond 0.085 km/s/km (model 13) would no longer be compatible with the measured travel times (see Fig. 4.4b). In spite of the large scatter in the data, it is possible to draw the same conclusion from the amplitudes: as the gradient is decreased, the local maximum moves out to larger distances and becomes less pronounced, while an increase of the gradient sharpens the peak and moves it closer to the shot point. Beyond about 70 km, the thickness of the sedimentary layer increases and its velocity is not well known: thus sediment velocities can be varied in order to compensate for travel-time effects due to different velocities in the lower of the two gradient zones mentioned before. As shown by the comparison between models 3 and 4 in Figure 4.5a, a stronger gradient, extending over a sufficiently large depth range, produces a second amplitude increase beyond about 65 km. Models 6 and 7 illustrate the effect of small changes in the extent of a second zone with a slight gradient. From those results it follows that the lower gradient is not greater than 0.04 km/s/km and that the velocity at the bottom of the gradient zone, which is reached at a depth between 8 and 10 km, is less than 6.1 km/s.

As discussed in Section 2.3 and illustrated by model 3L in Figure 4.5, a velocity inversion below the zone with positive gradient significantly accelerates the amplitude decay at large distances. The data points beyond about 80 km would indeed suggest a more rapid decay than that calculated for models with constant velocity below a depth of 8 to 10 km. However, without further calculations, it is difficult to assess the influence of the Molasse basin, which the profile crosses beyond 90 km, on the amplitude behaviour; on the one hand, one would expect an amplitude increase due to the steeper angle of incidence associated with the lower surface velocities, and on the other, a decrease due to higher anelastic attenuation in the softer and thicker sediments. Moreover, as mentioned before, the signal-to-noise ratio becomes rather poor in this distance range, so that these data points cannot be regarded as conclusive evidence. Nevertheless, a velocity inversion at a depth of about 10 km is certainly compatible with the Pg-amplitude data.
4.5 Structure of the middle crust

Looking at refraction data alone, evidence for a velocity discontinuity or inversion in the upper or middle part of the crust should manifest itself as one or two intermediate reflections between the Pg- and PMP-arrivals. Although some of the records from Sulz contain large amplitudes in this interval, phases are very difficult to correlate over more than a few traces (see Fig.4.3). Only in the distance range between 70 and 90 km does there seem to be a coherent arrival at about 0.5 s after the Pg (Fig.4.6a).

In order to explain the travel time of these arrivals, several ray-trace models were calculated, both with a positive velocity discontinuity and with an inversion. A simple positive velocity jump below the gradient zone cannot simultaneously account for the large delay of this phase and for the close distance from the shot. Therefore a low-velocity layer had to be taken into consideration. The most satisfactory fit of the travel time data and of the Pg-amplitudes was obtained with model 3L, which is characterized by a velocity drop from 6.06 to 5.6 km/s at a depth of 9 km and a gradual increase to 6.2 km/s between 11 and 12 km (lower inset in Fig.4.5b). The Pg-amplitude-distance curve of this model is plotted with a dotted line in Figure 4.5b. Note the sudden amplitude decay beyond 80 km as compared to model 3 in Figure 4.5a, due to the velocity inversion below the gradient zone. The synthetic seismograms corresponding to model 3L, in the distance range between 65 and 95 km, are displayed in Figure 4.6b. A comparison of the amplitudes of the reflection from the bottom of the low-velocity layer (denoted by PcP), relative to those of the Pg, between data and synthetics shows a significant discrepancy. While in the calculated seismograms the amplitude ratios of PcP to Pg reaches a value greater than 2 at a distance of 85 km, in the data the maximum amplitude ratio is about 1, and occurs already around 76 km.

Changes in the gradient at the base of the inversion or a decrease of the velocity inside the channel will not simultaneously reduce the amplitudes of the PcP-phase and move the maximum to shorter distances (Braile and Smith 1975; Müller and Mueller 1979). Braile (1977) showed that lowering the Q-value and thus increasing the degree of anelastic attenuation in the channel will significantly decrease the amplitudes of the PcP-reflection. The synthetic seismograms presented so far were all calculated with a constant Q of 500 for the crystalline crust underlying a sedimentary layer with Q equal to 100. The seismograms shown in Figure 4.6c correspond to a model similar to the previous one, except for the velocity and Q-value of the inversion zone, which were lowered to 5.4 km/s and 50, respectively. Under these conditions, the PcP-reflection is so weak that it cannot be distinguished from the multiple surface reflections of the Pg-phase which penetrate into the basement and from P to S conversions at the surface and at the sediment-basement boundary. Though other more realistic Q-values in the range between 50 and 500 are likely to be compatible with the observations, the data are not judged to be distinctive enough to allow a simultaneous determination of both velocity and Q structure in the middle crust. Note also the synthetic seismogram example in Figure 2.1 for a model without any discontinuity or inversion beneath the
Fig. 4.6 a) Vertical component records, trace normalized, 4-16 Hz band-pass filtered, showing possible reflections from the base of an upper-crustal low-velocity layer (PcP).

b) Synthetic records for model 3L (Fig. 4.5) with an assumed Q=100 for the sediments and 500 for the rest of the crust.

c) Synthetic records for model 3L (Fig. 4.5) with Q=50 and Vp=5.4 km/s in the low-velocity zone. PgPg: Pg phase reflected once at the surface. PcP: reflection from the base of the low-velocity layer. Pw: interference head-wave, or "whispering gallery phase" (see Červený et al. 1977). The amplitudes of the synthetic records are multiplied by distance and correspond to ground velocity.
upper crustal gradient zone, in which the multiply reflected and converted phases within the sedimentary layer produce amplitudes similar to those observed here. Accurate modelling would also require a better knowledge of Q in the sediments: anelastic attenuation is likely to influence the intensity of the sedimentary reverberations as well.

Though the synthetic seismograms contain reflections from the upper boundary of the low-velocity layer, which closely follow the Pg-arrivals between 35 and 55 km (see the example in Banda and Deichmann 1983) they also are masked by sedimentary effects. Similarly, the reverberations observed in the data at these distances (Fig. 4.3a) do not exhibit the character of a distinct arrival, so that they cannot be relied upon for a unique interpretation.

Thus, based on the evidence from this data alone, it is not possible to resolve conclusively the problem of the possible existence of a low-velocity layer in the middle crust below the Sulz profile. Resorting, therefore, to the criterion which favours the simpler of those models that are not in conflict with the data, further calculations were performed assuming a constant velocity of 6.0 km/s for the middle crust (model 9 in Fig. 4.5b).

An alternative model, compatible with the Pg-amplitudes, could involve a somewhat higher velocity at the bottom of the upper-crustal gradient zone at around 10 km depth, followed by a layer with a negative velocity gradient reaching all the way down to the lower crust. However, as discussed at the end of the previous section, the evidence for this interpretation is not entirely conclusive either.
Fig. 4.7. Trace normalized, vertical component record sections with 10-24 Hz band-pass filter (a) and 5 Hz low-pass filter (b). Continuous travel-time curves correspond to the wide-angle reflections from the crust-mantle transition of the model in Fig. 4.13, calculated with a ray-trace program (Obre condu 1976) taking into account the curved sediment-basement boundary shown in Fig. 4.4c. The dashed curve in (a) indicates the PMP-precursor (PrP).
4.6 Lower crust

While the Pg-phase is quite impulsive, the PMP has a fairly well correlated precursor (denoted here by PrP) and is followed by a coda of irregular reverberations.

As discussed in detail in Section 2.4, the reverberations forming the coda of the PMP can be explained either as multiple reflections and P-S conversions of the PMP within the sediments or as a consequence of some kind of upper mantle lamination. Since the length of the profile is insufficient to detect a first arrival refracted from the upper mantle (Pn), the data cannot contribute anything to the knowledge of the structure below the Moho. Thus, following the reasoning of Edel et al. (1975), a constant upper mantle velocity of 8.0 km/s was adopted for further calculations.

The precursor to the PMP appears to be frequency dependent: it is enhanced in the high-pass record section, while it is only faintly visible in the low-pass section (see Fig.4.7). In fact, spectral analysis shows that the main signal energy lies in the frequency range between 2 and 16 Hz. However, the spectrum of the Pg-phase peaks around 8 Hz, while the wide-angle reflection from the lower crust and crust-mantle boundary often contains a maximum around 4 Hz and another one around 10 Hz. Since two different shots were recorded in the distance range in which the PMP and its precursor are observed, it is very unlikely that this signal character is merely a source effect. Thus the structure of the lower crust appears to be selective with regard to the frequency of the seismic signals returned from it. On the basis of the discussion in Section 2.4, we must conclude that the crust-mantle boundary beneath Sulz is a transition zone several kilometers thick extending over most of the lower crust, and that it does not consist of a smooth gradient but of a series of velocity jumps or even of a series of lamina-like inversions.

In order to fit the Sulz PMP-travel-time data as well as the amplitudes, the linear velocity increase from one step or lamella to the next in models 10 or 11 had to be modified (see Fig.2.9). The data require a thicker transition zone with velocities increasing slowly at the top and more rapidly below. This resulted in the models 10s and 11s, presented in Figures 4.8a and b, together with the corresponding synthetic seismograms. For comparison, the recorded seismograms in the same distance range are reproduced in Figure 4.8c.

A qualitative comparison alone already indicates that the recorded signal is better matched by the lamella structure. Only the most significant part of the record section is shown: beyond about 85 km, the precursors interfere with the emerging Pn phase, while at distances shorter than 68 km, they are masked by strong signal-generated noise (see Figs.4.3 and 4.7). Some records at the shorter distances lack a clearly defined PMP-arrival, so that for a quantitative evaluation not all amplitudes could be measured. While both the PMP-amplitude-distance curves and the Pg to PMP amplitude ratios of the step- and of the lamella-model fit the data equally well, the PrP to PMP amplitude ratios allow one to distinguish between the
Fig. 4.8 a and b) Synthetic seismograms (vertical component, ground velocity, dominant frequency 8 Hz) for the crust-mantle transitions shown in the insets. Upper crustal model used corresponds to model 9 in Fig. 4.5b. Note the phases between Pg and PMP-precursors, caused by multiples and conversions of Pg within and beneath the sediments.

c) Vertical component records, 4-16 Hz band-pass filtered. Vertical lines indicate amplitude of precursors as plotted in Fig. 4.9. Amplitudes of synthetics and data are scaled by multiplying with distance.
Fig. 4.9 Maximum amplitudes of the PMP main phase (top) with thin curve corresponding to the smoothed data and thicker ones to the step- (10S) and lamella-model (11S) in Fig. 4.8. Amplitude ratio of Pg to PMP (bottom left) and of PMP-precursor (PrP) to PMP main phase (bottom right), with the corresponding model curves. Crosses correspond to data (from 4-16 Hz band-pass filtered records) with shot numbers above and station numbers below.
two models (Fig. 4.9). In the distance range over which the precursors are clearly identifiable, the lamella-model fits the data very well, while the step-model deviates by about a factor of 3. Since errors due to faulty instrument gains or to differences in local site responses have no effect on the amplitude ratios this deviation can be regarded as significant. From this it follows that, of the models discussed here, a laminated lower crust is the most likely explanation for the observations.

In order to account for high-frequency precursors before the lower-frequency PMP-phase observed along several seismic refraction profiles, Fuchs and Schulz (1976) proposed a model of the crust-mantle transition consisting of a thin high-velocity lamella over a zone of strong velocity gradient. However, in this case, the large velocity discontinuity at the top of the transition zone formed by the lamella would produce precursor amplitudes larger than those observed under Sulz.

Two of the models consisting of a series of step-like gradients proposed by Edel et al. (1975) for the southern Black Forest were also investigated: while the travel times can be made to fit the Sulz data quite accurately, the amplitudes of the PMP and its precursor do not match the observations sufficiently well. Indeed, it is likely that a quantitative amplitude interpretation of Edel's data might reveal that the PMP-precursors visible in some of his record sections are more adequately explained by some kind of lamination similar to that proposed here.
4.7 The travel-time delays between Rhine and Laegeren

As mentioned before, both Pg- and PMP-arrivals in the distance range between 91 and 101 km show a considerable delay with respect to the general trend of the travel-time curves. This delay, which reaches 0.2 s at stations 37 and 38, can be seen particularly well by correlating individual peaks and troughs of the PMP-phase in the low-pass filtered record section (Fig.4.7b). Despite the poor signal-to-noise ratio, this same delay can also be found among the Pg-arrivals on several of the appropriately filtered and amplified records in Figure 4.4a. Since both Pg and PMP are similarly affected, the cause of the anomaly must be located close to the surface between stations 35 and 40. This section of the profile coincides exactly with the Molasse Basin between the Rhine at 90 km and the Laegeren Mountain at 102 km (Fig.4.3b).

The most obvious explanation would attribute the delay to the low velocity of the Molasse sediments embedded in faster Mesozoic limestones. Own velocity measurements at shallow depth just north of the Rhine, gave P-wave velocities of 2.4 km/s for the Molasse and 4.7 km/s for the Malm limestones beneath. Rybach (1962) found average values between 2.6 and 3.0 km/s for Molasse sediments and between 4.1 and 4.9 km/s for various Malm formations just northwest of the Laegeren Mountain. From several borehole measurements in northeastern Switzerland, Lohr (1967) reports values of 2.62 km/s at the surface, increasing to 3.3 km/s at 1 km depth for the Tertiary, and of 4.6 km/s for the Mesozoic sediments at depths between 1 and 2 km.

Even assuming an unreasonably high contrast of 2.0 to 5.0 km/s between Tertiary and Mesozoic sediments, a structure as sketched in the geological cross-section (Fig.4.3b) will not fully account for the observed delays. In fact, ray-trace calculations showed that at station 37 only half of the 0.2 seconds delay can be explained by such a model. Moreover, a relatively small velocity contrast between Tertiary and Mesozoic sediments seems much more likely in view of the fact that no delay is observed in the Molasse Basin south of the Laegeren beyond 104 km.

Thus one is forced to conclude that the sediment-basement boundary is not as smooth as shown in Figure 4.3b, but that there must be some kind of depression reaching all the way into the crystalline basement. This hypothesis has in the meantime been fully corroborated by the results of a set of short refraction lines recorded in this area (Sierro et al. 1983) and by VIBROSEIS reflection profiles as well as by a borehole drilled all the way into the basement at a site close to station 37 (NAGRA 1983). The crystalline basement, consisting of biotite-gneiss, was reached only at a depth of 2020 m, instead of the expected 950 m, where it forms the bottom of a more than 1000 m deep trough filled with Permo-Carboniferous sediments.
4.8 Crustal Poisson's Ratios

Refraction seismology has contributed significantly to the identification of possible rock types constituting the earth's lithosphere through the determination of their elastic parameters. However, most interpretations have been limited to the study of P-wave velocities, neglecting the additional information contained in the S-wave data. This is due to the greater difficulty associated with recording and analyzing S-waves. In recent years, with the advent of three-component recordings and the availability of digitized record sections, more sophisticated processing techniques have been applied to the data, resulting in several successful shear-wave studies of the earth's crust (e.g. Braile et al. 1974, Keller et al. 1975, Zschau and Koschyk 1976, Assumpção and Bamford 1978, Haggag 1980, Banda et al. 1981, Gajewski 1981, Stroessenreuther 1982).

Since quarry blasts are in general a good source of shear waves, the Sulz data allow one to determine Poisson's ratio (sigma) for the crust in this area. To give an overall impression of the quality of the S-wave data, the record section of the transverse component, band-pass filtered between 4 and 16 Hz, is plotted with a reduction velocity of 3.5 km/s in Figure 4.10. The radial component records are of similar quality. Though the signal-to-noise ratio is poorer than in the P-wave record sections, onsets of Sg and SMS can be identified with good reliability over distance intervals of sufficient length. These intervals are plotted at a larger scale in Figure 4.11.

A study of shear waves along several profiles in Great Britain (Assumpção and Bamford 1978) showed that particle motion plots alone do not significantly increase the accuracy of S-wave onset determinations. Polarization filters, as used by Haggag (1980) and Gajewski (1981) can improve the detectability of SV-arrivals, but require that the angle of incidence at the surface be less than \( \sin^{-1}(vs/vp) \), where \( vs/vp \) is the ratio of S- to P-velocities in the surface layer. In the case of Sulz, due to the relatively hard sediments at the surface, this condition is not fulfilled. Therefore, the S-wave onsets were determined simply by the traditional criteria of amplitude and frequency change. The uncertainty of the picked PMP-, SMS- and Sg-arrivals was estimated to be \( \pm 0.1 \) s, which is indicated by the circles in Figure 4.11. The average of the differences between the radial and transverse component readings for Sg is equal to 0.002 s, which is practically 0. Since a significant amount of SV-SH velocity anisotropy should manifest itself as a systematic time shift between S-wave arrivals on radial and transverse components, this effect was thus assumed to be negligible. Therefore, whenever possible, arrival times used to calculate Poisson's ratio were determined from the average of the readings of the two horizontal components. The average of the absolute values of the differences between these two readings amounts to 0.09 s, which indicates the error estimate of \( \pm 0.1 \) s to be reasonable.

Since \( Tp = D/Vp \) and \( Ts = D/Vs \), where \( D \), the distance along the ray path between shot and receiver is assumed to be the same for both P- and S-waves, the travel-time ratios \( Ts/Tp \) are equivalent to the velocity ratios \( Vp/Vs \).
Fig. 4.11 Transverse (T), radial (R) and vertical (V) records (4-16 Hz band-pass filtered, trace-normalized) showing Pg, Sg, PMP and SMS arrivals. The dashed line indicates the PMP precursors reflected from the laminated crust-mantle transition.
These ratios are plotted for each station as a function of distance in Figure 4.12 together with the average values computed separately for Sg/Pg and SMS/PMP. From these average values, Poisson's ratio can be calculated directly using the standard formula:

\[ \sigma = \frac{1}{2} \left( \frac{V_p^2}{V_s^2} - 1 \right) \]

This method bypasses the need to construct a separate S-wave model (Assumpcao and Bamford 1978). The result is an average sigma-value of 0.238 ±0.002 for the sediments and the upper crust, corresponding to a depth range of 4 to 5 km. Since sigma for sediments is generally higher (Assumpcao and Bamford 1978, obtained values between 0.27 and 0.33), the decreasing trend of the first three values of TS/TP in Figure 4.12 may be due to a sedimentary influence, so that the actual sigma value for the upper crystalline basement is probably somewhat lower than 0.238. Based on the SMS/PMP data, the resulting average value for the whole crust is 0.244 ±0.002.

Fig. 4.12 Ts/Tp arrival-time ratios— which are equivalent to Vp/Vs velocity ratios — and the corresponding values of Poisson's ratio.
As indicated by the dashed line in Figure 4.11 and discussed in Section 4.6, the P-wave reflections from the crust-mantle transition consist of a high-frequency precursor, corresponding to the subcritical reflections from the individual lamellae, followed by a lower frequency wide-angle reflection with larger amplitudes. While the arrivals picked for the PMP clearly correspond to the wide-angle reflections, it should be noted that the S-wave reflections from the crust-mantle transition do not exhibit the same distinctive pattern of precursor and main phase as the P-wave reflections. This could be due to the fact that the S-waves, which have slightly lower dominant frequencies (see Fig.4.11), are less affected by the lamella structure than the P-waves. On the other hand, the separation between precursor and main phase may just be less distinct in the case of the S-waves, and the arrivals picked in Figure 4.11 may actually correspond to the precursors. In the latter case, the average sigma value for the whole crust given above would be too low. It is not possible to give an accurate value of sigma in the lower crust, but one can conclude from these data that it is higher than the average value given for the whole crust and that Poisson's ratio in the crust beneath Sulz increases with depth. In other words, there is evidence for a slight decrease with depth of the velocity of S-waves relative to that of P-waves.
4.9 Summary of results

The complete velocity-depth model proposed for the crustal structure along the eastern margin of the southern Black Forest is shown in Figure 4.13 and listed in Appendix D. Its main features are a sedimentary layer with an average thickness of about 1 km, a strong gradient in the upper crust and a lower crust composed of a series of lamella-shaped velocity inversions.

The sedimentary or surficial weathered layer has a significant and often neglected influence on the character of the record section and on the model as a whole: small variations in sedimentary structure or velocity can produce significant travel-time effects, while multiple reverberations and P-S conversions between the top of the basement and the earth’s surface can mask phases from deeper discontinuities. For example, without independent information from borehole data about the dip of the sediment-basement boundary under Sulz, the travel-time interpretation would have resulted in a significantly different velocity gradient in the upper crust, and it would have been difficult to achieve an agreement with the amplitude data. A more accurate knowledge of the sedimentary structure at the southern end of the Sulz-profile would put stronger constraints on the lower part of the upper crustal gradient as well as on the shape of the velocity increase in the transition zone above the Moho.

![Graph showing velocity-depth model of P-waves for SULZ 11S.](image)
The results of the interpretation of the Sulz-south profile can be summarized as follows:

1) The upper crustal gradient under the Sulz-south profile is exceptionally well constrained by information about the basement topography (boreholes) and by travel-time as well as amplitude data.

2) Unambiguous evidence for or against a discontinuity or low-velocity zone in the middle crust could not be found from these data alone. For this reason, the middle crust in Figure 4.13 has been drawn with a dashed line. The possible alternatives are discussed in light of other data in Section 6.2.

3) The evidence for the lamination in the lower crust is based on a quantitative comparison of the amplitudes of the PMP and its precursors with those of theoretical seismograms. It also correctly accounts for the observed travel-times and for the frequency dependent nature of the reflections from the crust-mantle transition. As discussed in more detail in Section 6.2, both the depth of the Moho and its characteristic structure agree well with results obtained in adjacent areas.
5. SOUTHERN ALPS

5.1 Introduction

The crustal structure of the Southern Alps, situated south of the
Periadriatic Lineament and extending roughly from the Ivrea Zone in the west
to the Dinarides in the east, has been investigated by means of several
seismic refraction profiles over the past 22 years (Fig. 5.1). The early
work, including several lines from shotpoint Lago Lagorai (Behnke et al.
1962, Giese 1968), was supplemented in 1963 and 1964 by two lines to the
southeast and east from shotpoint Lago Bianco (Giese 1968, Behnke 1969,
conjunction with the large-scale seismic refraction project along the axis
of the Alps in 1975 (ALP 75), a reversed north-south profile was recorded
between Innsbruck and the Po Plain (d-h in Fig. 5.1) (Italian Explosion
Seismology Group 1978). In 1977, an international effort (SUDALP 77) was
undertaken to obtain information about the crustal structure beneath the
Southern Alps by recording several shots along a west-east trending line
between Lago Maggiore and the Yugoslav border (A,B,C,D in Fig. 5.1) and two
shorter profiles between Lago Maggiore and the Insubric Line (05 and 07 in
Fig. 5.1). A first interpretation of the section from shotpoint A to B was
presented by Ansorge et al. (1979) and Mueller et al. (1980).

A uniform interpretation for the whole line from A to D, together with
a short NW-SE line in the eastern part, recorded in 1978, was presented by
the Italian Explosion Seismology Group (1981). The model proposed by these
authors, based mainly on travel-time interpretation of the reflected phases,
includes a layer of high velocity (7.0-7.2 km/s) in the middle crust,
embedded between two prominent low-velocity zones. This has been
interpreted as evidence for a wedge-like form of crustal doubling beneath
the Southern Alps. Giese et al. (1982) regard this model as typical for
the mechanism of continental collision during Alpine orogeny in the whole
Mediterranean region (see Fig. 5.2).

The present work reevaluates the western part of the SUDALP 77 data,
including the profiles A-05 and A-07, on the basis of not previously
available, amplitude controlled, digitized record sections. Since the
postulated high-velocity layer in the middle crust functions as a strong
reflector, modelling the measured amplitudes with synthetic seismograms
provides an independent means for its verification and thus a test for the
hypothesis of crustal doubling in this region.
Fig. 5.1 Tectonic sketch map (after Frey et al. 1974) with major seismic refraction profiles.
a-i: ALP75; A-D: SUDALP77; LBV: Lago Bianco - Verona; LL: Shotpoint Lago Lagorai; T: Shotpoint Obersee.
Fig. 5.2 Model of crustal doubling beneath the Southern Alps (from Giese et al. 1982).
Fig. 5.3 Tectonic sketch map of the Southern Alps (after Staub 1949) with shot and station locations of the SUDALP77 profiles.
5.2 Geological setting

As shown by the station locations on the geological sketch map in Figure 5.3 and below on the elevation profiles in Figures 5.5 and 5.6, the seismic refraction lines discussed in this section cross regions of considerable tectonic complexity. Shotpoint A, at the western end, is situated in the Lombardy trough, consisting of basins with Mesozoic sediments several kilometers thick, separated by ridges, Permian intrusions and faults.

In the central part of the main profile, the crystalline basement, composed mainly of metamorphic schists and gneisses, is covered by a relatively thin layer of Triassic and Permian sediments. In the east, the profile crosses the Adamello intrusion, the Giudicaria fault and the quartz porphyry of Bolzano, where shotpoint B is located.

Profiles A-05 and A-07 terminate in the south-Alpine schists and gneisses, bounded by the Insubric Line in the north (see e.g. Staub 1949, Angenheister et al. 1972, Winterer and Bosellini 1981).

5.3 Acquisition and presentation of the data

At shotpoint A, at the southern tip of Lago Maggiore, four borehole shots were fired with charges between 400 and 800 kg. Shotpoint B actually consists of two distinct quarries about 2 km apart, from which one blast each with charge sizes of 4.3 and 4.6 tons was recorded (see Ansorge et al. 1979 for details). Recording instruments were all of the MARS type, and the data were digitized and processed as described in Chapter 3. The resulting trace-normalized record sections are shown in Figures 5.4, 5.5 and 5.6, bandpass filtered between 2 and 16 Hz with a recursive, zero-phase digital filter, to remove some of the noise. Since several records of shots B1 and B2 overlap, they were plotted separately (Fig.5.4).

Although the charge sizes of shots B1 and B2 were up to ten times larger than those at shotpoint A, their efficiency was considerably smaller; thus, beyond about 170 km, the signal to noise ratio became so poor as to make the records unusable. Calibration recordings at the same site for each shot are not available to correct for different charge sizes and shot efficiencies. However, since all shots from a particular shotpoint were recorded over the same distance range, amplitudes could be adjusted, by correcting for systematic deviations between them. Only shot A1, recorded along the main profile, deviated systematically by more than the internal scatter of the data, so that the corresponding amplitude values had to be reduced by a factor of 2.
Fig. 5.5 Trace-normalized record section, filtered 2-16 Hz, from shotpoint A to the east, along profile 01 and part of 02. Station elevations are plotted with fivefold vertical exaggeration. Geology: Q, T: Quaternary and Tertiary sediments of the Lombardy trough, L: Liassic sediments; Tr, P: Triassic and Permian sediments of the Bergamasca Alps and Dolomites; Kr: South Alpine crystalline basement (schists and gneisses); Ad: Alpine granitic intrusion of the Adamello complex; B, Qp: quartzporphyry of Bolzano; C.A.: Cima d'Asta crystalline.
Fig. 5.6 Trace-normalized record sections, filtered 2-16 Hz, from shotpoint A along profiles 05 and 07. Station elevations are plotted with fivefold vertical exaggeration. For explanation of geology see Fig. 5.5.
5.4 Upper crust

In order to enhance the first arrivals, which for the most part emerge only gradually from the background noise, Figure 5.7 shows only the first few seconds of the seismograms along the three profiles originating from shotpoint A. The arrival times of those signals which could be determined with sufficient confidence are plotted in a single composite diagram (Fig.5.7). The large travel-time discrepancies (almost 1 s between profile A-01 and A-05 at 30 km) and different apparent velocities clearly illustrate the extreme lateral heterogeneity around shotpoint A, which is already evidenced by the complex surface geology. Since the experiment was devised to provide information mainly about the structure at greater depth, station spacing is insufficient to resolve near-surface details of such complexity. Thus only a qualitative interpretation shall be attempted here.

Common to all three profiles, originating from shotpoint A, is an uppermost sedimentary velocity of about 4.2 km/s. Going north from the main profile, the apparent velocity of the next layer increases from 5.0 to 5.8 km/s and the first arrivals occur earlier. At larger distances, profiles A-01 and A-07 together seem to define a common Pg-velocity for the crystalline basement of about 6.05 km/s. As will be shown later, arrivals from the middle crust and from the crust-mantle boundary match also quite closely along the two profiles. Thus, the strong lateral heterogeneity seems to be limited to the first 50 km around the shotpoint. The delay of up to 0.4 s between the arrivals on profile A-01 and A-07 at about 30 km is due to the influence of the Lombardy trough deepening very rapidly as one goes south from the Lugano ridge.

Profile A-05 exhibits a somewhat different character: in addition to the much earlier arrivals and to the higher apparent velocities (6.2 km/s for the Pg), the signal form of the first arrivals is much more impulsive and the amplitudes are larger. Moreover, the subsequent reflections are more distinct than on the other two record sections. This points to some kind of fundamental structural change as one goes north from profile A-07 towards profile A-05.

As mentioned earlier, shotpoint B is located in the quartz porphyry plate of Bolzano. As shown by the composite arrival plot of profiles B-west and B-east (Fig.5.8), it is characterized by a velocity of 4.6 km/s. Despite the geological complexities, the arrival times over the first 40 km are practically the same in both directions and define a velocity of about 6.0 km/s. Beyond about 40 km, the apparent velocity seems to increase, but the arrivals exhibit a much larger scatter. Since the digitized section of B to the east ends already at 52 km and the manually plotted records used by the Italian Explosion Seismology Group (1981) do not have sufficient resolution, it is not possible to say whether the relatively early arrival of the last three digitized signals of this profile is due to a local surficial effect, or to a structural change at depth.
Fig. 5.7 Record sections of first arrivals along profiles from shotpoint A.
Fig. 5.8 Record sections of first arrivals along profiles from shotpoints B1 and B2.
Fig. 5.9 True-amplitude record section, filtered 2-8 Hz, of profile A-01 with picked phase correlations. P1P, P2P and P3P denote crustal reflections in the order of their arrival; PMP denotes the reflection from the Moho. Pg is determined from Fig. 5.7 and P2P is better visible in Fig. 5.10. Amplitudes are multiplied by distance.
Fig. 5.10 True-amplitude record section, filtered 2-8 Hz, of profile A-07 with picked phase correlations. Amplitudes are multiplied by distance.
Fig. 5.11 True-amplitude record section, filtered 2-8 Hz, of shots B1 and B2 along profile 01 with picked phase correlations. Amplitudes are multiplied by distance. Overlapping seismograms are omitted (see Fig. 5.4).
5.5 Middle crust and crust-mantle boundary

The record section from shotpoint A along the main profile (A-01) is characterized by very prominent PMP-reflections (Fig.5.9). They also appear with almost the same arrival times on the last records of profile A-07 (Fig.5.10). Three intermediate phases are also present, but are much less coherent. The first intermediate phase (P1P) follows the first arrivals quite closely and is best visible on the record section of shotpoints B to the west (Fig.5.11). It is interesting to note that the second phase (P2P in Figs.5.9 and 5.10) is quite prominent on profile A-07, while it is hardly visible at the same distance on profile A-01. However, the P2P-phase in record section A-07 is matched quite well by the intermediate arrivals along A-01 in the distance range between 89 and 126 km. Thus, if one associates these arrivals with the same reflector beneath the two profiles, the different amplitude behaviour in the two record sections could be due to scattering effects from a non-planar surface of the reflector. In the record section from shotpoint B to the west, as well as along A-01, there is evidence for an additional intermediate reflection, designated as P3P in Figures 5.9 and 5.11.

Fig.5.12 Velocity-depth models from profiles A-01 (A) and B-01 (B) as derived by the Italian Explosion Seismology Group (1981). Dotted model in (A) is from Ansorge et al. (1979).
For the crust-mantle transition beneath profile A-01, Ansorge et al. (1979) proposed a model consisting of a step from 6.5 to 7.1 km/s just above the Moho. The Italian Explosion Seismology Group (1981) and Cassinis (1982), on the other hand, favour a transition in the form of a smooth gradient (Fig. 5.12). As shown in Section 2.4, the amplitude of the PMP-phase is sensitive to the abruptness of the velocity change just above the mantle: the slope of the amplitude-distance curve at distances below the amplitude maximum, allows one to distinguish between a model with a smooth transition all the way to the mantle and one with a sharp discontinuity at the bottom of the lower crust (compare models 4 and 9 in Fig. 2.8). Figure 5.13 shows the PMP-amplitude-distance data for the two profiles A-01 and B-west together with the theoretical curves calculated for a model with a sharp discontinuity at the base of the crust. Except for three values of the A-01 data, which deviate by more than a factor of three from their neighbouring points, theoretical curves and data show good agreement. For both profiles, the slope of the amplitude-distance curves of the synthetic seismograms at distances below that corresponding to the peak of the curves matches the data very well. This is good evidence for a sharp discontinuity between the lower crust and upper mantle, rather than a smooth transition.

Fig. 5.13 PMP amplitude-distance data for profiles A-01 and B-01 together with calculated curves for the Moho models in the insets. Data points are denoted by crosses with shot (above) and station numbers (below).
Since profile B-west is too short and the signal-to-noise ratio at large distances along profile A-east is too poor, no reliable velocity could be determined from these data for the wave refracted in the uppermost mantle (Pn). Rough values could be extracted from three of the older profiles of which Behnke (1969) compiled the analog record sections: Lago Bianco to Lago Lagorai to Tarvisio = 8.05 km/s, Lago Lagorai to Leibnitz = 8.28 km/s, Lac Negre to Brescia to Lago Lagorai = 8.12 km/s. None of these profiles were reversed, so that these values correspond only to apparent velocities. An intermediate value of 8.1 km/s was chosen for the calculations in this study.

According to Ansorge et al. (1979), the relatively high velocity above the Moho is required to explain the high apparent velocity of the PMP-phase along profile A-01, between about 140 and 170 km (see Figs. 5.5 and 5.9). Whether the step is as sharp as assumed here, or whether it is a smoother transition or even a series of small steps cannot be answered by the amplitude-distance data (see Section 2.4). However, the absence of distinct precursors to the PMP-phase between 90 and 120 km in the case of profile A-01 and between 110 and 140 km for profile B-west, as observed along the Sulz-south profile, indicates that the transition from 6.3 or 6.4 km/s to 7.1 or 7.2 km/s may be less abrupt than assumed here and that the lower crust does not contain any significant velocity reversals.

Because of the difficulty in accurately determining the arrivals of the intermediate phases mentioned above, various models for the middle crust can be found which satisfy the travel-time data alone. Thus, after determining several possible models compatible with the travel-times, amplitude calculations were performed. Due to the extreme lateral heterogeneity around shotpoint A, amplitude interpretations were attempted only for profile B-west. The maximum amplitudes for each phase as a function of distance are plotted in Figure 5.14. Except for the well defined PMP-phase, the values exhibit too much scatter as to allow any conclusions to be drawn from the shape of the amplitude-distance curves. Thus, only the general amplitude level of each phase, as denoted by the thick curves in Figure 5.14, were taken into consideration. These smooth curves are reproduced in a composite diagram in Figure 5.15, showing the amplitude level of the three phases relative to each other. Figure 5.15 also shows the theoretical amplitude-distance curves, calculated by the reflectivity method for four different models compatible with the travel-time data.

Though the agreement is not perfect, model C fits the observations best. In this case, PIP with the lowest amplitude level and a rapid decay with distance corresponds to the reflection from the top of a velocity inversion situated at a depth of 7 km, P2P is a reflection off the bottom of this low-velocity zone at 12 km, and P3P comes from a velocity increase from 6.2 to 6.4 km/s at a depth of 18 km. In the case of model A, with the low-velocity zone situated at greater depth, and of model B with three positive velocity jumps and no inversion, the relative amplitude levels of the first and second phases are reversed with respect to that observed in the data. Model D is an approximation to that proposed by the Italian Explosion Seismology Group (1981) and discussed by Giese et al. (1982) with
Fig. 5.14 Amplitude-distance data and smoothed curves for the four reflections correlated along profile B-01 in Fig. 5.11. Numbers next to the crosses designate shot (above) and station (below).
Fig. 5.15 Amplitude-distance curves calculated with the reflectivity method for models A-D. Numbers refer to the reflections from the corresponding discontinuities (4 is equivalent to PMP). The data plot is a composite of the smoothed curves in Fig. 5.14.
Fig. 5.16 Ray-trace model from shotpoint A to the east along profile 01. Crosses correspond to calculated travel-times; continuous lines correspond to the correlations in Fig. 5.9.
Fig. 5.17 Ray-trace model from shotpoint B to the west along profile 01. Crosses correspond to calculated travel-times; continuous lines correspond to the correlations in Fig. 5.11.
the high-velocity "tooth" in the middle crust (see Figs. 5.2 and 5.12 (B)). In this case, the amplitudes of P-P, corresponding to the reflection from this high-velocity layer, are much too strong, surpassing even those of the PMP by more than a factor of two. Thus the existence of the proposed mid-crustal high-velocity layer beneath profile B-west must be regarded as very unlikely. Whether it exists to the east of shotpoint B cannot be checked at this stage, since the corresponding data are not available in digital, amplitude corrected form.

The crustal structure of model C, derived here for the profile from shotpoint B towards A on the assumption of a one-dimensional structure, differs only slightly from that derived for the reversed profile by Ansorge et al. (1979): The low-velocity zone, situated at approximately the same depth, has a velocity of 5.7 instead of 5.4 km/s, the middle crust consists of two layers of 6.2 and 6.4 km/s, instead of a single layer of 6.3 km/s, and the transition layer at the base of the crust has a velocity of 7.1 km/s instead of 7.2 km/s. The main difference consists in the significantly greater depth of the Moho beneath the eastern part of the profile (45 km) than beneath the western part (35 km).

Two-dimensional ray-tracing was performed for the two shotpoints in an attempt to extend this one-dimensional model over the laterally heterogeneous structure present here. The results are shown in Figures 5.16 and 5.17. The computer program used, developed by Gebran (1976), calculates only diving rays and wide-angle reflections. Thus the subcritical branches of the travel-time curves and the reflection from the top of the low-velocity layer (P1P) are not matched by calculated arrivals. However, the one-dimensional travel-time and amplitude calculations for profile B-west show a good agreement between model and data for these phases as well.

5.6 Near-vertical reflections from shotpoint A

As can be seen from the reflection points of the rays in Figure 5.17, the depth of the crust-mantle boundary beyond about 160 km and in particular directly beneath shotpoint B, is given merely by extrapolation. Below shotpoint A, on the other hand, the model agrees with some near-vertical reflection profiles recorded concurrently with the refraction data.

The corresponding seismograms, as published by the Italian Explosion Seismology Group (1981), are shown in Figure 5.18. The travel-time curves drawn between the record sections were not derived from these data, but were calculated from the dipping-layer model on the left side of the figure, which corresponds to the first 20 km east of shotpoint A as shown in the ray-trace diagrams in Figures 5.16 and 5.17. Nevertheless, some clear correlations can be seen: in addition to the prominent upper crustal reverberations at the top, there are some conspicuous bands of energy around 7 seconds and between 10.5 and 11.5 seconds. The former is interpreted in
terms of the base of the low-velocity zone, and the latter corresponds to
the crust-mantle transition. Note that only the depth of the latter was
adjusted on the basis of these reflection data, while the other correlations
merely correspond to the interfaces extrapolated from the ray-trace models.

Fig. 5.18 Near-vertical reflection sections in the vicinity of shotpoint A
(from Italian Explosion Seismology Group 1981) with phase correlations for
the velocity-depth model shown. The depth of the crust-mantle transition
was adjusted to match the reflected energy visible around 11 s, all other
discontinuities and the corresponding correlations were extrapolated from
the ray-trace model in Figs. 5.16 and 5.17.
Fig. 5.19 Record section of profile Lago Bianco - Verona from Behnke (1969). Phase correlations correspond to the ray-trace model below.
5.7 Comparison with profile Lago Bianco-Verona

As shown in Figure 5.1, at a distance of about 105 km east of shotpoint A, the main SUDALP77 profile intersects an older line recorded from Lago Bianco, in the central Alps, and ending southeast of Verona. Part of the data was first evaluated by Giese (1968) and then, after it had been compiled in complete form by Behnke (1969), reevaluated by Egloff (1979). The record section, available only in analog form, is shown in Figure 5.19. The intersection with the SUDALP profile is at about 120 km and the Insubric Line is crossed at about 75 km from the shotpoint.

Based on the results from Ansorge et al. (1979), Egloff (1979) fixed the thickness of the crust beneath the intersection with profile A-01 at 35 km, assumed the upper mantle velocity to be 8.2 km/s, and adjusted the dip of the Moho in order to match the high apparent Pn-velocity of about 9.2 km/s. This resulted in a continuously rising Moho from a depth of 50 km below Lago Bianco to 20 km below Verona.

On the basis of the present evaluation of the Southern Alps profile from shotpoint B to A, crustal thickness increases from west to east, resulting in a Moho depth of 43 km below the intersection of the two profiles. Though the amplitude of the Pn-phase is weak and its onset often difficult to pick with precision, the arrivals between -4 and -6 seconds beyond 250 km are clearly visible. In order to explain these early arrivals, a very pronounced step-wise thinning of the crust must occur to the southeast of Lago di Garda. The ray-trace diagram illustrating the resulting structure is shown in Figure 5.19, together with the calculated travel-time curves.

No attempt was made to derive a detailed model of the whole crust along this profile, which - due to the intersection of the Insubric Line - is probably more complex than that shown. However, in the region of concern, it corresponds exactly to the structure derived for the Southern Alps, and the calculated travel-times show reasonable agreement with the data. Neither this nor Egloff’s interpretation explain the prominent secondary arrivals at about 1.5 seconds, between 220 and 260 km: a correlation with the PMP-phase, as seen between 120 and 200 km, is irreconcilable both with the depth of the Moho as derived from the SUDALP data, and with the early arrivals of the Pn. Whether they are due to some greater complexity in the lower crust in connection with the Insubric Line, or whether they are due to some kind of constructive interference, caused by multiple reflections within the crust, cannot be answered on the basis of the presently available data.
5.8 Summary of results

The reevaluation of the western part of the SUDALP77 refraction profiles between shotpoints A and B, and of the Lago Bianco - Verona profile can be summarized under six points:

1. The well-known Lombardy trough consists of sediments with velocities between 4.2 and 5.0 km/s, extending to a depth of 7.5 km below shotpoint A. To the east, in direction of the main profile, it thins out at about 50 km from the shotpoint, while to the north, towards profiles 07 and 05, it is bordered by very steeply rising flanks. Judging from the crystalline rocks outcropping only 10 to 20 km west of shotpoint A, the trough must be bounded quite sharply in that direction as well, while towards the south, below the Po Plain, its extent and thickness is difficult to estimate.

2. The existence of an upper-crustal low-velocity zone is in agreement with travel-time as well as with amplitude data. The highly variable amplitude character of the corresponding reflections along the two closely neighbouring profiles 01 and 07 is an indication of strong lateral variations within such a zone, though the good match of the travel-times indicates no systematic change between the two lines. If one thinks of such upper-crustal velocity inversions to be due to a series of mushroom- or pillow-like granitic intrusions, as postulated by Mueller (1977), one would indeed expect to see a somewhat erratic character in the seismic signal.

3. An additional intermediate reflection is evidence for a discontinuity in the middle crust. The amplitudes of this phase can be satisfactorily modelled with a small velocity increase from 6.2 to 6.4 km/s. A high-velocity tooth, as postulated by the Italian Explosion Seismology Group (1981), is contradicted by the amplitude calculations.

4. The lower crust represents a transition to the mantle. It is modelled here as a single step from 6.4 to 7.1 km/s with a thickness ranging from 2 km in the west to 7 km in the east. The data do not allow the determination of the detailed form of this transition zone: it could just as well consist of a smooth gradient or a series of small steps, but a lamella-like structure, as found beneath the Sulz profile seems unlikely. However, the amplitude character of the PMP-phase is evidence for a sharp velocity jump of about 1 km/s at the Moho itself.

5. Along the main profile, crustal thickness varies from 31 km in the west, below shotpoint A, to 46 km below the Adamello-Giudicaria region. The 0.3 s delay of the PMP-reflection at a distance of 90 km along profile A-07 relative to the corresponding phase along A-01 indicates a rapid deepening of the Moho towards the north.

6. To the southeast, along the Lago Bianco - Verona profile, the early arrivals of the Pn-phase are explained by a stepwise thinning of the crust to a depth of about 22 km in the region near Verona.
5.9 Comparison with gravity measurements

Given such pronounced variations in Moho topography, it is instructive to examine these results in light of what is known about the gravity field in the same area. Figure 5.20 displays part of a Bouguer anomaly map compiled by Schwendener (unpublished), based on data from the Swiss Gravity Map (Klingelé and Olivier 1979) and from the Italian Gravity Map (Ballarin et al. 1972), as well as on detailed measurements in the Ivrea Zone (Kissling 1980) and along a traverse from the Bodensee to Bergamo (Meyer 1982, Schwendener 1983). For the most part, the profiles evaluated here intersect the iso-anomaly lines at rather small angles or run parallel to them over large distances. A rigorous treatment of the gravity effect would therefore require three-dimensional modelling, taking into account the influence of laterally varying structures, which, however, shall not be attempted here. For the qualitative comparison that is of interest in this discussion, Figure 5.21 shows the trends of the Bouguer anomaly along the two main profiles, together with the depth of the Moho as derived from the refraction seismic data. In both cases, the good correlation between the two curves is quite striking: the thin crust at the southeastern end of the Lago Bianco - Verona profile corresponds exactly to the well-known Verona gravity high, and the dip in the Moho from shotpoint A to B along profile O1 is well matched by the negative trend of the Bouguer anomaly.

As shown in Figure 5.20, the influence of the Po Plain sediments can be neglected along both profiles, while the positive anomaly of the Ivrea body extends out to a distance of 50 km along profile O1. According to the model by Kissling (1980), the Ivrea effect amounts to as much as 50 mgal at shotpoint A. The fact that, in spite of this, the Bouguer anomaly plotted in Figure 5.21 correlates so well with the Moho depth could indicate that the positive Ivrea effect is largely counterbalanced by the opposite effect of the Lombardy trough sediments.

The good correlation between Bouguer anomaly and Moho depth tempts one to extrapolate the Moho beneath profile O1 further to the east, beyond shotpoint B, on the basis of the gravity trend. The resulting rise of the crust-mantle boundary, shown by the dashed line in Figure 5.21, agrees at least qualitatively with the fact that the PMP-phase arrives about a second earlier on the profile from shotpoint B to the east than on the one to the west (see Italian Explosion Seismology Group 1981, Figs. 3 and 4). This is, of course, valid only under the assumption that the crustal structure above is essentially the same in the two directions.
Fig. 5.20 Bouguer gravity map of part of the Southern Alps (Schwendener, unpublished). The dashed line corresponds to the -5 mgal contour of the gravity effect due to the Po plain sediments (Schwendener 1983), and the dash-dotted line delineates the +10 mgal contour of the gravity contribution of the Ivrea body (Kissling 1980). A and B = shotpoints of SUDALP77 profiles; LB = shotpoint Lago Bianco; V = Verona.
Fig. 5.21 Comparison of Bouguer gravity anomaly and Moho depth along profiles A-01 and Lago Bianco - Verona. Stars indicate the positions of the shotpoints. The dashed part of the Moho curve (top) is merely an extrapolation based on the trend of the Bouguer anomaly.
5.10 Comparison with previous results

Since the record sections beyond 50 km east of shotpoint B are not available in digital amplitude controlled form, it is not possible to reexamine the postulated existence of a mid-crustal high-velocity layer below the eastern part of the Southern Alps on the basis of an amplitude evaluation. Nevertheless, except for the earlier PMP-arrivals and the anomalous Pg-onsets around 50 km discussed in Section 5.4 (see Fig.5.8), the analog record section from shotpoint B to the east (Italian Explosion Seismology Group 1981, Fig.4) does not differ significantly from the one to the west. Thus, following what was stated in Section 5.5 regarding the nonuniqueness of models based on travel-time correlations alone, the data as evaluated by the Italian Explosion Seismology Group (1981) cannot be regarded as sufficient evidence for the existence of the mid-crustal high-velocity layer to the east of shotpoint B either.

Additional information can be obtained from the north-south profile d-h, recorded during the ALP75 experiment (see Fig.5.1), which intersects profile 02 through the Southern Alps at almost right angles. The point of intersection of the two profiles is situated about 80 km north of shotpoint h and 25 km east of shotpoint B. An examination of the two reversed record sections along profile d-h, as reproduced by the Italian Explosion Seismology Group (1978), reveals a peculiar feature. In the record section recorded from shotpoint d to the south, the PMP-phase is well defined all the way out to 200 km, with an arrival of about 1.6 seconds in reduced time at shotpoint h. In the reversed record section, on the other hand, the PMP-phase becomes unclear beyond about 180 km, but an extrapolation to a distance corresponding to the position of shotpoint d leads to an arrival time of only one or two tenths of a second in reduced time. Since travel-times between shotpoints of a reversed profile must always be equal, this discrepancy of over one second means that either an error was made in aligning the signals in one of the record sections, or the crust-mantle boundary drops very steeply at about 90 km, going north from shotpoint h (just beneath shotpoint Lago Lagorai), causing a shadow zone in the PMP-reflection, when shot from south to north.

The Italian Explosion Seismology Group (1978) offers two alternative interpretations for this profile. The first extends the shallow crust-mantle boundary from below the southern end, as a high-velocity layer into the middle of the thicker crust in the north, which leads to the model of crustal doubling advocated by Giese et al. (1982) (see Fig.5.2). The second is characterized by a rapid thickening of the lower crust beneath the central part of the profile, in a manner similar to that proposed here for the SUDALP profile between shotpoints A and B. Assuming the record sections were correctly assembled, the shadow zone in the PMP reflection, discussed above, favours the second interpretation, and must be considered as further evidence against the existence of an intermediate high-velocity layer in the crust of the southern Alps.
6. CONCLUSIONS

6.1 Methodological considerations

The present study constitutes an instructive example of the use of amplitude information in seismic refraction experiments to constrain the number of possible interpretations based on travel-time data alone. The conclusions derived from the numerical experiments and from the application of synthetic seismogram calculations to actual data are discussed in detail in the previous chapters and shall not be reiterated here. However, the experience gained in this study suggests a few remarks concerning improvements in experimental and interpretational procedures to make seismic refraction results more conclusive.

Long refraction profiles are usually the result of a cooperative venture involving several research groups from different institutions. Even though instrument types may be basically the same, modifications and repairs often alter the original specifications, causing differences in recording characteristics and sensitivity. The consequent ambiguities which arise during interpretation of the data can be largely avoided by recording a small test shot before every major campaign with all instruments set up in one point. Such a test recording also reduces the risk of unexpected instrument malfunctions and allows the correction of seismometer polarity reversals, which, if left uncorrected, increase the difficulty of proper phase correlations significantly.

In order to achieve sufficiently high station density over a long profile, it is often necessary to record several shots from the same shotpoint at different recording sites. Even though charge sizes may be the same, shot efficiency can vary quite drastically. It is therefore very helpful for an absolute amplitude interpretation to have one instrument recording all shots at a single position, so that the corresponding amplitude differences can be corrected for.

Signal character and amplitudes are very sensitive to the local geological conditions at the shotpoint as well as at the recording locations. Both travel-time and amplitude interpretations become increasingly uncertain when, in order to increase station density, a profile is recorded from several different shotpoints a few kilometers apart. The case of the profiles from shotpoints B1 and B2 in the Southern Alps, where shotpoints and recording stations were moved simultaneously, illustrates the difficulties which arise from the resulting scatter in the data. Even moving only the shotpoint while keeping the recording instruments fixed, as is often done in experiments using marine shots, can significantly alter the signal character as a consequence of different water depth and sedimentary structure, thus making reliable phase correlations impossible.

Indeed, the importance of careful selection of the shot site is
illustrated very clearly by comparing the signal character from the Sulz quarry blasts with that from shotpoint A of the SUDALP profile. In the latter case, the complex sedimentary structure of the Lombardy trough has certainly contributed significantly to the diffuse appearance of the record section, characterized by the prominent reflections over the first 70 km, the strongly attenuated first arrivals at larger distances, and the irregular reverberations of the later phases. The simpler geological situation around Sulz, on the other hand, resulted in a much "cleaner" record section, with well-defined impulsive first arrivals, even though the final record section consists of several quarry blasts of different size and emplacement.

As interpretational techniques become more refined, enabling one to model complicated laterally varying structures, and interest turns to structural details all the way down to the mantle, accurate knowledge regarding the influence of the near-surface layers becomes increasingly important. The accuracy of topographic corrections depends heavily on the velocity value in the uppermost layer at each station and local variations of sedimentary velocity and thickness can contribute significantly to the scatter of travel-times, thus masking possible crustal features at greater depth. Since the angle of incidence of a seismic signal at the surface is governed by the ratio of near-surface to refractor velocity - and this influences the distribution of the total seismic energy to the amplitudes of each directional component - variations of near-surface geology also contribute to the scatter of amplitude data. While the travel-time effect could be corrected for by complementing long-range refraction profiles with a number of short lines, providing information about the near-surface velocity structure along the profile, this procedure would give only a first approximation for the amplitude effect, since local structural features can produce additional frequency-dependent resonance and interference phenomena, which influence the amplitude of a given phase. Careful selection of the recording sites and sufficient station density can, however, reduce the effect of this amplitude scatter as well.

In conclusion, it can be said that, with proper precautions during planning and execution of seismic refraction experiments, the evaluation of amplitudes can contribute significantly to a more accurate interpretation of the data.
6.2 Tectonic considerations

The Moho constitutes one of the major structural discontinuities in the lithosphere. It is defined as the transition from lower-crustal P-wave velocities of around 7 km/s to upper-mantle velocities of about 8 km/s. Depth variations of this discontinuity from one region to another are interpreted as clues for differences in tectonic regime and evolutionary history. Figure 6.1 shows the Moho depths below the southeastern Black Forest and the Southern Alps, as derived from the two seismic refraction experiments evaluated in this study, in a map with values from other investigations in and around Switzerland. Cross-sections summarizing the present knowledge about the seismic velocity structure of the crust along a north-south profile across the Alps from Basel to Como, across the Rhinegraben, the Black Forest and Urach geothermal anomaly, and along the Southern Alps from Lago Maggiore to Trento are shown in Figure 6.2.

Sulz-south

While the structure of the middle crust along the west-east cross-section in Figure 6.2b exhibits considerable lateral variation, the upper crust and the depth to the Moho are relatively uniform. In fact, the pronounced upper-crustal velocity gradient found below Sulz has also been derived from studies of local earthquakes in the Swabian Jura (Peterschmitt 1979, Mueller and Peterschmitt 1981), from a reinterpretation of two refraction profiles in the Rhinegraben (Zucca 1984), and from a combined refraction and reflection experiment in the area of the Urach geothermal anomaly situated about 60 km ENE of Sulz (Gajewski and Prodehl 1983, 1985).

The main line of the Urach refraction experiment crosses the Sulz-south profile at a distance of about 37 km from the shotpoint. Figure 6.3 shows the corresponding record sections as published by Gajewski and Prodehl (1983, 1985), together with their phase correlations and the resulting model. The point of intersection lies near Willingen, at a distance of about 60 and 90 km along profiles U2-240 and U1-240, respectively (see Fig.6.3). Thus, the two data sets from Sulz and Urach provide information about the crustal structure from almost the same area. While the gradient zones in the upper part of the crystalline basement agree well with each other, the velocities of 6.2 to 6.3 km/s, at depths between about 10 and 20 km shown in Figure 6.3, are significantly higher than the 6.0 or at most 6.1 km/s found in the Sulz data. However, an enlarged plot of the first arrivals in the Urach record sections would probably show that phase a3 in WSW direction, corresponding to Pq, could be traced to much greater distances than shown in Figure 6.3 and that a correlation with a lower velocity gives a better fit. This would bring the two models into good agreement with each other.

From the geological setting (Fig.4.2), it is to be expected that the granites and gneisses of the Black Forest extend to greater depths below the
Fig. 6.1 Crustal thickness (km) for Switzerland and neighbouring areas (modified after Kissling 1980). New values based on the Sulz-south and the SUDALP77 refraction profiles are circled.
Fig. 6.2 a) Crustal cross-section along the Swiss Geotraverse (after Mueller et al. 1980), modified below Como to include the revised Southern Alps model.

b) Crustal cross-section across S.W. Germany (after Mueller et al. 1973), modified for the Sulz-south model, and including results from Edel et al. (1975), Bartelsen et al. (1982) and Gajewski and Prodehl (1983). Locations showing evidence for a laminated crust-mantle transition are denoted by horizontal striations.

c) Crustal cross-section along the Southern Alps (profile A-01).
Fig. 6.3 Record sections and crustal model for the Swabian Jura near Urach, from Gajewski and Prodehl (1983, 1985). U1 and U2 designate shotpoints, 60 and 240 correspond to the azimuths of the profiles to the ENE and WSW, respectively, and SS marks the point of intersection with profile Sulz-south.
Sulz profile. Seismic velocities and velocity-depth gradients measured in situ depend on several factors:

1) intrinsic elastic properties of the rock, 
2) temperature, 
3) confining pressure, 
4) degree of fracturing, 
5) degree of fluid saturation, 
6) pore pressure.

When comparing field and laboratory measurements one must take all these factors into consideration (see e.g. Smithson and Shive 1975). Depending on the rock type and heat flow regime under consideration, the effect of increasing temperature and pressure, in the case of dry unfractured rocks, can produce either a slight positive or a slight negative gradient (see e.g. Hughes and Maurette 1956, Kern and Richter 1981). At low confining pressures, fractures and open pore spaces tend to lower seismic velocities, and, as a consequence, the closing of these voids under the influence of increasing pressure leads to strong velocity-depth gradients. Though qualitatively this agrees quite well with field measurements, closer examination reveals that, in general, gradients near the surface are weaker and velocities at depth lower than expected. This is due to the fact that fractures and pore spaces are more or less water saturated. At low confining pressures this causes a higher velocity as compared to that of dry rock, while at greater depth the increased pore pressure effectively keeps the cracks open, which in turn lowers the velocity (see Nur and Simmons 1969, Mueller 1977). A comparison of the Sulz results with several laboratory measurements reveals that, indeed, fractures, fluid inclusions and pore pressure must play a significant role down to depths of 5 or even 10 km.

As discussed in Chapter 4, the Sulz data does not allow to resolve unambiguously the question of structure in the middle crust. Of the Urach refraction experiment, the two profiles to the east (U1-60 and U2-60) show a very clear intermediate reflection (phase b) between the upper crustal diving wave (a3) and the FMP (c) (Fig.6.3). The very high-quality records of section U2-240, on the other hand, do not show such a clear intermediate phase. Thus the low-velocity body is limited to the thermally anomalous region around Urach (Fig.6.3). Based on this evidence, a simple model with constant velocity in the middle crust below the Sulz profile seems more likely than one with a discontinuity or even a velocity inversion.

On the other hand, it is possible that near-vertical reflection data would reveal additional structure at depths of 10 to 12 km: in the Rhinegraben, echoes with a travel-time of about 4 seconds have been interpreted as reflections from the top of a low-velocity layer in this depth range (Mueller et al. 1969, 1973), which could extend to the east below the Black Forest.

Indeed, the generalized structural model for the crust beneath the Swabian Jura derived from observations of local earthquakes includes the existence of a pronounced low-velocity zone (Peterschmitt 1979, Mueller and
Peterschmitt 1981). This means that such a low-velocity layer could be laterally discontinuous. As such, this would be quite compatible with the idea that velocity inversions are due to local intrusions of lower velocity granitic material into a prevalently gneissic crust (Mueller 1977). As discussed in detail in Sections 4.4 and 4.5, an additional alternative compatible with the data presented here would consist of a zone with negative velocity-depth gradient below about 10 km, caused by appropriate conditions of confining pressure, temperature and pore pressure (see e.g. Smith et al. 1977).

The depth of the Moho, on the other hand, seems to be rather uniform, with values around 25 to 26 km ranging over about 200 km, from the southern Rhinegraben to the region around Urach. Such a rather thin crust seems to be typical for both a rift structure, such as the Rhinegraben, and for a region with magmatic intrusions and geothermal activity, such as Urach. The Moho depth below Sulz must thus be viewed as a consequence of its proximity to these two anomalous areas.

Several other authors have postulated a laminated lower crust before, based partly on near-vertical reflection data (Meissner 1967, 1973, Fuchs 1969, Clowes and Kanasewich 1970, Davydova 1972, Bartelsen et al. 1982, Barton et al. 1984). The present study provides additional independent evidence for the existence of such a structure.

While the vertical extent of the transition zone and the general velocity increase with depth is directly supported by the data, the regularity of the lamella thickness and the strength of the inversions is largely an arbitrary artifact. Whether an existing lamella-structure is seen or not is in part a function of the wavelength of the recorded signals, which in turn puts certain constraints on the average thickness of the individual lamellae. However, the real structure is probably much less regular than suggested by this model. Moreover, nothing can be said about the lateral extent of such a structure.

Hale and Thompson (1982) review a number of seismic reflection lines which show evidence of lamination in the lower continental crust, and note that such a structure seems to be relatively discontinuous. In addition, taking the lithologic sequence of the Ivrea-Verbano zone (northern Italy) as a model of the lower crust (Fountain 1976), they computed a synthetic vertical incidence reflection seismogram, which show striking similarities to recorded ones. For comparison, this model has been superimposed on the Sulz-model in Figure 6.4. Though the two models differ in detail, the laminated sequence of high- and low-velocity layers is common to both.

The existence of a laminated crust-mantle transition several kilometers thick has been postulated not only for the Sulz-south profile, but on the basis of reflection data, also below the Urach region (Bartelsen et al. 1982). Reflection data recorded in the Southern Rhinegraben (Dohr 1970), showing a wide band of striated reflections between about 7 and 9 seconds, suggest a similar kind of lamination (Mueller et al. 1973), (see Figs. 6.2b and 6.5a). The depth of this laminated structure under the Rhinegraben would correspond roughly to that of the gradient zone between about 20 km and
the Moho in the models by Edel et al. (1975). It is thus quite possible that such a laminated crust-mantle transition extends over an even greater area.

Since the detection of a laminated structure at the base of the crust by refraction measurements is only possible under favourable circumstances (appropriate source signal, careful recording procedures and an otherwise not too complicated crustal structure), it is impossible to assess the significance of such a structure on a global level. Haie and Thompson (1982) review most of the evidence for a laminated Moho reported so far, and conclude that it might be a more common feature of the continental crust than was thought until now. The increased use of deep reflection soundings in the future will undoubtedly help to delineate those areas where it exists, thus providing an answer to its tectonic significance.

Fig. 6.4 Velocity-depth model for the lower crust of SULZ 11S (thick line) and velocity profile across the Ivrea body (thin-lined lamination), after Haie and Thompson (1982).
Fig. 6.5 a) Velocity-depth columns for the Sulz and Southern Alps models. The horizontal striations in the Sulz model denote the laminated crust-mantle transition zone.

b) Part of the crustal evolution scheme by Mueller (1978).
Southern Alps

Moho depth along the Southern Alps profile varies quite considerably: crustal thickness between 31 and 35 km is found in the region between Lago Maggiore and Lago di Como; thickness increases going east-northeast into areas with greater topographic elevations, reaching a maximum of 46 km below the Adamello massif (see Figs. 6.2c and 6.5a). To the southeast, on the other hand, there is a sudden updoming of the mantle, which coincides with the Tertiary volcanism of the Colli Euganei and causes the prominent gravity maximum around Verona. The significance of this mantle bulge is difficult to assess without more detailed knowledge regarding the crustal thickness associated with the Po Plain.

At first glance, total crustal thickness below shotpoint A, in the vicinity of the southern tip of Lago Maggiore, would suggest a normal continental crust. However, it is important to note that the basement is reached only at a depth between 7 and 8 km beneath the sediments of the Lombardy trough. If the thickness of this sedimentary cover amounted to a more usual value between 2 and 3 km, the Moho would come to lie at a depth of about 26 km, which is comparable to that below Sulz or below the Rhinegraben. This means that the crystalline part of the crust thickens by more than 10 km between Lago Maggiore and Lago di Como, which are only 50 km apart, and by another 10 km towards the Adamello-Giudicaria region. This rather dramatic effect is illustrated by the velocity-depth columns in Figure 6.5a. One is thus inclined to consider the possibility of a categorical difference between structure and evolution of the crust beneath western Lombardy on one hand and below the more mountainous parts of the Southern Alps on the other.

Several comparative schemes of crustal structures have been devised to classify different tectonic units and to describe the evolutionary process linking a typical continental crust to an oceanic one (see e.g. Berckhemer et al. 1975, Mueller 1978, Smithson et al. 1981). In an attempt to classify the results derived from the Sulz-south profile and for the Southern Alps, as shown by the velocity-depth columns in Figure 6.5a, part of the evolutionary scheme proposed by Mueller (1978) is reproduced in Figure 6.5b.

As one would expect, the results from the Sulz-south profile for the region just east of the Rhinegraben, fit very nicely into this scheme, as a transition between a "normal" continental crust and a graben structure.

The results from the Southern Alps, on the other hand, are difficult to fit into a unified interpretation. The mountainous parts of the Bergamasc Alps and the Adamello massif must be viewed in terms of the Alpine orogeny and do not fit into this scheme of crustal evolution. The area under western Lombardy on the other hand, characterized by the small overall thickness of the crystalline part of the crust, and, in addition, by the thin mid- and lower-crustal layers with velocities around 6.4 and 7.1 km/s, would fit into a similar position as the Sulz data. This would mean that
Fig. 6.6 a) Palinspastic reconstruction of Alpine orogeny from Lower Cretaceous to present (after Truempy 1980 and Buechi and Truempy 1976). Arrow denotes position of Southern Alps profile. B=Basel, L=Lugano, A=Aare, T=Tavetsch, G=Gotthard, Br=Briançonnais, M=Margna, O=Austroalpine, S=Southern Alps.

Fig. 6.6 b) Palinspastic reconstruction of Eastern Alpine orogeny from Late Permian to Early Cretaceous (after Dietrich 1980). Arrow indicates presumed position of Southern Alps profile.
the Lombardy trough constitutes a relic of an ancient rift structure.

The Southern Alps must be considered as part of the Adriatic-African plate, which, during the Late Jurassic and Lower Cretaceous was separated from the Eurasian plate to the north by the actively rifting South Penninic Ocean. It is thus tempting to explain the anomalous crustal structure below the Lombardy trough in terms of the southern margin of this oceanic basin. However, according to most palinspastic reconstructions of the Alps (see e.g. Fig.5.6a), that zone which now constitutes the Southern Alps was separated from the South Penninic Ocean by the Austroalpine domain, which, during orogeny, was thrust over the Penninicum and now lies north of the Insubric Line (see e.g. Buechi and Truempy 1976, Truempy 1980). In a palinspastic reconstruction of the Eastern Alps, Dietrich (1980) postulates, on geological evidence, an additional "southern rift system" within the eastern part of the Southern Alps, which was active during late Permain and Triassic age (Fig.5.6b). Thus the abnormal crust beneath the Lombardy trough may be considered as evidence for an analogous ancient rift system in the western part.

Indeed, geologic evidence suggests two distinct pre-Jurassic rifting phases in western Lombardy. A first one was formed during Late Hercynian times, with largely WSW-ENE trending structures, documented by the Lugano volcanites, and is analogous to or even a continuation of the southern rift system in the reconstruction by Dietrich (1980). A second one is of Upper Triassic and Liassic origin, characterized by several deep sedimentary basins separated by wells forming a horst and graben structure delineated by N-S striking fault systems. A possible palaeotectonic reconstruction of the sedimentary basins during the Jurassic, along the whole Southern Alps profile is reproduced in Figure 6.7a (Bernoulli 1980), and in more detail for the area around shotpoint A between Lago Maggiore and Lago di Lugano, in Figure 6.7b (Kaelin and Truempy 1977). It is thus very likely that the thinning of the continental crust associated with these rifting phases has been preserved beneath western Lombardy until present times in the form of an abnormally thin crystalline part of the crust, as revealed by the seismic results. To which of the two rifting phases just discussed the observed crustal thinning corresponds cannot be determined with certainty. However, it is possible that it originated already in the Hercynian as an upwelling of the mantle accompanied by the evidenced volcanic activity, and that, under a subsequent tectonic regime during Triassic-Liassic times, this area with an already weakened crust could subside to accommodate the sedimentary basin still visible today. The complete role of the Lombardy Basin in the course of evolution of the Southern Alps can only be assessed properly on the basis of more knowledge about the crustal structure further west in the transition zone to the Ivrea Body and further south towards the Po Basin.

Further to the east along the Southern Alps profile, subsequent orogenic activity, as a result of the continental collision between Eurasia and Africa, then led to a thickening mainly of the lower crust and to an uplift of those sediments which now constitute the Bergamasco Alps and the Dolomites. As discussed in detail in Chapter 5, evidence for a simple overthrust of the African plate onto the Eurasian one, in the sense of a
Fig. 6.7 a) Paleotectonic reconstruction of the Southern Alps (after Bernoulli 1980). Stars denote positions of shotpoints of profile A-01.

Fig. 6.7 b) Paleotectonic reconstruction of Western Lombardy (after Kaelin and Truempy 1977). The star and dashed line in the inset indicate the position of shotpoint A and profile A-01.
crustal doubling hypothesis (e.g. Giese et al. 1982), could not be found. It is thus more likely that the lower-crustal material was either thickened in situ as a result of the continental collision or that it forms a seismically amorphous melange of northern and southern crustal material interpenetrating each other as suggested by Mueller et al. (1980).

It is interesting to note that all refraction profiles recorded so far in and around the Alps exhibit clear PMP-reflections and provide no evidence for any interruptions of the Moho, at least of a dimension that is resolvable by refraction measurements, which corresponds to one or two kilometers. At first glance, it is difficult to imagine that the subduction mechanism associated with the continental collision, as proposed by several authors (e.g. Laubscher 1974, Panza and Mueller 1979), would not destroy the Moho at least in some places. However, it is possible that the Moho, with its large density contrast, is the level of differentiation at which subduction during a continent-continent collision can occur: the denser upper-mantle material is forced downward, while the lighter material above the Moho is partly pushed together, thus causing a thickening of the lower crust, and partly ejected upwards, forming the complicated system of nappes at the surface (see e.g. Miller et al. 1982).

Fig. 6.8 Epicenter map of recent earthquakes (1971-1975) in the Central, Southern and Eastern Alps (after Bonjer and Gelbke 1981). The hatched area indicates the relatively aseismic Southern Alps region.
This, however, gives neither an explanation for the structural asymmetry between the two continental blocks north and south of the Insubric Line (compare the structural differences between Central and Southern Alps in Fig. 6.2a), nor for the role of this important lineament in the course of the Alpine orogeny. An answer to these questions should be provided by more detailed investigations of the transition zone between the Central and Southern Alps, but methods, such as reflection profiling, with greater resolving power than refraction seismology, must be resorted to, in order to delineate structural changes over such short distances.

A look at the seismicity map in Figure 6.8 (Bonjer and Gelbke 1981) shows the area of the Southern Alps as a largely aseismic zone, bounded to the north by a band of seismic activity associated with the overthrust of the Calcareous Alps onto the northern foreland Molasse and to the south by a line extending through the Friuli earthquake region into the Po Plain. The time period covered by the seismicity map in Figure 6.8 is very short; however, compilations of historical earthquakes corroborate the aseismic nature of the Southern Alps. This, together with the relatively undisturbed geology, at least compared to the situation north of the Insubric Line, gives the Southern Alps the appearance of a stable continental block acting like a bumper of the Adriatic-African plate in the orogenic collision process. Fault plane mechanisms of the 1976 Friuli earthquakes constitute conclusive evidence for an underthrusting of the southern foreland beneath the South-Alpine block (Müller 1977a, Gebrande et al. 1978, Wittlinger and Haessler 1978). The direction of this mechanism is exactly opposite to that expected from the crustal doubling hypothesis, advocated by Giese et al. (1982), and must be viewed as further evidence against the latter.
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Given strong variations in topography along a profile it is often necessary to correct arrival times for elevation differences. Usually, all recording stations are reduced to the average elevation along the profile, or, if one end of the line is significantly higher than the other, to a regression line through the elevation versus distance plot. For a refracted arrival, the correction $T$ is given by the well known formula,

$$T = \frac{H}{V_1} \cos i$$

where $H$ is the elevation difference, $V_1$ the surface velocity and $i$ the angle between the emerging ray and the vertical (see e.g. Dobrin 1976). This can be rewritten to give,

$$T = H \sqrt{\frac{1}{V_1^2} - \frac{1}{V_a^2}}$$

where $V_a$ is the apparent velocity of the refracted wave. Assuming $H = 1$ km, the values of $T$ as a function of $V_1$ are shown by the curves in Figure A.1, each of which corresponds to a different refractor velocity $V_a$. The velocity $V_a$ is usually known quite well from the apparent velocity of the arrivals, while the surface-layer velocity is known only near the shotpoint, and, unless additional data is available, it can only be estimated very roughly beneath each station.

The diagram enables both the correction and its uncertainty to be determined at a glance. For example, given $V_a = 6.0$ km/s and $V_1 = 4.5 \pm 0.5$ km/s, an elevation difference of 1 km corresponds to an arrival time correction of $0.15 \pm 0.03$ seconds. The correction for a different elevation is obtained by multiplying these values by the actual elevation difference $H$. Shot elevations can of course be corrected in the same way, and the above formula applies to diving waves as well as to pure head-waves. However, for reflected waves it is valid only in the vicinity of the critical point.
Fig. A.1 Elevation correction curves for refractions. 
$V_a$=refractor velocity, $V_1$=surface layer velocity, 
$T/H$=travel-time correction for 1 km elevation.
APPENDIX B. THE DIGITIZING SYSTEM

The digitizing system employed at the Institute of Geophysics of the ETH-Zuerich to digitize seismograms recorded on FM magnetic tape was implemented between 1979 and 1980. In its present form it is subdivided into two separate parts: the digitizing system proper, based on a Hewlet-Packard (HP) mini-computer system at the Abteilung fuer Industrielle Forschung (AFIF) of the Institut fuer Elektrotechnik of the ETH-Zuerich, and a data handling program package, based on the main CDC-computer of the ETH computing center.

The digitizing system proper was developed along the lines of an analogous system designed by W. Kaminski at the University of Karlsruhe, and was adapted and optimized for the hardware available at the ETH-Zuerich on the basis of a rudimentary program package written by A. Hardmeier for the above mentioned HP-computer. Though the programs had to be written for the particular input-output and file handling structure of the computer used (HP 21MX with a RTE-VI operating system), it is sufficiently flexible to have been transferred successfully to a Digital PDP-11 computer at the University of Utah and to a hybrid system in Milan. A similar system, also developed on the basis of the one in Karlsruhe, has been implemented at the University of Munich (see Stroessenreuther 1982).

The following presentation is subdivided into two parts: first an overview of the whole hard- and soft-ware in form of flow-diagrams with brief explanations of the function of each program, and second, an explanation of the computational methods behind the time decoding and filter programs.
PLAY-BACK AND DIGITIZING HARDWARE (Fig. B.1)

Analog part:

The frequency modulated multiplexed recordings of the MARS instruments are played back, demultiplexed and demodulated in the standard way. The frequency counter and oscilloscope are used to adjust the tape speed and head position of the play-back tape recorder for optimum signal quality. Before input into the analog-digital converter, the signals are amplified to the ±10 volts required as input by the converter. This amplification must be adjusted individually to the level of each seismogram as displayed on the strip-chart recorder and the value must be entered interactively into the digitizing program. The strip-chart recorder is also used to determine the onset time of the digitized sequence if the time signal is not a coded one or if it is too poor to be decoded automatically. The pilot frequency of 6.4 kHz is reduced to 400 Hz and used to drive the A-D converter. Possible interruptions of the pilot signal are bridged by a phase-locked-loop (PLL) circuit, thus always insure a continuous digitizing sequence.

Digital part:

In the present configuration, the peripheral units are connected to two separate computers, which are linked to each other in a master-slave mode. The A-D converter has a resolution of 10 bits, so that ±10 volts input correspond to ±511 counts. The digitized data are transferred from the converter onto disc, processed there and then stored on digital magnetic tape. The digitizing procedure is initiated and monitored from a single graphics terminal, and a permanent record of the digitized signals is provided in form of a raw plot on the electrostatic printer-plotter.
Fig.B.1 Digitizing hardware.
THE DIGITIZING PROGRAMS (Fig. B.2)

ADS: master program which initiates the digitizing procedure and schedules the subroutines and programs with the necessary parameters for steps 1 through 8.

TTRV: subroutine of ADS which transfers data from the A/D converter to file DATAFI on disc.

INFR: subroutine of ADS which searches file INFOFI for the corresponding header information and transfers it to the first record of file DATAFI.

TIME: program calling subroutines TSTAC and TCODE.

TSTAC: subroutine of TIME which calculates the exact sampling rate relative to the time signal and the second fraction of the time of the 0-th digit. The graphic output is plotted on the graphics terminal and stored in file STACFI (not shown).

TCODE: subroutine of TIME which decodes DCF and MSF time signals to determine the complete date and time of the O-th digit. The results of TSTAC and TCODE are automatically written onto the header record of file DATAFI.

ROPLO: produces a control plot of the raw data (all three seismic channels and time signal) and of the complete header information as well as of the graphics output of program TIME which was stored in file STACFI.

DSEE: displays any desired signal interval on the graphics terminal and allows errors in the data, such as spikes, to be corrected interactively. It can also display the digit number and exact time of any desired sample.

REDU: separates the 4 multiplexed channels on file DATAFI, filters the seismic signals with a 32 Hz low-pass filter, reduces the sampling rate to 100 Hz and creates three new single channel files containing only the signal interval of interest. The time signal is discarded.

TADUM: stores the final data on magnetic tape and updates the tape directory on file TAPEFI.

INFO: converts the binary header information of the data files into a coded file (HEADFI) and back.

EDIT: system program with which the header information in file HEADFI can be edited if necessary.

Additional subroutines used by these programs:

HEAD and TIBL: see Fig. B.3.
Fig. B.2 Flow-diagram of digitizing procedure.
PROGRAMS TO PLOT RECORD SECTIONS ON THE HP-COMPUTER (Fig. B.3)

PROPL: master program to plot record sections on the HP-computer. Input and output parameters are stored on files IPROFI and OPROFI.

TADUM: reads each datafile from tape and stores it in file DATAFP, from which program PROPL picks the relevant signal intervals for the record section.

HEAD: reads and writes the binary header information.

TIBL: converts the given time value, in seconds, of a sample to the corresponding record and word numbers.

VERSAPLOT routines:

PLTSP and PLTS3: convert plot calls to unsorted vectors stored on file VECTR1.

SORT: transforms the unsorted plot vectors in file VECTR1 to a sorted file VECTR3 suitable for output on the VERSATEC plotter, using scratch file VECTR2.

RASS: vector to matrix conversion and output to the VERSATEC plotter. File PARM contains the corresponding processing parameters.
Fig. B.3 Flow-diagram of record section plot on Versatec.
DATA HANDLING SOFTWARE ON THE CDC-COMPUTER (Fig. B.4)

COPHTFF: reads data from a Hewlett-Packard 7-track tape and stores them in a sequential discfile on the CDC-computer of the ETH computing center. Each 16 bit HP-word is converted to a 60 bit CDC integer. (Program courtesy of AFIF, ETH-Zuerich)

HPSEIS: converts those integers corresponding to characters or real numbers in the signal headers into their correct CDC representation, and transforms the sequential output of program COPHTFF into a binary random access file.

INFO: reads, displays and, if necessary, corrects the binary header information of each signal.

NEWFI: creates a new file with selected signals from a single old file or merges two old files into a new one.

FSFQ: plots any desired signal interval and calculates the corresponding amplitude, power or phase spectra, using subroutine RELFFT, written by F. Bonzanigo (software library of the Institut fuer Technische Physik, ETH-Zuerich).

PROFL: plots record sections from data stored on binary random access files.

BINCODE: converts data on binary random access files on the CDC computer to coded sequential files, transferrable to other computers.

PROFSEQ: plots record sections from data stored on coded sequential files.

FILTR: subroutine of programs PROFL and PROFSEQ: first order recursive, zero-phase, band-pass filter.

DEPOSIT/REVIVE and SCOPY: system programs on the ETH computing center to transfer files between disc and tape.

Additional subroutines used by these programs:

GETLABL: subroutine to read label and directory record of the random access files.

GETPOS: subroutine which reads the positions of the desired signals on a random access file in various input formats and converts them to the corresponding record numbers as stored in the file directory.

GETSEI: subroutine to read the desired seismogram interval and header from a random access file.

HEAD and TIBL: see Fig. B.3.
Fig. B.4 Data-handling and plotting software on CDC.
COMPUTATIONAL METHODS

The following gives more detailed explanations of the computational methods behind some of the key programs employed in the digitizing system.

Program ROPLO:

As illustration, Figure B.5 shows a raw-plot of the three components of a digitized seismogram plus time signal, as it is plotted by program ROPLO on a VERSATEC electrostatic printer-plotter. The digitized data are stored on disc in multiplexed form in the following sequence:

$$S(1,1) S(1,2) S(1,3) S(1,4) S(2,1) S(2,2) S(2,3) \ldots S(I,J)$$

where $I =$ sample number, $J =$ channel number.

Sorted in this manner, they are thus ideally suited for being displayed on a matrix plotter, which plots one line after another as the paper is fed out of the machine, each line containing the values corresponding to all four channels at a time. Program ROPLO drives the plotter directly by assigning the amplitude value of each trace to the proper bit in an integer array corresponding to the nibs of the electrostatic plotter. This means that it is not necessary to employ the complex and time consuming VERSAPLOT software shown in Figure B.3 for plotting record sections. The program does not interpolate, so that the maximum time scale is limited by the resolution of the plotter (with 200 nibs/inch this corresponds to 200 samples/inch, or at a sampling rate of 400 Hz, to 5.08 cm/s). However, it is designed in such a way that it can compress the time scale without losing the information of any data points. Thus, even high-frequency signals are plotted with full detail and at great speed, for which a conventional pen-plotter needs a long time and large amounts of ink.
Fig. B.5 Control plot as produced by program ROPLO on Versatec.
Program TIME:

Since neither the exact sampling rate nor the time of onset of the digitized sequence is known beforehand, program TIME analyses the recorded and digitized time signal to determine the deviation of the true sampling rate (ABTF), which corresponds to the actual value of the pilot frequency in the recording instrument, from the nominal one (IDFQ = 400 Hz) and to decode the date and time of the 0-th digit (TO). The fictitious 0-th digit was chosen as reference point, such that the time of the I-th digit is given by:

\[ T(I) = TO + I/ABTF. \]

The time signal is composed of a sequence of second marks, consisting of rectangular pulses of variable duration at the onset of each second. In the DCF time signal, discussed here, the information regarding year, month, day, hour and minute is coded by a variable succession of short (0.1 s) and long (0.2 s) pulses. The minute mark is identified by the fact that the 59th second pulse is always missing (see Fig. B.5).

---

**TIME SIGNAL STACK OF: R41-4E-18**

<table>
<thead>
<tr>
<th>DAUR</th>
<th>ISEC1</th>
<th>ISEC2</th>
<th>IDFQ</th>
<th>ABTF</th>
</tr>
</thead>
<tbody>
<tr>
<td>90.804</td>
<td>160</td>
<td>180</td>
<td>400</td>
<td>400.333</td>
</tr>
<tr>
<td>TO</td>
<td></td>
<td></td>
<td>84.128 3:35 26.6003</td>
<td></td>
</tr>
</tbody>
</table>

![Stacked time signal of the first and last 30 s of the digitized sequence.](image)

---

Fig. B.6 Stacked time signal of the first and last 30 s of the digitized sequence.
Program TIME is subdivided into two separate parts: subroutine TSTAC calculates the sampling rate and the fractional part of the second corresponding to the O-th digit, while subroutine TCODE decodes the time signal to determine the remaining information.

Subroutine TSTAC:

In principle, the sampling rate is calculated by determining the samples corresponding to the onsets of the first and last time marks in the digitized sequence, and then dividing the number of samples between them by the corresponding number of second intervals. In order to automate the procedure and to insure sufficient reliability even with poor reception of the time signal, the signal-to-noise ratio must be improved: by using the nominal sampling rate (400 Hz) as interval length, 30 one-second intervals each at the beginning and end of the digitized sequence are summed in a stacking procedure. The true sampling rate is calculated from the relative shift of the onsets of the two stacked second intervals and from the time difference between them.

The computational procedure is as follows (refer to Fig. B.5):

1) perform stacking operation over the two time intervals:

\[
\text{LSTAC} = \sum_{K=1}^{\text{LSTAC}} T(I+1) + IDFQ - IDFQ \quad (I=1, IDFQ)
\]

\[
T = \text{original time signal samples}
\]

\[
\text{TS} = \text{stacked time samples}
\]

\[
\text{LSTAC} = \text{number of stacked second intervals} = 30
\]

\[
\text{IDFQ} = \text{nominal sampling rate} = 400
\]

2) find the onset of each stacked time mark:

2.1 determine the maximum and minimum value of each stacked second interval,

2.2 perform a rough search for the rising flank of each second mark, by determining point IS, where the value of the time stack first surpasses \((\max+\min)/2\).

2.3 in the interval between IR and IS, where \(\text{IR} = \text{IS} - \text{IDFQ}/20\), calculate the second derivatives (IDD) according to the formula,

\[
\text{IDD} = \text{TS}(I+2) - 2*\text{TS}(I+1) + \text{TS}(I) \quad (\text{I=IR,IS}),
\]

The onsets of each stack (ISEC1 and ISEC2) are the points in each interval where IDD is a maximum.
3) calculate the true sampling rate ABTF:

\[
ABTF = IDFQ + \frac{(ISEC2 - ISEC1)}{DT}
\]

\[
DT = INT(DAUR) - LSTAC \quad \text{time difference}
\]

\[
DAUR = DIGS/IDFQ \quad \text{signal duration}
\]

\[
DIGS \quad \text{number of samples in signal}
\]

In the example in Figure B.6:

\[
ABTF = 400 + \frac{(180 - 160)}{60} = 400.333 \text{ Hz}
\]

4) calculate second fraction of the time of the 0-th digit:

\[
\text{if } ABTF > IDFQ, \ TO = 1 - \frac{ISEC1}{ABTF}
\]

\[
\text{if } ABTF < IDFQ, \ TO = 1 - \frac{ISEC1}{ABTF} - LSTAC \times \frac{(IDFQ - ABTF)}{ABTF}
\]

For the case of strongly disturbed or weak time signals, the program has an option for interactive determination of the onsets of the time stacks on the graphics terminal.

Subroutine TCODE

The procedure to decode the DCF time signal is based on the determination of which time pulses are 0.1 s and which are 0.2 s long. This is done as follows:

1) starting from the position of the first second mark, which is known from the value of TO calculated by subroutine TSTAC, and incrementing it each time by a number of samples equal to the sampling rate, integrate each time pulse over a duration of 0.2 s. This gives a value for the area of each time pulse \( A(I) \).

2) determine the maximum and minimum value of all these areas.

3) normalize the areas as follows:

\[
A = -1 \quad \text{for } \min \leq A < \min + \frac{(\max - \min)}{4}
\]

\[
A = 0 \quad \text{for } \min + \frac{(\max - \min)}{4} \leq A < \max - \frac{(\max - \min)}{4}
\]

\[
A = 1 \quad \text{for } \max - \frac{(\max - \min)}{4} \leq A \leq \max
\]

Thus, a missing second mark corresponds to -1, one of 0.1 s duration to 0 and one of 0.2 s to 1.
4) using the information about the date and time of the shot, stored in the header record, set the elements of a second array B to -1, 0 and 1.

5) cross-correlate array A and B. The position of the maximum of the cross-correlation gives directly the value of the whole second corresponding to TO and also the position of the minute mark.

6) knowing the position of the minute mark, decode the remaining time information by determining the positions of the elements equal to 1 in array A.

7) with the shot time and distance between shot and receiver stored in the header record, calculate the time between the 0-th digit and 0 s in reduced time:

\[ RSEC = TS - TO + \frac{DIST}{VR}. \]

8) check the results against the known shot time and if consistent, enter the results into the header record.

Using the information regarding date and time of the shot, and crosscorrelating a synthetic time signal with the recorded one, assures a large degree of reliability: even with several second marks missing because of bad signal reception, the correct position of the minute mark can be found. In addition, a number of checks allow the program to spot errors and ask for immediate corrective action from the operator via input from the terminal. An analogous subroutine was written to decode an MSF coded time signal.

Program DSEE

Defects of the recording system (e.g. dust on the tape recorder head and faulty tapes) or disturbances of the carrier frequencies can cause sharp spikes or drop-outs to occur on the record (see Figs. B.5 and B.7a). If not removed, these spikes can make it impossible to normalize the amplitudes to the maximum value of each trace upon plotting the final record section. As illustrated in Figure B.7b, low-pass filtering alone will not remove them; on the contrary, by reducing the amplitude and broadening the pulse, it can become indistinguishable from a seismic phase, thus falsifying the record. Program DSEE, which allows any desired data interval to be displayed on the graphics terminal, is used to remove such spikes.
Figures B.7c and d illustrate the procedure: the spike is displayed with sufficient magnification, the samples to be modified are entered directly on the screen by moving the cursor to the desired position, and the cleaned data sequence is written back onto the file. If desired, the program can also perform a linear interpolation across the spike; in this case, only the two endpoints of the spike need to be specified with the cursor.

Fig.B.7 Signals as displayed by program DSEE on a graphics terminal.

a) Enlargement of spike on vertical component trace in Fig.B.5.

b) The same spike filtered with a 32 Hz anti-aliasing filter; note broadening and addition of negative side-lobes.

c) Enlargement of the unfiltered spike: the samples to be modified (denoted by small dots) were entered with the cursor.

d) same as (c) after removal of the spike; the dots at the bottom mark the position of each sample in (c) and (d).
Program REDU

In order to avoid aliasing during digitizing and to increase the accuracy of the time determination, the signals are digitized at a relatively high sampling rate (usually 400 Hz). After the time of onset has been determined and possible spikes have been removed, the sampling rate can be reduced (usually by a factor of 4) without losing any significant information. This is done by program REDU. Again, however, aliasing due to possible high-frequency noise is to be avoided, and therefore, the signals must be subjected first to an anti-aliasing filter. The impulse and frequency response of the low-pass filter used for this purpose are shown in Figures B.8a and b. Its coefficients were calculated with program PARKS, developed by McClellan et al. (1973). It is essentially flat up to 32 Hz and provides more than 60 db damping beyond 50 Hz, which is the Nyquist frequency of the reduced data. Since its impulse response is symmetric, once the time delay of half the filter length is subtracted, no additional frequency dependent phase shift is introduced. However, very sharp first arrivals at short distances from the shotpoint are somewhat broadened and are sometimes preceded by a few fictitious wavelets, corresponding to the side-lobes of the filter.

The filtering procedure itself is accomplished by a convolution in the time domain, applying the filter only to every fourth sample. This is the most time consuming step in the whole digitizing sequence; it could be accelerated significantly by the use of a recursive filter, with a somewhat more gradual high-frequency cut-off.

Program PROFL

In order to suppress unwanted noise or enhance particular frequency components of the signal, an additional low-, high- or band-pass filter can be applied to the data, during plotting of the record sections, which is done with program PROFL.

The algorithm used corresponds to a first order recursive filter, for which the coefficients can be calculated easily according to any desired specifications. As illustrated in Figure B.9 for a band-pass filter between 1 and 8 Hz the cut-off is very smooth, but it has the advantage of being computationally very efficient. By applying it to the signal the same number of times in each direction, i.e. forewards from the first to the last sample and backwards from the last to the first, a frequency dependent phase shift can be avoided.
Fig. B.8 a) Impulse response of the 32 Hz anti-aliasing filter.

Fig. B.8 b) Frequency response of the 32 Hz anti-aliasing filter.

Fig. B.9 Frequency response of a 1-8 Hz band-pass filter.
APPENDIX C. THE REFLECTIVITY METHOD

The following is a brief summary of the main steps in the derivation of the reflectivity method for the computation of synthetic seismograms, and a discussion of some of the parameters as used in the calculations in this study.

From the stress-strain relationships which govern the behaviour of elastic media one can show that the particle displacement due to an elastic disturbance traveling through a body is described by the wave equation. In general, the material properties, defined by the density, the compressional- (P) and shear-wave (S) velocities are a function of the spatial coordinates. In the method under consideration, the structure of the earth is assumed to vary only with depth and is approximated by stacks of horizontal layers, each with constant elastic parameters. Thus, the problem is reduced to the consideration of two-dimensional solids and in this case, the equations of motion are expressed in terms of two wave equations for the displacement potentials in the i-th layer:

\[ \nabla^2 \phi_i = \frac{1}{\alpha_i^2} \frac{\partial^2 \phi_i}{\partial t^2} \quad \text{for compressional waves} \]

and

\[ \nabla^2 \psi_i = \frac{1}{\beta_i^2} \frac{\partial^2 \psi_i}{\partial t^2} \quad \text{for vertically polarized shear waves}, \]

where the displacements are given by,

\[ u_i = \frac{\partial \phi_i}{\partial x} - \frac{\partial \psi_i}{\partial z} \quad \text{(horizontal component)} \]

and

\[ w_i = \frac{\partial \phi_i}{\partial z} + \frac{\partial \psi_i}{\partial x} \quad \text{(vertical component)}, \]

with \( \alpha_i \) the compressional and \( \beta_i \) the shear wave velocities in the i-th layer.
The plane-wave solution for the case of a compressional wave can be written in terms of its potential as,

\[ \phi_i(t,x,z) = A_i \exp\left[j(\omega - k_i x - l_i z)\right] + B_i \exp\left[j(\omega + k_i x - l_i z)\right] \] (3)

with \( k_i \) = horizontal wavenumber, \( l_i \) = vertical wavenumber and \( \omega \) = circular frequency, which are related to the angle of incidence \( \varphi \) as follows:

\[ k_i = \frac{\omega}{\alpha_i} \sin \varphi_i ; \quad l_i = \frac{\omega}{\alpha_i} \cos \varphi_i ; \quad k_i^2 + l_i^2 = \frac{\omega^2}{\alpha_i^2} \] (4)

This is an expression for a plane harmonic wave, with \( A_i \) the incident and \( B_i \) the reflected wave in the \( i \)-th layer.

The solution for the shear potential \( \psi_i \) is analogous.

The boundary conditions for the determination of the constants of integration in the solution of the wave equations follow from the requirement that stress and displacement be continuous across the interface between two layers. The explicit formulation of these boundary conditions allows the amplitude in the \( i \)-th layer to be expressed in terms of that in the \((i-1)\)-th layer. This can be written in matrix form as follows:

\[ \begin{pmatrix} A_i \\ B_i \end{pmatrix} = m_i \begin{pmatrix} A_{i-1} \\ B_{i-1} \end{pmatrix} \] (5)

where \( m_i \) is the so-called layer matrix, containing all the reflection and transmission coefficients for the interface between the \( i \)-th and the \((i-1)\)-th layer. To obtain the response of the \( n \)-th layer in terms of the first, it is sufficient to multiply the matrices \( m_1, m_2, \ldots, m_n \) with each other, giving a product matrix \( M \) such that:

\[ \begin{pmatrix} A_n \\ B_n \end{pmatrix} = M \begin{pmatrix} A_1 \\ B_1 \end{pmatrix} \] (6)

with \( B_n = 0 \), since there is no reflection from infinity. This is the so-called Thomson-Haskell matrix formalism. After rearrangement of the elements of \( M \), one obtains the so-called reflectivity, \( R = B_1/A_1 \), which is the ratio of reflected to incident amplitudes for the whole stack of layers and from which this method derives its name.
For the simple case of a pure P-wave reflected from a single interface between two solid half-spaces, the formalism reduces to the following expression:

$$\begin{pmatrix} A_2 \\ B_2 \end{pmatrix} = \frac{1}{2 l_2 s_2} \begin{pmatrix} l_1 s_2 + l_2 s_1 & -l_1 s_2 + l_2 s_1 \\ -l_1 s_2 + l_2 s_1 & l_1 s_2 + l_2 s_1 \end{pmatrix} \begin{pmatrix} A_1 \\ B_1 \end{pmatrix}$$  \hspace{1cm} (7)

$l$ is defined in (4) and the density $\rho$ enters through boundary conditions involving the stresses. Since there is no energy reflected from below the second half-space, $B_2 = 0$ and the plane-wave reflection coefficient is given by:

$$R_{pp} = \frac{B_1}{A_1} = \frac{l_1 s_2 - l_2 s_1}{l_1 s_2 + l_2 s_1}$$  \hspace{1cm} (8)

with (4) this can be written as:

$$R_{pp} = \frac{s_2 \alpha_2 \cos \varphi_1 - s_1 \alpha_1 \cos \varphi_2}{s_2 \alpha_2 \cos \varphi_1 + s_1 \alpha_1 \cos \varphi_2}$$  \hspace{1cm} (9)

where $\varphi_1$ is the angle of the incident and reflected ray above, and $\varphi_2$ is the angle of the ray refracted below the interface, in accordance with Snell's law. For vertical incidence, this reduces to the well-known formula for the acoustic impedance:

$$R_{pp} = \frac{s_2 \alpha_2 - s_1 \alpha_1}{s_2 \alpha_2 + s_1 \alpha_1}$$  \hspace{1cm} (10)

For the case of a single interface between two half-spaces, the reflectivity is only dependent on the material properties above and below and on the angle of incidence. For multiple layers, on the other hand, the layer matrices contain additional terms of the form,

$$\exp(j \omega \alpha_i d_i) = \exp(j \omega \alpha_i d_i \cos \varphi_i)$$

where $d_i$ is the thickness of the $i$-th layer. This leads to a frequency dependence of the total reflectivity for a stack of layers, due to interference effects of the reflected and transmitted waves from one boundary with those of another. This matrix algorithm automatically takes into consideration the contribution of all possible multiples and conversions affecting the final amplitude.

In practice, the seismograms with which the synthetics are to be compared represent the response of the medium to spherical waves, which, in
the distance range from the source and for the model dimensions under
corpus, cannot be simulated correctly by a plane-wave response. The
compressional potential of a spherical wave, which is the solution of the
wave equation has the following general form:

\[ \phi(r, z, t) = \frac{1}{D} \int f(t - \frac{D}{\alpha}) \]

where \( r \) = radial distance, \( z \) = vertical distance from the source,
and \( D = \sqrt{r^2 + z^2} \).

Its Fourier transform is

\[ \overline{\phi}(r, z, \omega) = F(\omega) \frac{e^{-j\frac{\omega D}{\alpha}}}{D} \]

where \( F(\omega) \) is the frequency domain representation of the source function.

With the aid of the Sommerfeld integral, the Fourier transform of the
spherical wave potential can be rewritten as follows:

\[ \overline{\phi}(r, z, \omega) = F(\omega) \int_0^\infty J_0(kr) e^{-jlz} \frac{k}{j^l} dk \]

where \( J_0(kr) \) is the Bessel function of the first kind and order 0, while \( k \)
and \( l \) are the horizontal and vertical wavenumbers. This is equivalent to
synthesizing a spherical wave front out of a sum of infinitesimal plane wave
segments, each with a different angle of incidence.

For a given model with the corresponding velocities and densities \( \alpha_i, \beta_i, \gamma_i \)
the total reflectivity \( R = R(\omega, k) \) is a function of frequency and wavenumber
(the latter being equivalent to the angle of incidence). If one regards the
integrand in equation (13) as the plane wave contribution to the incident
spherical wave, the corresponding reflected wave is obtained simply by
multiplying with \( R(\omega, k) \). By means of an inverse Fourier transformation, one
finally obtains the potential for the reflected spherical wave in the time
domain:

\[ \phi_r(r, z, t) = \int_{-\infty}^{+\infty} F(\omega) e^{j\omega t} \int_0^\infty \frac{k}{j^l} R(\omega, k) e^{-j\frac{\omega D}{\alpha}} dk d\omega \]

An analogous argument leads to the shear wave potential. The displacements
and thus the desired synthetic seismograms are calculated from equations
(2).
In practice, the two integrations, corresponding in effect to a double Fourier transformation from the frequency-wavenumber domain to the time-distance domain, are performed numerically on a computer. For computational reasons, the first integration is actually performed over the angle of incidence or slowness instead of the wavenumber. The limits of integration are chosen according to the requirements of the problem: the frequency limits are given by the desired frequency content of the source signal, and the slowness limits are governed by which phases out of the whole seismogram one wishes to model. The most time consuming part of the calculation consists of the computation of $R(\omega,k)$, which must be repeated for each frequency and each wavenumber step, and must run over the entire stack of layers each time.

Since the first versions of the method, as developed by Fuchs (1968) and by Fuchs and Müller (1971), based on earlier work by Thomson (1950), Haskell (1955), Harkrider (1964) and Dunkin (1965), numerous improvements and extensions have been made (Kennett 1975a, Kind 1976, 1978, 1979, Baumgardt 1980, Kind and Odom 1983). The computer program used in this study was written by Kind (1978) with the possibility of taking into consideration anelastic attenuation (Kennett 1975b), and was adapted to run on a CDC-computer by L.W. Braille, Purdue University. The first integration is performed over slowness, and the source signal corresponds to an explosive point source placed in the reflectivity zone itself (at a depth of 100 m). All multiple reflections and P-S conversions, as well free-surface effects within the chosen limits of integration are taken into account. In the synthetic seismogram calculations presented in this study, the horizontal phase-velocity interval used to fix the limits of integration ranged between 0.2-0.4 km/s below the minimum P-wave velocity in the model and 1.0-2.0 km/s above the maximum. The ratio of compressional to shear-wave velocities, $V_p/V_s$, was assumed to be equal to the square-root of 3 (Poisson's coefficient = 0.25) for all models.

In addition to the P- and S-velocities, the model is also determined by the density in each layer. In general, the calculations are performed on the basis of some empirical relationship between densities and compressional wave velocities. In the program version used here, the relation is:

$$\rho = 0.252 + 0.3788 \, V_p$$

Anelastic attenuation is described by the quality factor $Q$. It is defined by the relationship,

$$\frac{2\pi}{Q} = \frac{dE}{E}$$

where $dE$ is the amount of energy dissipated per cycle of harmonic motion in a given volume, and $E$ is the total elastic energy in that volume (Knopoff 1964). This means that a low $Q$-value corresponds to high attenuation. Values of $Q$ for the earth's crust and mantle have been determined by a number of authors using various methods (see e.g. Knopoff 1964, Clowes and Kanasewic 1970, Hill 1971, Braille 1977). Peterschmitt (1979) made a compilation of a large number of $Q$-value determinations, which demonstrates the enormous scatter of the results. Amplitudes of body waves, such as $P_g$
refracted in the upper-crustal basement, are affected only to a minor degree by variations in $Q$ (Braile 1977, Banda et al. 1982), so that, unless stated otherwise, a representative $Q_p$-value of 500 for P-waves was chosen for the crystalline part of the crust in the calculated models. Attenuation in sedimentary rocks being somewhat higher than in crystalline ones, a $Q_p$-value of 100 was chosen for the sediments. Following Kennett (1975b) shear-wave $Q$-values were set according to the relation $Q_s = 4Q_p/9$. 
## APPENDIX D. MODEL PARAMETERS

### TABLE D.1: Pg-models

<table>
<thead>
<tr>
<th>Model</th>
<th>Lower limit of layer thickness (km)</th>
<th>Layer thickness (km)</th>
<th>Vp (km/s)</th>
<th>Number of layers</th>
<th>Gradient (km/s/km)</th>
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TABLE D.4: Southern Alps models

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Geophysics is a relatively young science, which attempts to interpret some of the phenomena associated with the earth as a planet in terms of physical laws. Thus, some of the basic principles of physics, such as Newton's laws, the laws of thermodynamics, Maxwell's equations of electromagnetism and many more, have been applied with great success to deepening our understanding of the true nature of the earth. However, geophysicists have not been content with merely applying theoretical results developed by their glorious predecessors, but have established themselves with major contributions to physical theory, on their own right. The most significant piece of insight into the nature of physical reality thus arrived at is well-known under the name of Harvard Law:

Under the most rigorously controlled conditions of pressure, temperature, volume, humidity and other variables, the earth will do as it damn well pleases.

It is indeed typical for the modesty which characterizes earth scientists, that neither the exact date nor the author of this fundamental principle can be ascertained with confidence. In its presently known form, it was most likely formulated sometime after the early sixties of this century, at Harvard University. However, an earlier formulation of it can be found in the literature dating as far back as 1923. It is reproduced here as a small contribution to the reconstruction of the historical development of our science (e. e. cummings, 1923):

O sweet spontaneous
earth how often have
the
doting

fingers of
prurient philosophers pinched
and
poked
thee
,has the naughty thumb
of science prodded
thy

beauty . how
often have religions taken
thee upon their scraggy knees
squeezing and
buffeting thee that thou mightest conceive
gods
(but
true
to the incomparable
couch of death thy
rhythmic
lover

thou answerest
them only with

spring)
Nicholas Deichmann: Curriculum vitae

1949 Born in New York City, USA, as the son of Carl Deichmann and Daphne von Grunelius;

1955-1960 Primary school in Ascona, Switzerland;

1960-1964 Secondary school, Collegio Papio, Ascona, Switzerland;

1964-1967 High School, Pomfret School, Pomfret, Ct., USA;

1967 High School Diploma;

1967-1971 Student of philosophy, McGill University, Montréal, Canada;

1971 Bachelor of Arts;

1971-1972 Photography assistant, Atelier Groebli, Zuerich;

1972-1977 Student of geophysics, ETH-Zuerich;

1974 Field assistant with North Water Project, Canadian Arctic;

1977 Diploma in natural sciences;

1977- Research assistant with the Swiss Seismological Service;