RAY-BASED IMAGE RECONSTRUCTION IN CONTROLLED-SOURCE SEISMOLOGY WITH AN APPLICATION TO SEISMIC REFLECTION AND REFRACTION DATA IN THE CENTRAL SWISS ALPS

ABHANDLUNG

Zur Erlangung des Titels
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der
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TO
MY BELOVED PARENTS
AND
MY BROTHER KASPAR PHILIPP
HORATIO: O day and night, but this is wondrous strange!

HAMLET: And therefore as a stranger give it welcome.

There are more things in heaven and earth, Horatio,

than are dreamt of in your philosophy.

(Shakespeare, *Hamlet*, Act I, Scene 5)
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ABSTRACT

In this work I have tried to explore the potential of ray theoretical migration and forward modelling techniques for the joint - albeit not synoptic - interpretation of deep seismic reflection and refraction data. Simple analytical considerations show that for the velocities and travel times relevant for deep seismic reflection data migration displacements easily exceed 5 km vertically and 10 km laterally. This implies that virtually every deep seismic reflection profile needs to be migrated and that a minimum profile length of at least some 30 to 50 km is required to allow structural interpretation at greater depths. Estimates of the influence of uncertainties in velocity upon migration show that the error in the average velocity must not exceed 0.2 km/s at Moho depth in order to allow a meaningful comparison of the reflectivity imaged by normal-incidence profiling and the crustal velocity structure inferred from seismic wide-angle data. Conventional migration schemes based on the solution of the scalar wave equation rarely produce satisfying results when applied to deeper crustal data. An extensive review of the corresponding algorithms shows why these methods are highly sensitive to lateral velocity variations as well as to the short, laterally discontinuous reflection segments and high noise levels in conjunction with the high velocities and long travel times characteristic of deep seismic reflection data. These problems can be largely overcome by ray theoretical depth migration of digitised line drawings. As a case study the individual deep seismic reflection profiles of the eastern (ET) and southern (ST) traverses across the Swiss Alps, which were acquired as part of the National Science Foundation Program 20 (NFP20), have been combined along the course of the European Geotraverse (EGT). The resulting reflectivity distribution was simultaneously depth migrated with the contours of the smoothed, laterally consistent velocity field obtained by the reinterpretation of the seismic wide-angle profiles running parallel to the strike of the Alpine arc. This led to an overall excellent agreement between the most prominent reflectivity patterns and the strongest wide-angle reflections, which is considered to be an important criterion for successful migration. Assuming that the result of this migration represents an unbiased acoustic image of the present-day tectonic configuration of the crust below the central Alps low-angle subduction of the lower crust and uppermost mantle of the European plate below the Adriatic promontory of the African plate is clearly depicted. Orogenic crustal thickening is interpreted to arise from the stacking of nappes onto the European upper crust and from wedging of the European and Adriatic middle crusts. At least part of the south-vergent upper crustal thrusting in the Southern Alps can be accounted for by the inferred northward downbending of the Moho and lower crust of the Adriatic plate.
Hochfrequenzapproximation der Wellentheorie, stellt daher zur Zeit und wohl auch in absehbarer Zukunft für krustenseismische Reflexionsdaten die praktikabelste Lösung dar.

1. INTRODUCTION

Reflection seismology was developed in the 1920s and by the 1930s was being used on a large scale for oil exploration (cf. Sheriff and Geldart 1983; Robinson 1983). Various geometric transformation schemes to make the picked travel times of the observed reflections more accessible to geological interpretation have been in use since the early beginnings. These transformation techniques were labelled with the pictorial name migration as their purpose was to move or migrate the apparent positions of the reflections on the recorded seismic time sections to their true position in space. This terminology survived the advent of digital signal processing in the 1960s and early 1970s despite the fact that it became evident that the migration process represents a boundary/initial value problem to the scalar wave equation aiming at extrapolating the observed seismic wave field back through the imaged medium (cf. Claerbout 1970, 1971; Claerbout and Doherty 1972).

The purpose of this work is to illuminate the potential of ray theoretical migration and forward modelling methods for the joint - albeit not synoptic - interpretation of deep seismic reflection and refraction data. As a case study I have chosen the deep seismic reflection profiles across the eastern and southern Swiss Alps, which were shot as part of the Swiss National Science Foundation Program 20 "Geologische Tiefenstruktur der Schweiz" (NFP20) (cf. Schweizerische Arbeitsgruppe für Reflexionsseismik 1988; Frei et al. 1989). The major advantages of this dataset are the well-studied surface geology and tectonics, the possibility to extrapolate gross structures observed at the surface to depths of 10 km due to the eastward axial plunge of the Alpine structural grain (cf. Pfiffner et al. in press) and the unsurpassed wealth of complementary geophysical information, particularly seismic wide-angle profiles (cf. Mueller et al. 1980). The disadvantages of the NFP20 deep seismic reflection data are the commonly low signal to noise ratio (S/N) and their fragmentation into sometimes very short and laterally offset profiles.

As already indicated migration is kind of an unfortunate term not only because it fails to describe the underlying physics of the process but also because it is restricted to the repositioning part sensu strictu. Therefore I have decided to use the term image reconstruction instead of migration in the title. This is semantically more correct for high-frequency approximations of the wave equation and implicitly avoids the unnatural
separation between the critically interdependent processes of velocity estimation and repositioning of the imaged reflectivity.

In my opinion the fundamental difference between a PhD thesis and a corresponding publication in a scientific journal is that the former allows the author not only to document the results of his work but also their theoretical and methodological background. Ideally this makes PhD monographs suitable introductory texts for future generations of PhD or research students working in the same field. Therefore I have written this text in a tutorial and informal style, which hopefully may comfort those who feel somewhat bewildered by the rather formal title of this doctoral dissertation. As part of this policy I have also tried to achieve an acceptable balance between the internal coherency of the text as a whole and loose coupling between the individual chapters and sections in order to ease selective reading. Instead of lengthy phrasings this way of writing almost inevitably leads to the use of commonly accepted abbreviations and idioms, the most important ones of which are explained in Table 1.1.

The text is structured as follows:

• Chapter 2 reviews the most common ways to solve the scalar wave equation for the boundary/initial value problem of the migration of stacked seismic data. The relative merits of the various algorithms are worked out as well as the inadequacy of all these schemes to properly migrate deep seismic reflection data.

• Chapter 3 discusses ray theoretical aspects of migration and the possibilities of this approach to overcome some of the fundamental shortcomings of the previously presented migration schemes based on the wave equation. Using simple analytical considerations the general importance of migration of deep seismic reflection data as well as the necessity of an accurate parametrisation of the velocity field are demonstrated.

• Chapter 4 addresses general problems of velocity estimation on a crustal scale from spatially aliased seismic wide-angle data and the application of the thus derived velocity information to the parametrisation of the velocity field suitable for the migration of deep seismic reflection data. As a practical example a series of seismic wide-angle profiles running roughly parallel to the strike of the Alpine arc are reinterpreted in order to provide a smooth, laterally consistent velocity model across the central Alpine arc, which can subsequently be used for the migration of the eastern and southern segments of the NFP20 deep seismic reflection traverses across the central Alps (Frei et al. 1989).
Chapter 5 treats the migration of the NFP20 eastern (ET) and southern (ST) traverses as a case study. The individual seismic reflection profiles are combined and the reflectivity of the resulting profile is ray theoretically depth-migrated simultaneously with the contours of the velocity model previously derived from the reinterpreted wide-angle profiles. The tectonic implications of the resulting reflectivity distribution are discussed under the assumption that it represents a truthful acoustic snapshot of the present-day physical state of the crust below the central part of the Alpine arc.

Chapter 6 discusses fundamental aspects of the migration and interpretation of deep seismic reflection data and relates them to the considered Alpine dataset in order to assess validity of the approach taken and the thus resulting interpretation.

Chapter 7 is intended to be comprehensive list of the key references on the topics covered in this work.

The appendices contain user’s guides and descriptions of the code of the two major migration programs which I have either substantially modified or newly developed in the course of my PhD work.
Two-way travel time. Time it takes a seismic wave to travel from the source down to the target depth and back up to the receiver.

Common-mid-point. Mid-point between a particular source and receiver. Equivalent but nowadays less common expressions are CDP (common-depth-point) or CRP (common-reflection-point). Sorting data according to CMPs results in so-called CMP gathers. Reflections on CMP gathers exhibit a quasi-hyperbolic dependence of TWT versus spatial offset even if they arise from dipping reflectors.

Root-mean-square. The square root of the average of the squares of a series of measurements. In reflection seismology the rms velocity $v_{\text{rms}}$ is commonly used to "smooth out" ray bending effects in a vertically stratified velocity structure.

Normal moveout. Approximate description of the TWT versus offset dependence on CMP or shot gathers under the assumption of straight ray paths in a vertically stratified velocity field: $t^2(x) \approx t^2(0) + (x/v_{\text{rms}})^2$.

NMO-correction aims at removing the effect of NMO, which in the case of a CMP results in flattening the reflection "hyperbolae".

Summing all NMO-corrected traces of a CMP into one single central trace. Ideally this process collapses the offset domain completely and hence results in a so-called zero-offset section, i.e. a seismic section with coincident source and receiver locations.

Any kind of energy on a seismic section that does not arise from primary reflections or diffractions. Any non-coherent noise is commonly referred to as random noise. Truly random noise in a CMP gather containing $n$ traces is reduced by a factor of $n^{1/2}$ by stacking.

| S/N | Signal-to-noise ratio. |

Table 1.1. Frequently used abbreviations and idioms.
2. WAVE EQUATION MIGRATION

2.1 From Phenomenology to Theory

Most of the popular hand-migration techniques were based on Huygen's principle (cf. Hagedoorn 1954; Robinson 1983), i.e. the subsurface was considered to consist of point diffractors. So by summing along diffraction hyperbolae corresponding to the rms velocity of the overburden laterally coherent assemblages of point diffractors, i.e. reflectors, were moved to their true spatial position, individual diffraction hyperbolae were collapsed to their apexes and randomly distributed point diffractors were cancelled out. Although these manual techniques were painstaking to carry out they were very successful and had a clear and simple phenomenological basis. It was therefore only natural that with the commercial availability of digital computers in the early 1960s the first migration algorithms implemented by the oil industry were so-called "diffraction stack migrations" which - as indicated by their name - just emulated the previous manual techniques. However, though obviously the phenomenology of the migration process was well understood, the theory was not and these early diffraction stack migrations never worked satisfactorily (cf. Robinson 1983). In the early 1970s Jon Claerbout substantially contributed to the solution of the problem of computer migration by formulating the migration process as a boundary/initial value problem to the scalar wave equation (Claerbout 1970, 1971; Claerbout and Doherty 1972). The theoretical clarity and the practical success of this formal approach fruited in the development of a variety of migration algorithms in the following years all of which were explicitly based on the wave equation (e.g. Loewenthal et al. 1976; Claerbout 1976; Larner and Hatton 1990; Schneider 1978; Stolt 1978). Not unexpectedly it was found that the old diffraction stack migration only needed a little mathematical patching up to be a correct solution to the given boundary/initial value problem of the wave equation via the Kirchhoff integral and as such to do as good a job as any of the other methods (Larner and Hatton 1990; Schneider 1978). Common to all these early methods was that they produced an output in two-way travel time (TWT) and that they required a 1-dimensional or at least locally a laterally homogeneous definition of the velocity model. The first is very convenient for the interpreter and the second often is a sufficient approximation. However, little did anyone suspect in the beginning that in the case of strong lateral velocity heterogeneities
all these so-called time migration algorithms would break down and that a still deeper look into the theory was required to solve the problem (Gazdag 1980; Judson et al. 1980; Hatton et al. 1981). It was Hubral (1977) who - based on ray theoretical considerations - showed that in the presence of any lateral velocity variations time migration cannot depopagate, i.e. migrate, the observed wave field correctly.

The Exploding Reflector Model
Deep seismic reflection data still are almost exclusively migrated after stack and hence this chapter is restricted to stacked data. Assuming that NMO and stack have correctly collapsed the offset domain the unmigrated stacked section can be looked at as a zero-offset section, in which sources and receivers coincide and the energy travels the same way down to the reflectors and up again. According to Fermat's principle reflections at zero-offset always occur at a right angle to the reflector. Moreover, the zero-offset model assumes that no shear waves and no multiples are generated and consequently that the scalar or acoustic wave equation does model the data correctly.

Since the waves are assumed to travel the same way down and up this model can be further simplified by just considering the upgoing waves. The two-dimensional zero-offset wave field $p(x,z=0,t)$ which we record at the surface can then be imagined to have been generated by a reflector "exploding" at time $t=0$ (Loewenthal et al. 1976). For this purpose one can either half the observed two-way travel time or half the velocity of the medium through which the wave has travelled. The latter approach is more common and will be pursued in the following.

The scalar wave equation relates the acoustic pressure field $p(x,z,t)$ and the compressional wave velocity $v(x,z)$ as follows:

$$\frac{\partial^2 p(x,z,t)}{\partial t^2} = \frac{v^2(x,z)}{4} \nabla^2 p(x,z,t) = \frac{v^2(x,z)}{4} \left\{ \frac{\partial^2 p(x,z,t)}{\partial x^2} + \frac{\partial^2 p(x,z,t)}{\partial z^2} \right\}$$

(2.1)

The factor four on the right hand side of (2.1) arises from the transformation of $v(x,z)$ to $v(x,z)/2$ required by the exploding reflector model. Given the seismic zero-offset (i.e. CMP-stacked) section observed at the earth's surface $p(x,z=0,t)$ as the boundary/initial condition we want to solve the scalar wave equation (2.1) for the reflectivity $p(x,z,t=0)$ of the subsurface. This solution corresponds to the migrated seismic section. The reverse procedure is called forward modelling. Migration and
forward modelling are therefore very closely related; whilst migration *depropagates* a wave field observed at the surface in order to unravel the reflectivity of the subsurface, modelling provides an observed wave field by *propagating* it through the reflectivity (Figure 2.1).

![Figure 2.1. The relation between migration and forward modelling of zero-offset seismic data.](image)

In the following several basic methods to solve this boundary/initial value problem are reviewed and their relative merits and drawbacks are discussed. From this discussion it will become evident quite naturally that - even under the best possible circumstances - wave equation migration of deep crustal data is problematic. All the problems are formulated in two dimensions; this has two reasons: first, crustal reflection data are and will be virtually exclusively acquired along profiles (i.e. in two dimensions); second, an extension to the third dimension is in the considered cases straightforward to formulate but painstaking to type. Moreover, there arise formidable conceptual problems in three dimensions, such as the parametrisation of the velocity field, which are well beyond the scope of this text.

### 2.2 Time Migration with Transform Methods

**Phase Shift Migration (Gazdag 1978)**

If the velocity is kept constant ($v=\text{constant}$) it is amazing how simple things become. The scalar wave equation has then, as is well known, a set of harmonic plane wave solutions of the form
exp\{i(\omega t-k_x x-k_z z)\} \quad (2.2)

constrained by the so-called "dispersion relation"

\[ k_x^2 + k_z^2 = \frac{4\omega^2}{v^2} \quad (2.3) \]

Where \( k_x \) and \( k_z \) are the wave numbers with respect to \( x \) and \( z \) (positive downward) and \( \omega \) is the angular frequency. In the following I have adopted the common practice of seismic data processing and not explicitly indicated the integration boundaries whenever they are constrained by the considered data.

Taking the 2-dimensional Fourier transform of \( p(x,z,t) \) over \( x \) and \( t \) yields

\[
P(k_x,z,\omega) = \int \int p(x,z,t) \exp\{i(\omega t-k_x x-k_z z)\} \, dx \, dt \quad (2.4)
\]

Considering just upgoing waves (2.4) can be rewritten according to (2.2) and (2.3) as

\[
P(k_x,z,\omega) = P(k_x,z=0,\omega) \exp\{ik_z z\} \quad (2.5)
\]

To obtain again \( p(x,z,t) \) just the 2-dimensional inverse Fourier transform has to be taken

\[
p(x,z,t) = \frac{1}{(2\pi)^2} \int \int P(k_x,z=0,\omega) \exp\{i(k_x x+k_z z-\omega t)\} \, dk_x \, d\omega \quad (2.6)
\]

For \( t=0 \) (2.6) evidently becomes a migration equation, which is commonly known as phase shift migration (Gazdag 1978)

\[
p(x,z,t=0) = \frac{1}{(2\pi)^2} \int \int P(k_x,z=0,\omega) \exp\{i(k_x x+k_z z)\} \, dk_x \, d\omega \quad (2.7)
\]

Although the vertical coordinate in (2.7) is formulated as depth \( z \) it is - due to the constant velocity - nothing but scaled travel time and the final output is generally computed as such.
Formula (2.7) is strictly valid only for a constant velocity of the medium. However, for a vertically stratified medium \( v = v(z) \) a generalized approximation is possible (Gazdag 1978)

\[
P(k_x,z,\omega) = P(k_x,z=0,\omega)\exp\{i \int_0^z k_z(z')dz'\} \tag{2.8}
\]

Thus (2.7) becomes

\[
p(x,z,t=0) = \frac{1}{(2\pi)^2} \iint \exp\{i(k_x x' - k_y y' - k_z z')\} p(x',z=0,t')dx'dk_x d\omega \tag{2.9}
\]

where

\[
\gamma = \frac{k_y v}{2} = \omega \sqrt{1 - \frac{k_x^2 v^2}{4\omega^2}} \tag{2.10}
\]

The necessity of a detailed knowledge and specification of the interval velocities \( v(z) \) of the medium can be circumvented by expanding the square root

\[
\sqrt{1 - \frac{k_x^2 v^2}{4\omega^2}} \approx 1 - \frac{k_x^2 v^2}{8\omega^2} \tag{2.11}
\]

so that the so-called phase shift term becomes

\[
\int_0^z \gamma dt' = \omega t - \frac{k_x^2 v^2}{8\omega^2} \nu_{\text{rms}} t \tag{2.12}
\]

which now depends only on the rms velocity. In ray theory this approximation corresponds to a straight ray path approximation in a vertically stratified medium whose inaccuracy increases with increasing deviation of the ray from the vertical, i.e. with increasing dip of the reflector.
Practice has shown that in the case of a constant velocity medium, both accuracy and computational efficiency of formula (2.7) can be enhanced by a change of integration variable from $\omega$ to $k_z$. From the dispersion relation (2.3) we get for upgoing waves

$$\omega = -\frac{k_z v}{2} \sqrt{1 + \frac{k_x^2}{k_z^2}}$$

and thus

$$\frac{d\omega}{dk_z} = -\frac{k_z v^2}{4\omega}$$

With these two relations the constant velocity phase shift migration (2.7) becomes

$$p(x,z,t=0) = r(x,z) = -\frac{v^2}{4(2\pi)^2} \int P(k_x,z=0,\omega) \exp\{i(k_x x + k_z z)\} \frac{(k_z/\omega)}{dk_x \, dk_z}$$

This equation is commonly known as frequency-wave number or f-k migration (Stolt 1978) because the same formula results by applying three Fourier transforms (with respect to $t, x$ and $z$) to the wave equation, which transforms it into an algebraic equation, solving the boundary/initial value problem in the resulting frequency-wave number (f-k) domain and finally transforming the solution back into the time-space domain.

The major difference between the phase shift and the f-k migration is that in the latter the depth domain $z$ is Fourier-transformed as well. The Fourier transform, like all other integral transforms, is a global operator which by definition limits the possibility to adjust to local conditions. As a consequence no velocity variations, not even 1-dimensional ones as in the case of the phase shift migration (2.12) can be explicitly incorporated in the f-k migration formula.

There are, however, several implicit ways to accommodate velocity variations into the f-k migration:

- Migrate the entire section with different velocities and get the final migrated section by "patching together" the correctly migrated parts.
• Stretch the unmigrated time domain according to the desired 1-dimensional velocity function \( v(z) \). This is commonly referred to as f-k time stretch (Stolt 1978).

2.3 Kirchhoff Time Migration

As already mentioned above this migration technique simply represents a mathematically refined version of the early diffraction stack migration. The Kirchhoff integral theorem expresses the value of the wave field at an arbitrary point in terms of the value of the wave field and its normal derivatives on an arbitrary closed surface. In reflection seismology measurements are made only at the surface and therefore, instead of a closed surface, the integration must be limited to the surface of the earth. Furthermore, in seismic practice, only the wave field is recorded and its normal derivative is not available. In terms of the Kirchhoff integral the migrated wave field at some point \((x',z')\) in the subsurface is obtained as follows (cf. Larner and Hatton 1990; Schneider 1978):

\[
p(x',z',t=0) = \int \frac{\cos \theta}{\sqrt{\pi r v}} \frac{\partial^{1/2}}{\partial t} p(x,z=0,t=2\pi/v) dx
\]

(2.16)

\( v \) is assumed to be constant and thus is the straight line connecting the observation point \((x,z=0)\) at the surface and the subsurface "source" point \((x',z')\), \( \theta \) the angle between the z-axis and the line joining \((x,z=0)\) and \((x',z')\) and \( \frac{\partial^{1/2}}{\partial t} \) the half derivative with respect to the travel time \( t \). A simple expression for the operator \( \frac{\partial^{1/2}}{\partial t} \) can be given in the frequency domain where its transfer function is given by \( \sqrt{i\omega} = \sqrt{\omega} \exp(i\pi/4) \), which corresponds to a 45 degree phase shift.

Hence the Kirchhoff migration operator has to major parts:

1) a 45 degree phase shift

2) a weighted sum of all the points along each diffraction hyperbola into its apex

These two features had been "forgotten" in the early diffraction stack implementations and resulted in their unsatisfactory performance (Robinson 1983).
Formula (2.16) can be either derived directly from the Kirchhoff integral (Larner and Hatton 1990) or by rearranging terms and integrals and a stationary phase approximation of the f-k migration formula (2.15) (Stolt and Benson 1986).

Kirchhoff migration is physically very satisfying. It shows that - in agreement with Huygen's principle - any reflection can be considered as a laterally coherent superposition of point scatterers. Each point scatterer will appear in the unmigrated data as an event with hyperbolic moveout and will have the 45 degree phase shift which we associate with diffractions. Each such hyperbolic event is summed back into its apex, correcting phase and amplitude for each term in the sum.

When the velocity of the medium varies with depth only the constant velocity \( v \) in (2.16) may be approximated by the corresponding rms velocity \( v_{\text{rms}} \), which corresponds to a straight ray path approximation in a vertically stratified medium. If the medium velocity does not only vary vertically but also laterally, the summation curves cannot be approximated by analytical diffraction hyperbolae. Formula (2.16) must then be generalized for \( r \) and the corresponding summation curves and the wave paths must be computed separately. As outlined in the following chapter this is commonly done by ray tracing.

2.4 Finite Difference Time Migration

The previously introduced transform and integral migration schemes enjoy some clear advantages in terms of simplicity and understandability. However, these schemes are global operations in the sense that each output is computed from the entire dataset, which makes adjustments to local velocity variations difficult or impossible.

Partial differential equations as such describe basic physical principles at an infinitesimally small scale and hence local solutions are most naturally obtained by approximating them by the corresponding finite difference equations. It was in fact this kind of an approach which led to the breakthrough of the wave equation migration method (Claerbout 1970, 1971; Claerbout and Doherty 1972; Loewenthal et al. 1976).

At first sight it seems to be straightforward to formulate a suitable finite difference approximation to the scalar wave equation and then - explicitly or implicitly - solve it by stepwise downward continuation in time or depth. However, the boundary/initial value
problem is ill-posed for such a brute force approach because the second vertical derivative of the wave field recorded at the surface is not known. Claerbout (1970, 1971) elegantly circumvented this problem by the following coordinate transformation

\[
x' = x \\
z' = z \\
t' = t + \frac{2z}{v_c}
\]

(2.17)

\[p(x,z,t) = q(x',z',t')\]

Here \(v_c\) is a constant reference velocity.

For a medium of constant velocity \(v_c\) the scalar wave equation (2.1) thus becomes

\[
\frac{\partial^2 q}{\partial z' \partial t'} = -\frac{v_c}{4} \left\{ \frac{\partial^2 q}{\partial x'^2} + \frac{\partial^2 q}{\partial z'^2} \right\}
\]

(2.18)

The new coordinate system defined by (2.17) travels with the upcoming wave front from the exploding reflector at depth to the receivers at the surface, which allows a much coarser discretisation in time and depth and thus a cruder approximation of the second vertical derivative. The new variable \(t'\) is commonly referred to as "advanced travel time" (Stolt and Benson 1986) and is a measure for the deviation of the wave from the vertical. So for nearly vertically travelling waves \(q(x',z',t')\) will change slowly with respect to \(t'\) and \(z'\) and hence (2.18) can be approximated as

\[
\frac{\partial^2 q}{\partial z' \partial t'} = -\frac{v_c}{4} \frac{\partial^2 q}{\partial x'^2}
\]

(2.19)

Comparing the dispersion relation of (2.19) with that of (2.1) shows that accurate results (error <0.2%) will be obtained for deviations from the vertical of less than 15 degrees, and hence (2.19) is generally referred to as the "15 degree equation" (Claerbout 1976).

(2.19) is a constant velocity approximation to the wave equation and thus there is little sense in looking for a finite difference solution to it. So the obvious question is what happens if we allow the velocity in (2.17) to vary both vertically and laterally. The answer is: very little! Neglecting velocity gradients the 15 degree approximation of the scalar wave equation (2.19) just becomes
Claerbout (1976) has shown that (2.20) can be extended to handle dips up to 45 and even 55 degrees properly, and so one might suspect that the corresponding finite difference solution would allow to depropagate the observed wavefield correctly through media of random velocity. As outlined below this is not the case.

\[
\frac{\partial^2 q}{\partial z' \partial v'} = -\frac{v(x,z)}{4} \frac{\partial^2 q}{\partial x'^2} \tag{2.20}
\]

2.5 Depth Migration

Hubral (1977) has shown that no time migration scheme could depropagate the wavefield correctly in the presence of lateral velocity variations. This is easy to reconcile in the case of transform and integral methods which simulate lateral velocity variations as a series of 1-dimensional velocity functions. Such a model parametrisation approximates a dipping layer in a staircase sense and hence Snell's law is bound to be violated because the waves are refracted either at horizontal or vertical boundaries but never at the true slope of the interface.

For the finite difference approach (2.20) and its extensions it is much less clear what went wrong because the velocity obviously is allowed to vary laterally. The reason is that we have tricked ourselves by neglecting the second vertical derivative of the wavefield and of the advanced travel time (cf. (2.18)), which describes the deviation of the upward travelling wavefield from the vertical. In doing so we implicitly approximate the medium as being locally laterally homogeneous which results in a violation of Snell's law in the same sense as in the case of the transform and integral migration methods.

So for a correct depropagation of the wave field through a heterogeneous medium there is no other way but to solve the full scalar wave equation. As already mentioned the major difficulty arises from the second vertical derivative of the wave field which is not contained in the boundary/initial conditions provided by the observed wave field. Feasible approaches have been presented e.g. by Gazdag (1980), Judson et al. (1980), Hatton et al. (1981) and Kosloff and Kessler (1987).
2.6 Migration Velocity

From the fact that migration represents a depropagation of an observed wave field it is already obvious that the migration velocities, very much unlike stacking velocities, do correspond to the true medium velocities. Although this is in principle true for both time and depth migration schemes, time migration schemes - due to the essentially 1-dimensional definition of the velocity field - are much less sensitive to uncertainties in the specification of the velocity model than depth migration schemes in the sense that considerably different velocities will lead to only minor differences in the time migrated position of a particular reflector. This treacherous robustness of time migration algorithms with respect to uncertainties in velocity is largely responsible for their popularity and led Hatton et al. (1986) to the sarcastic remark: "Migration velocities tend to be treated in a cavalier fashion to say the least. For many years, the standard doctrine has been to use the stacking velocity, give or take an apparently random few per cent to account for the day of the week, the weather, or some other whim."

The effects of uncertainties in the parametrisation of the velocity model upon depth migration will be examined quantitatively in the next chapter using ray theoretical considerations.

*Wave Equation Migration and Spatial Resolution*

Due to the geometric spreading of the wave front the spatial resolution of unmigrated seismic data decreases with increasing travel time and velocity. Spatial resolution may be quantified by the diameter of the so-called 1st Fresnel zone, which corresponds to the minimum lateral extent a reflector must have so that for a given dominant wave length $\lambda$ all incident energy interferes constructively. In other words:

- For reflectors smaller than the corresponding Fresnel zone the amount of diffracted incident energy increases with decreasing reflector dimension.
- Individual reflectors laterally separated by less than one Fresnel zone diameter cannot be distinguished.
Assuming that maximum constructive interference occurs at a phase shift of one quarter of the dominant wave length $\lambda$ the diameter of the 1st Fresnel zone $D_1$ may be readily defined for a spherical wave front as a function of the average velocity $v$ of the overburden, the dominant frequency $\nu$ and two-way travel time $t$ (e.g. Sheriff 1980)

$$D_1 = v \sqrt{\frac{t}{\nu}} \tag{2.21}$$

Let us consider the NFP20 eastern traverse (see Chapter 5) as an example: assuming a dominant frequency of 25 Hz (Valasek et al. 1990) and an average crustal velocity of 6.0 km/s we get 4.2 km for the diameter of the 1st Fresnel zone at Moho depth (12 s TWT) at the northern end of the traverse.

As seen before, migration is a process that depropagates or downward continues the recorded wave field and optimally this procedure not only moves dipping events into their correct spatial position but also increases the spatial resolution of the seismic data by removing the effect of geometric spreading, i.e. by collapsing the Fresnel zone. Ideally therefore the spatial resolution of migrated seismic data depends only on the dominant wave length $\lambda$ and and hence a migrated seismic section may be considered to represent the convolution of the seismic source signal with the reflectivity of the imaged subsurface. As a consequence migration is useful even if no dipping events are present in the unmigrated data (cf. Berkhout and Van Wulfften Palthe 1979; Berkhout 1984; Stolt and Benson 1986; Lindsey 1989).

**Parametrisation of the Velocity Model**

Another pertinent question concerns the scale, both lateral and vertical, at which the migration velocity model should be defined. Standard and generally successful practice of the oil industry is to use the interval velocities between major reflectors (e.g. Stolt and Benson 1986; Berkhout 1987; Keho and Beydoun 1988) though we know from well information that significant velocity variations occur at a much smaller scale. This may be explained as follows:

In terms of linear system theory the recorded seismic signal represents the convolution of the source signal with the Green's function of the medium. The latter may be split up into a "macro" part representing the large-scale velocity variations and a "micro" part representing the small-scale velocity variations, i.e. the reflectivity of the

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*References*

- Berkhout and Van Wulfften Palthe 1979
- Berkhout 1984
- Stolt and Benson 1986
- Lindsey 1989
subsurface. The desired output is - as outlined above - the convolution of the input signal with the reflectivity of the medium, and hence the effects of the large-scale velocity structure must be removed from the observed signal. From this it is evident that the migration velocity model is correctly described by the large-scale features, i.e. the macro model of the velocity structure. As to the scale of the macro model there are no firm rules. Successful practice of the oil industry is to stick to those reflecting horizons separated by several dominant wave lengths which are believed to be associated with the most significant velocity changes.

Migration as a Velocity Analysis Tool

There are two basic approaches. The first one involves migration with a range of constant velocities and measuring the amount of focused energy either numerically or by eye (e.g. Sattlegger 1975; Yilmaz and Chambers 1984; Stolt and Benson 1986). This approach, which in principle is very similar to NMO velocity analysis by constant velocity Stacks, is useful provided that the lateral velocity variations are mild and time migration is adequate.

In the case of severe lateral velocity variations Stolt and Benson (1986) suggest a somewhat more sophisticated technique which they descriptively call "self-consistent velocity analysis". This interpretive depth migration technique, which is sketched in Figure 2.2, again illustrates that forward modelling (wave field propagation) and migration (wave field depropagation) should not be regarded as two separate procedures but rather as one single iterative process.
Digitise the key horizons from the unmigrated time section and convert them to depth using best-guess interval velocities

Modify this model by trial-and-error forward normal-incidence ray tracing until the resulting synthetic sections resembles the corresponding unmigrated data

Use this model as a starting model for depth migration

Severe disagreement?

Modify the geometry and/or the interval velocities of the migration velocity model

Is the agreement between the depth-migrated key-horizons and the used velocity model satisfying?

Final depth-migrated section and interval velocity model

2.7 Migration of Deep Seismic Reflection Data

All the migration schemes presented in this chapter are based on the one-way scalar wave equation and therefore cannot correctly handle P-S conversions, multiples, incompletely sampled parts of the wave field and (coherent and random) noise.

Due to the steep incidence angles one may expect that little or no P-S conversions do occur. Multiples and coherent noise can, in general, be suppressed to the level of the random noise in the processing stage. If this is not possible they will have a destructive, albeit unpredictable, effect on the result of the migration.
Wave Equation Migration and Incomplete Sampling of the Wave Field

Deep seismic reflection elements are commonly laterally discontinuous and often shorter than the 1st Fresnel zone. If such features were representative of the full wave field they should be associated with diffractions. This is not the case and Warner (1987) as well as Raynaud (1988a) argue that the lateral discontinuities of deep seismic reflections represent "data shadows" which arise from overburden complexities particularly near the surface. This argument is supported by Wever (1988) who statistically showed that the average length of deep reflectors decreases with increasing static complications. Wave equation migration overcompensates the lacking diffractions in the shadow zones which results in so-called "smiles" at the reflector ends. In the case of short reflectors the "smiles" become dominant and the migrated section thus gets uninterpretable. Warner (1987) has convincingly demonstrated that this effect is negligible for travel times and velocities characteristic of near-surface data but becomes disastrous at lower crustal levels.

In order to reduce the smiles and make the section look "properly" migrated it is not uncommon to use significantly too low (up to 30%) migration velocities. However, this can by no means be considered as a true remedy because - as demonstrated in the following chapter - the resulting positioning errors rather decrease than increase the structural interpretability of the migrated section compared to the unmigrated section.

Wave Equation Migration and Random Noise

In reflection seismology the term random or Gaussian noise is used for any unwanted incoherent parts of the recorded seismic signal. Compared to industry data the level of random noise is uniformly high in deep seismic reflection data. It is therefore interesting to examine the effect of random noise upon migration in more detail. This is most easily done on the example of the phenomenologically clear Kirchhoff summation approach which was introduced in Section 3 of this chapter. The consequences will, however, be the same for all migration schemes based on the wave equation.

The essential contribution to each Kirchhoff summation arises from the region of stationary phase, which increases with travel time $t$ proportional to $(v_{rms}^2 t)^{1/2}$ (Hatton et al. 1986). The amplitude variation of previously uniform Gaussian noise will therefore be proportional to $1/(v_{rms}^2 t)^{1/2}$ after migration (cf. Hatton et al. 1986). Hence, against common prejudices migration does reduce and not increase the content of random noise.
There is, however, also a major drawback. For long travel times and/or high velocities the summation hyperbolae become nearly flat and thus migration essentially results in a spatial mixing of neighbouring traces. Spatial mixing does, as is well known, enhance the lower wave numbers and thus gives the data a smeared and "wormy" look which drastically reduces their resolution and interpretability. For signal-to-noise ratios of 1 or less, which are common for deep seismic reflection data, the contribution of noise and thus the smearing will become dominant. Wave equation migration of noisy, high-velocity seismic data from deep reflectors will therefore always be problematic and even harsh noise reduction procedures either before (Valasek et al. 1990) or during migration (Milkereit 1987) can only partially alleviate the problem.

**Migration of Deep Seismic Reflection Data**

In this chapter I have tried to illustrate the fundamental migration problems for deep seismic reflection data by a rather extensive treatment of the most common algorithms based on the scalar wave equation. It turns out that the more accurately a migration algorithm models the actual wave propagation the more sensitive it is to insufficiencies of the input data and the velocity model: depth migration is much less "robust" than time migration and the accurate algorithms based on integral or transform schemes are more delicate than approximate finite difference solutions (cf. Hatton et al. 1986).

The characteristic properties of deep seismic data such as short discontinuous reflection segments, the combination of long travel times and high velocities and generally high noise levels make it unlikely that any scheme directly based on the scalar wave equation will ever work universally. One potential - and so far probably the only universally applicable - solution is to use ray theory to migrate a line drawing of the primary reflections and diffractions. As a crude first-order approximation to the wave equation ray theory does not suffer from an incomplete sampling of the wave field and noise problems are avoided by the line drawing preparation process, which ideally results in a noise-free version of the data with all the amplitude and phase information neglected. Moreover, ray theory can handle strong lateral velocity variations and thus naturally produces a migrated depth section as output. Drawbacks are the inherent subjectivity of the line drawing preparation and the fact that ray theory does assume perfect resolution and hence does not collapse the Fresnel zone during migration.
The remaining part of this work is dedicated to the methodological background and to an application of ray-theoretical depth migration to digitised line drawings of deep seismic reflection data from a structurally complicated area.
3. RAY-THEORETICAL MIGRATION

3.1 Outline of Ray Theory

The governing equations and the geometric concepts of ray theory have been known for a long time. As a high-frequency approximation of the wave equation it fails to correctly predict any frequency-dependent features such as diffractions and interference effects, which are characteristic of a wave field containing finite frequencies. Nevertheless ray theory has been successfully used in seismology for at least one hundred years and many important concepts, such as the CMP-stacking and migration in reflection seismology, were and still are based on ray-geometric concepts. The advent of digital computers and thus the possibility to solve the governing equations numerically for media of "arbitrary" complexity has led to a stunning renaissance of the ray method in seismology (e.g. Cerveny et al. 1977).

Ray Series - Asymptotic Ray Theory (ART)

Babich and Alekseyev (1958) in the USSR and Karal and Keller (1959) in the USA independently developed a series solution of the elastodynamic equation for inhomogeneous isotropic media

\[
\frac{\partial^2 u_i}{\partial t^2} = (\lambda+\mu) \nabla(\nabla \cdot u_i) + \mu \nabla^2 u_i + \nabla \lambda (\nabla \cdot u_i) + \nabla \mu \times (\nabla \times u_i) + 2(\nabla \mu \cdot \nabla) u_i
\]  

(3.1)

where \( u_i \) is the displacement vector (\( i=1,2,3 \)), \( \rho \) the density, \( \lambda \) and \( \mu \) Lamé's parameters and "\( \times \)" the vector product.

This solution is commonly referred to as ray series approach and in the frequency domain has the following form (cf. Cerveny et al. 1977):

\[
ui = \exp\{ -i\omega(t-\tau(x_i)) \} \sum_{n=0}^{\infty} U_i^{(n)} (-i\omega)^n
\]  

(3.2)

Here \( \omega \) denotes the angular frequency, \( t \) the travel time, \( U_i^{(n)} \) the complex-valued amplitude coefficients and \( \tau(x_i) \) the real valued phase function or eikonal.
The full ray series (3.2) is asymptotic, i.e. for $n = \infty$ represents the full wave field, and hence is commonly referred to as asymptotic ray theory (ART) solution of the elastodynamic equation. The leading term of (3.2)

$$u_i = U_i \exp \{-i\omega(t-\tau(x_i))\}$$

(3.3)

corresponds to the well-known ray-theoretical approach of geometric optics and represents the ART solution for $\omega = \infty$. It therefore is also referred to as zero-order ART (e.g. McMechan and Mooney 1980).

Whilst the leading term (3.3) of the ray series (3.2) is strictly only valid for infinitely high frequencies, higher-order terms of (3.2) do allow to accommodate frequency-dependent effects and in seismology can be used to model "higher-order waves" such as head waves (Cerveny and Ravindra 1971). However, the practical determination of the coefficients $U_i^{(n)}$ is difficult and hence most seismological approaches are restricted to the leading term of (3.2).

Surfaces of constant phase $\tau = \tau(x_i) = \text{constant}$ represent wave fronts which are diagnostic of the travel time behaviour of the propagating wave field. Inserting (3.3) into the elastodynamic equation (3.1) leads to an eigenvalue problem the solution of which is the so-called "eikonal equation" which has the same form for P- and S-waves and is a key to the study of the travel times of seismic waves (cf. Vidale 1988, 1990):

$$\left(\frac{\partial \tau(x_i)}{\partial x_i}\right)^2 = (\nabla \tau(x_i))^2 = \left(\frac{1}{v}\right)^2$$

(3.4)

where $v$ is the velocity of P- and S-waves, respectively.

One method to solve first-order non-linear partial differential equations such as (3.4) is to integrate them along their so-called characteristics (see e.g. Courant and Hilbert 1966).

It can be shown that the characteristics of the eikonal equations are the extremals of Fermat's functional (e.g. Cerveny et al. 1977)

$$\int_{s}^{d} \frac{ds}{v(s)} = \text{EXTREMAL}$$

(3.5)
where $S$ represents the total length of the ray path.

These extremals or characteristics are called rays. They form a system of trajectories perpendicular to the eikonal surfaces defined by $\tau = \tau(x_j) = \text{constant}$ and are termed "ray equations"

\[
\frac{dx_i}{d\tau} = v^2(x_i) \, p_i \\
\frac{dp_i}{d\tau} = -\frac{1}{v(x_i)} \frac{dv(x_i)}{dx_i}
\]  

(3.6)

$p_i$ is the so-called slowness vector which is perpendicular to the wave front, i.e. parallel to the ray, and has the length $v^{-1}$, $\tau$ denotes the travel time along the ray measured from a given reference point.

It should be noted that there are numerous alternative ways to derive the eikonal and ray equations. The possibly most straightforward approach is to look for harmonic plane-wave solutions of the scalar wave equation for the asymptotic case of infinitely high frequencies. Following Cerveny et al. (1977) I have chosen the so-called ray series approach because it nicely relates the zero-order ART, i.e. geometric ray theory, to higher-order, finite-frequency solutions.

**Solution of the Ray Equations**

In three dimensions (3.6) represents a system of six ordinary linear differential equations of the first order. Analytical solutions exist for a few rather special definitions of the velocity field (cf. Gebrande 1976; Camina and Janacek 1984; Cerveny 1985) but also the numerical integration of these equations is unproblematic even in complex media (Cerveny et al. 1977). System (3.6) describes the geometry of rays and the associated travel times and thus may be useful to study the travel time behaviour of seismic waves. As long as the velocity field and its first spatial derivatives are continuous a particular ray is uniquely defined by its starting point and its initial slowness vector and we can - in a stepwise fashion - calculate the coordinates $x_i$ and travel times $\tau(x_j)$ along this ray.
Three special cases of velocity structures deserve special attention:

- \( \frac{dv(x_j)}{dx_i} = 0 \), i.e. \( v=\text{constant} \).

In this case the right-hand side of (3.6) vanishes and the integration of the left-hand side shows that the ray path is straight.

- \( \frac{dv(x_i)}{dx_i} = \text{constant} \).

In this case the ray describes a circle (cf. Camina and Janacek 1984). This may be exploited to avoid the numerical solution of (3.6) (e.g. Gebrande 1976; Schrader 1990), however, at the price of a rather special and in some aspects limited definition of the velocity field.

- One or several components of \( \frac{dv(x_j)}{dx_i} \) are infinite.

This means that the ray has hit an interface and the new initial values for the slowness vector \( p_j \) are given by Snell's law.

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**Limitations of Ray Theory in Seismology**

The ray equation system (3.6) is an asymptotic approximation of the wave equation for infinitely high frequencies. It therefore predicts perfect resolution and excludes any frequency-dependent phenomena such as diffraction and interference effects. Moreover, ray theory is not able to describe the amplitude behaviour of waves in the critical region, at caustics and in shadow zones (Cerveny et al. 1977). Clearly rays, i.e. infinitely high frequencies, lack physical existence and are nothing but a mathematical concept. Nevertheless, this concept has been successfully applied to seismic waves which range from less than 1 Hz to more than 50 Hz and to model scales ranging from several thousands of kilometres to a few hundred metres. There is no firm metrics determining the range of validity of the ray approximation in seismology but bearing in mind the inherent assumptions the theory is general leads to meaningful results.

- To avoid diffraction effects the minimum radius of the curvature of an interface must always be larger than the diameter of the 1st Fresnel zone (cf. formula (2.21)) of the modelled wave field at that depth (Figure 3.1a).

- To avoid interference effects all interfaces must be separated by at least one dominant apparent wave length of the modelled wave field.
• Velocity gradients must not be too large. This is commonly the case if \( \frac{v}{(|v_x| + |v_z|)} \gg \Lambda \), where \( \Lambda \) is the dominant wave length.

• Problems with caustics (Figure 3.1b) and shadow zones (Figure 3.1c) can be reduced by a careful design of the velocity model. However, the insufficiencies of the ray method at caustics, in shadow zones and in the critical region are severe only if amplitude considerations are important.

Under the above restrictions ray theory may be expected to simulate wave propagation accurately enough to allow even a qualitative consideration of ray-theoretically calculated amplitudes and signal shapes (Cerveny et al. 1977). Raynaud (1988a) has shown that even under considerably less restrictive conditions ray theory correctly predicts the travel time behaviour of the wave field. By comparative wave and ray-theoretical modelling of deep seismic reflection data he came to the conclusion that complexities in the overburden normally smaller than one dominant wave length were mainly responsible for the typically discontinuous nature of deep reflections. Although ray theory clearly failed to simulate the amplitude behaviour of the wave field it correctly reproduced the travel times to the major reflections and diffraction apaxes. This finding serves as a heuristic justification of a ray-theoretical depth migration of digitised seismic reflection data through complex media (Raynaud 1988a,b).

Finally one has to be aware of the fact that system (3.6) just represents the leading term of the ray series (3.2) and thus explicitly excludes higher-order waves such as head waves or multiples. To model these effects with (3.6) suitable new initial conditions have to be specified at the corresponding positions of the ray path. Even if a ray just hits a boundary we have to explicitly specify the new initial conditions for the reflected and/or refracted ray. In other words: Zero-order ART just does what you tell it to do! At first sight this may seem to be a nuisance and for certain applications it certainly is. For ray-theoretical migration, however, it turns out to be a blessing: Whilst incomplete sampling of the wave field jeopardizes wave equation migration, ray migration is not affected. Specify what you consider to be part of the primary wave field and it will migrate that - and only that!
(a) Wavefront and Wavelet

(b) Caustic and Focus

V1 > V2
Figure 3.1. (a) Compared to the dominant wave length of the incident signal and the width of the corresponding Fresnel zone the sketched interface is "rough". This implies that substantial amounts of the incident seismic energy are diffracted and hence that ray theory is inadequate in this situation. Sketches of a caustic (b) and a shadow zone (c). Dashed lines depict rays, arrows their direction. The wavelets at the top and the bottom of the figure qualitatively show the expected amplitude behaviour, which is not correctly reproduced by ray theory.

3.2 Ray Migration Concepts in a Constant Velocity Medium

Many wave phenomena can be best visualized using ray geometry. Migration is no exception. As we have seen in the previous section ray geometry becomes very simple for constant velocity since then the ray paths are straight. Figure 3.2 shows the relation between the unmigrated position A and the migrated position B of a dipping reflector on a stacked section. The velocity is assumed to be constant and hence the depth \( z \) and the TWT \( t \) only differ by a constant factor and the depth \( z \) may simply be considered as
scaled travel time $t$. The offset domain is assumed to have been collapsed perfectly by correction for NMO and stacking and hence the section can be considered to zero-offset, i.e. sources and receivers are coincident. Due to the plane-layer assumption inherent in the CMP-concept a dipping reflector $A$ is plotted vertically below the coincident source/receiver pair $R$. Elementary ray geometric considerations, however, make it evident that zero-offset can only be achieved if the ray hits the reflector at a right angle and therefore the reflector has to be moved updip along the wave front from its unmigrated position $A$ to its true spatial, i.e. migrated, position $B$. This allows us to establish a relationship between the apparent dip $\alpha$ of the reflector before migration and its true dip $\beta$ after migration.

From Figure 3.2 we take:

- Length of the ray path: $RA=RB=v\cdot t/2$
- Tangens of the apparent dip before migration: $\tan \alpha = RA/RY$
- Sine of the true dip after migration: $\sin \beta = RB/RY$
- Hence the relationship between $\alpha$ and $\beta$ is given by

$$\tan \alpha = \sin \beta$$

(3.7)

This simple but fundamental relationship between the unmigrated and the migrated dip angle has become famous under the name "migration formula". From this formula it is evident that the apparent dip angle before migration must not exceed 45 degrees and that any steeper dips on the unmigrated section represent artefacts.
Figure 3.2. Schematic illustration of the geometry of the migration process in a homogeneous medium. Vertical axis is depth but due to the constant velocity may be considered as scaled travel time. Before migration the line element A dips at an apparent angle $\alpha$ and is located vertically beneath the coincident source/receiver location $R$ on the stacked, zero-offset section; migration moves the element updip to position $B$ where it dips at the true angle $\beta$ and is parallel to the wave front, i.e. perpendicular to the incident ray.

Formula (3.7) may be extended to establish the relationship between the unmigrated $(X_{\text{unmig}}, Z_{\text{unmig}})$ and the migrated position $(X_{\text{mig}}, Z_{\text{mig}})$ of a reflector (Chun and Jacewitz 1981):

$$X_{\text{mig}} = X_{\text{unmig}} + \frac{v^2}{4} \frac{t \tan \Phi_t}{\Delta x}$$  \hspace{1cm} (3.8)

$$Z_{\text{mig}} = Z_{\text{unmig}} + \frac{v}{2} t \left\{ 1 - \sqrt{1 - \frac{v^2 \tan^2 \Phi_t}{4}} \right\}$$  \hspace{1cm} (3.9)

where $\tan \Phi_t = \frac{\Delta t}{\Delta x} = \frac{2 \tan \alpha}{v}$.
Figure 3.3 shows the thus calculated horizontal and vertical displacements for a range of apparent dip angles ($\alpha$) as a function of TWT. The velocity is taken to be 6.0 km/s which is a representative average value for crustal studies. It becomes evident that even for relatively small apparent dip angles around 10 degrees migration displacements, particularly the horizontal component, are substantial for TWT relevant for crustal studies (5 to 20 s). For bigger apparent dips displacements become huge. For example, an apparent dip of 40 degrees at 10 s TWT will result in a vertical displacement of more than 10 km and a horizontal displacement of more than 20 km! The following lessons should therefore be learnt from this simple analytical consideration:

- Virtually every deep seismic reflection profile needs migration to allow structural interpretation; a simple time-to-depth conversion along vertical ray paths, e.g. by the not so uncommon "multiply-by-three" rule in crustal reflection seismology, will rarely be significant.

- Deep seismic reflection profiles must not be too short. Otherwise dipping reflectors will migrate out of the profile into areas of no seismic coverage thus jeopardizing structural interpretation. In cases of short profile length and/or steep dips lateral migration paths may exceed the profile length thus virtually making any interpretation pure guess work. Estimates of the maximum expected dip angles and the resulting lateral displacements according to (3.8) should be used to estimate the minimum required profile length.

As a first approximation the formulae (3.8) and (3.9) and the results displayed in Figure 3.3 can still considered to be valid for vertically stratified media. The constant velocity $v$ has then to be replaced by the rms velocity $v_{\text{rms}}$, which corresponds to a straight ray path approximation. The accuracy of this approximation degrades with increasing reflector dip, i.e. with increasing deviation of the ray path from the vertical. In exploration seismology the rms concept is still considered to be valid for angles up to 45 degrees and more for velocity analysis, NMO-correction and CMP-stack.

More sophisticated analytical migration formulae which allow to accommodate constant vertical and horizontal velocity gradients are given by Michaels (1977) and Hubral (1978).
Figure 3.3. Horizontal (a) and vertical (b) migration paths as a function of apparent reflector dip and travel time (TWT). The velocity of the overburden is taken to be constant at 6.0 km/s, which is a representative average value for crustal studies.
3.3 Migration Error Analysis

For any computational method it is of utmost importance to obtain a quantitative estimate of the impact of uncertainties in the model parametrisation on the final result. In the case of migration this may be achieved by applying the differential $\delta$ with respect to the travel time $t$ to the analytical migration formulae (3.8) and (3.9). This yields

$$\delta X_{\text{mig}} = \frac{v \ t \ \tan \Phi}{2} \ \delta v$$

$$\delta Z_{\text{mig}} = t \ \delta v - \frac{2 \ t - \frac{v^2 \ t \ \tan^2 \Phi}{4} \ \delta v}{\sqrt{1 - \frac{v^2 \ \tan^2 \Phi}{4}}}$$

These formulae are strictly valid only for constant velocity media and yield only minimum estimates of migration errors because additional complications arising from ray bending in an inhomogeneous medium are neglected.

Figures 3.4 and 3.5 display the migration errors as a function of TWT resulting from velocity uncertainties of 0.1 and 0.2 km/s, respectively, for a range of apparent dips. Again the medium is assumed to have a constant velocity of 6.0 km/s. Figure 3.4 shows that - with the exception of the combination of very steep apparent dips and very high TWT - an uncertainty in velocity of 0.1 km/s only leads to only minor positioning errors. The situation is, however, no longer so favourable for a velocity error of 0.2 km/s (Figure 3.5). Here mispositioning - particularly in the vertical direction - becomes substantial even for only moderately steep apparent dips and moderately high TWT. Hence every effort must be made to determine the average velocity structure used for migration as precisely as possible, preferably within $\pm 0.1$ km/s. In crustal studies this accuracy can only be achieved by seismic wide-angle data. Experience from workshops (cf. Ansorge et al. 1982) in fact shows that for high-quality seismic wide-angle data the average crustal velocities independently inferred by different workers agree within 0.1 to 0.2 km/s.
Figure 3.4. Horizontal (a) and vertical (b) migration errors as a function of apparent reflector dip and travel time (TWT) for an uncertainty in velocity of 0.1 km/s relative to 6.0 km/s.
Figure 3.5. Horizontal (a) and vertical (b) migration errors as a function of apparent reflector dip and travel time (TWT) for an uncertainty in velocity of 0.2 km/s relative to 6.0 km/s.
Thus we may conclude:

- It is desirable to keep the error in the average velocity for the migration of deep seismic reflection data as small as 0.1 to 0.2 km/s. Wherever nearby seismic wide-angle data are available this should generally be feasible.

- In order to reduce "smiling" effects conventional wave equation migration of deep seismic reflection data is sometimes performed with up to 30% too low velocities (cf. Warner 1987). The resulting positioning errors rather reduce than increase the structural interpretability of the data.

In this and the preceding section we have only considered media of constant velocity for which - as seen before - the ray paths are straight. Because of the resulting geometric simplicity this has allowed an analytical treatment of certain important characteristics of migration. The thus inferred migration displacements (Figure 3.3) and migration errors (Figures 3.4 and 3.5) must be considered as minimum estimates since the effects of ray bending in more complicated heterogeneous media are neglected. The great potential of the ray method in seismology in general and in migration in particular, however, lies in its ability to accommodate complex velocity structures. This has been exploited both in forward modelling and in migration. The rest of this chapter will hence be devoted to the application of the ray method to migration in complex media.

### 3.4 Migration Methods Combining Wave and Ray Theory

*Time-to-Depth Migration: Image Rays*

Hubral (1977) was the first to realise that no time migration scheme - due to the locally laterally homogeneous definition of the velocity field - can depropagate the wave field correctly in the presence of lateral velocity variations. To understand this statement it is useful to think of migration in terms of the Kirchhoff summation concept introduced in the previous chapter where the subsurface is supposed to consist of point scatterers each of which gives rise to a diffraction curve. The migrated section, i.e. the depropagated wave field, is then obtained by focusing the seismic energy along each diffraction curve into its apex. If the velocity above a particular point scatterer does not vary laterally then the diffraction curve is approximately hyperbolic with respect to the corresponding rms velocity. Such a diffraction curve is rotationally symmetric and its apex lies vertically
above the corresponding scatterer. If on the other hand the velocity above a particular scatterer does vary laterally then the resulting diffraction curve is no longer rotationally symmetric and its apex does not lie vertically above the corresponding scatterer. Using ray geometric concepts Hubral (1977) has shown that the minimum travel time from each scatterer to the surface is given by the time along what he called an "image ray", a ray that takes off from the scatterer at an angle so that it emerges vertically at the surface. Hence the point where the image ray emerges at the surface corresponds to the lateral location of the apex of the diffraction curve for the corresponding scatterer which in turn corresponds to the point where we want the migration process to focus the seismic energy. Time migration implicitly assumes that the lateral position of the scatterers and the apices of the corresponding diffraction curves are always identical. Evidently, therefore, time migration can only be complete if the velocity structure is 1-dimensional, i.e. when all the image rays are straight over their entire length. If on the other hand the velocity does vary laterally then, as suggested by Hubral (1977) and implemented by Larner et al. (1981), the image ray concept can be used to obtain a correctly migrated section by the following two-step depth migration scheme:

- Migrate the data in the time domain using the best possible 1-dimensional approximation of the true 2- or 3-dimensional velocity structure.
- Convert the above time-migrated section to depth along image rays traced through the true 2- or 3-dimensional velocity structure of the medium.

Larner et al. (1981) have shown that this scheme is very satisfying and efficient provided that the time migration step manages to focus the energy properly. This is only guaranteed as long as the lateral velocity variations are mild enough to be reasonably approximated by a 1-dimensional model.

An important implicit conclusion to be drawn from Hubral's (1977) work is that any ray theoretical approach to migration necessarily results in a depth migration scheme. Apart from the time-to-depth mapping along image rays this has been used to extend the Kirchhoff migration procedure from a classical 1-dimensional time migration method to an efficient and versatile depth migration method or for the depth migration of correlated travel times by inverse normal-incidence ray tracing.
Ray-Kirchhoff Depth Migration

As we have seen in the previous chapter, conventional Kirchhoff migration essentially corresponds to a weighted sum over diffraction hyperbolae at each depth point of the section. The shape of these diffraction hyperbolae is defined by the velocity structure of the overburden which is commonly approximated in a 1-dimensional way. A direct extension of this 1-dimensional "analytical" Kirchhoff migration scheme to the second or third dimension is not straightforward and no practical implementations are known.

Ray theory offers an elegant solution to this problem and allows to overcome the limitations inherent in conventional Kirchhoff time migration. Within the previously mentioned natural restrictions of ray theory, ray-Kirchhoff depth migration schemes can accommodate much stronger lateral velocity variations than the time-to-depth migration scheme of Hubral (1977) and Larner et al. (1981).

There are two basic approaches:

- From every depth point of the migration velocity model a family of rays is traced to the surface. The shape of the resulting travel time curve defines the summation or diffraction curve needed for Kirchhoff migration and all that is left to do is to apply the proper weighting and phase shift factors. This approach has been shown to be applicable to 2- and 3-dimensional velocity structures as well as to stacked and unstacked data (Carter and Frazer 1984; Zhu 1988; Schrader 1990).

- The Kirchhoff migration formula is essentially defined by the Rayleigh II integral (Berkhout 1982) which corresponds to a surface integral over the convolution of the recorded wave field and the gradient of the Green's function of the medium. The Green's function can be determined by paraxial ray tracing (Cerveny et al. 1984; Beydoun and Keho 1987) through the medium. This approach has been followed by Keho and Beydoun (1988).

3.5 Ray-Theoretical Depth Migration of Digitised Line Drawings

At first glance, this approach to migration may look like a step back into the dark pre-computer ages when picked sections had to be painstakingly migrated by hand using ray theoretical concepts. For high-quality shallow data this is undoubtedly true; migration
methods based on the wave equation here normally do an excellent job. For deep seismic reflection data the situation is completely different - since as demonstrated by Warner (1987) and outlined in the preceding chapter - wave equation migration is ill-conditioned for this kind of data. Moreover, the numerical solution of the ray equations and thus ray tracing through truly complex media has just become practically feasible by the advent of fast digital computers.

The first computer-based ray migration method for digitised line drawings was published by Raynaud (1988b). With kind permission of the BIRPS group in Cambridge the code was transferred to the ETH Zürich and its interactive and graphics capabilities were enhanced in the course of this work. This "Zürich" version of the program now runs under the name MIGRAY and is documented in Appendix A.

The method is overall superbly designed and has successfully been used in numerous practical applications (e.g. Warner 1986, 1987; Holliger and Klemperer 1989, in press). The velocity field is parametrised in terms of constant velocity layers which themselves are defined in a rather inflexible way by a series of velocity-depth columns. This model parametrisation allows to avoid the numerical solution of the ray equation system (3.6) and hence makes ray tracing highly efficient but at the same time it is annoyingly tedious and time consuming to construct and change complex models. Moreover, the restriction to constant layer velocities in ray theory is an unnatural one. In certain cases - particularly where strong velocity gradients are present - this limitation does not allow to define the migration velocity model with the desired accuracy.

As outlined before the most accurate velocity information for the migration of deep seismic reflection profiles is provided by seismic wide-angle data. Normal-incidence seismic reflection profiles are intended to sample in detail an unaliased 2-dimensional cross section of the actual 3-dimensional wave field and hence are ideally oriented perpendicular to the structural grain. The opposite is true for reconnaissance seismic wide-angle profiles. Since the negative effects of spatial aliasing increase with decreasing lateral continuity of the "imaged" features seismic wide-angle profiles are preferably oriented parallel to the structural trend. The ideal velocity model for the migration of a deep seismic reflection profile would therefore be based on a series of perpendicular seismic wide-angle profiles. Nowadays seismic wide-angle data are routinely interpreted by forward ray tracing programs following Cerveny et al. (1977) which allow to accommodate lateral and vertical velocity gradients; these velocity gradients will be present in the resulting models of the velocity structure. Trying to simulate such a model
by layers of constant velocity is rather crude physics, may lead to structural features that violate ray theory and may introduce velocity errors that are greater than the 0.1 to 0.2 km/s which we have previously found to be tolerable. For these reasons a substantial part of this thesis work was dedicated to extend Raynaud's (1988b) ray migration algorithm to full ray theory. The corresponding program is called MIGART which is an acronym for Migration by Asymptotic Ray Theory and is described in Appendix B.

In the following the fundamental concepts of ray-theoretical depth migration of digitised line drawings will be discussed. These concepts are the same for any ray migration scheme based on digitised line drawings.

**Digitised Line Drawings as Input Data**

Input to ray-theoretical depth migration according to Raynaud (1988b) is a digitised line drawing of the primary reflections and diffractions encountered in the stacked section. In terms of filter theory the process of line drawing preparation may be described as a convolution of the original seismic data consisting of amplitude, phase and corresponding travel time information as well as noise with the coherency filter of the human eye. Ideally this filtering process would result in a noise-free copy of the original seismic data consisting of the travel times of all the primary reflections and diffractions with the amplitude and phase information removed. It is of course not possible to quantify the accuracy of a presumably "subjective" line drawing but experience has shown that - despite often substantially different interpretative approaches - line drawings independently made by different workers generally agree amazingly well (compare e.g. Gibbs (1987) with Holliger and Klemperer (1989) or Frei et al. (1989) with Figures 5.2 a, b, c, d).

The most important prerequisite for reducing the ambiguity inherent in line drawing preparation is to start off with carefully processed seismic data which not only has an optimized signal-to-noise ratio but also may be expected to be largely free of non-primary coherent energy. Should it not be possible to remove all the non-primary coherent energy - such as multiples and side-swipes - in the processing stage, this may be done in an interpretative way during line drawing preparation (Holliger and Klemperer 1989). A question that remains open for debate is whether line drawing preparation should be accompanied by any further subjective optical filtering, such as rejecting coherent energy below a certain spatial extent (manual spatial deconvolution) or removing the
reverberating character typical of many crustal reflections (manual vertical deconvolution). Assuming that a good job has been done in the processing stage all coherent energy must be considered as being part of the reflectivity of the subsurface, which we are looking for; and since Raynaud (1988b) has shown that corresponding travel times are correctly predicted by ray theory even in the presence of complex overburden I advocate the inclusion of all clearly coherent events in the original line drawing. Additional filtering of this kind, e.g. based on length, dip, bending or "amplitude" of reflection events, may still be performed numerically on the digital line drawing before and/or after migration. Practical problems concerning line drawing preparation are discussed in Warner (1986,1987), Holliger and Klemperer (1989) and Valasek and Holliger (1990).

In the following a digitised line drawing of an individual reflection or diffraction is called a reflection segment. Each reflection segment is defined by N points with N≥2. Thus a reflection segment is made up of N-1 line elements each of which is defined by its two ending points (cf. Figure 3.9).

**Initial Conditions**

As illustrated by Figure 3.2 the problem to solve is to shift each line element updip along the corresponding wave front until it is perpendicular to the incident ray. The geometric construction of the reflection point is the inverse of that used to produce a synthetic zero-offset section by inverse ray tracing (Taner et al. 1970). For a given 2-dimensional velocity model the location of the reflection point is thus uniquely defined by this construction. The problem hence is to define the correct initial conditions for this inverse normal incidence ray tracing procedure. As we have previously seen the initial conditions required for ray tracing are the coordinates of the starting point and the initial slowness of the ray. The former is given as the surface location (R) of the unmigrated reflection point (Figure 3.2); the latter requires the knowledge of the starting angle of the normal-incidence ray and has to be determined separately for each line element.

Figure 3.6 shows an unmigrated line element defined by two points $P_1(x_1,t_1)$ and $P_2(x_2,t_2)$. The vertical axis is TWT and is assumed to be scaled according to the so-called effective velocity $v_{\text{eff}}$ which corresponds to the average velocity along the corresponding normal-incidence ray and is defined as
\[ v_{\text{eff}} = \frac{\int v(s) \, ds}{S} \]

where \( S \) represents the total length of the ray path.

The apparent dip \( \varphi \) is therefore given by

\[ \tan \varphi = \frac{\frac{v_{\text{eff}}}{2} \frac{\Delta t}{\Delta x}} \]

where \( \Delta x = x_2 - x_1 \) and \( \Delta t = t_2 - t_1 \).

\[ \sin \theta = \frac{v_o \tan \varphi}{v_{\text{eff}}} \]

\begin{figure}[h]
\centering
\includegraphics[width=0.5\textwidth]{figure3_6.png}
\caption{Unmigrated line element defined by its end points P1 and P2 lying vertically below the corresponding coincident source/receiver locations R1 and R2. Vertical axis is two-way travel time (TWT) which is assumed to be scaled according to the "effective velocity" relevant for this line element.}
\end{figure}

As is evident from Figure 3.7, which shows the same line element after depth migration, the starting angle \( \theta \) for a particular ray element is determined by the length difference of the two normal incidence rays. Therefore
where $v_0$ is the near-surface velocity immediately below the corresponding source/receiver pair.

If the profile is at angle $\zeta$ to the dip direction $\Delta x$ in (3.14) has to be replaced by $\Delta x'$, which is defined as (see e.g. Levin 1971)

$$\Delta x' = \Delta x \cos \zeta \tag{3.15}$$

Please note that this concept is only valid if $v_{\text{eff}}$ is the same for both rays. Experience shows that this condition is virtually always fulfilled if the velocity model is sufficiently smooth for ray tracing and if the individual line elements are not too long. Otherwise the program can easily check this condition and remove the problem either by discarding the line element, chopping it up into several shorter elements or by slightly changing the crucial parts of the velocity structure.

![Figure 3.7. Line element of Figure 3.6 after ray-theoretical depth migration.](image)

**Ray Tracing**

Knowing the initial conditions of the ray tracing system the migration procedure for an individual reflection segment can now be outlined as follows:

1. Determine the starting angle for the first line element.
(2) For each of the two end points defining the line element trace one ray through the medium. The rays obey the ray equations and Snell's law at interfaces. Ray tracing for a point is finished when the ray has travelled for half the observed TWT of this point.

(3) Repeat (1) and (2) for all the remaining line elements of the reflection segment.

(4) The depth-migrated reflection segment is defined by joining the individually depth-migrated line elements.

The thus defined migration procedure is equivalent to the wave front migration method of Hagedoorn (1954) which not only correctly repositions reflections but also "collapses" diffractions by moving the individual line elements into the apex. Moreover, the change in length of a line element due to migration is indicative of focusing and defocusing effects of the overburden and may be used to calculate "ray migration amplitudes" (Raynaud 1988b). These focusing and defocusing effects predicted by ray theory are, however, not to be confused with the shrinking of the Fresnel zone and the increase in spatial resolution achieved by wave equation migration (cf. Berkhout and Van Wulfsten Palthe 1981; Lindsey 1989).

For the case of constant layer velocities ray theory optionally allows to migrate the velocity structure as well. This can be done by either defining the velocity model in TWT right away or by converting it from depth to TWT along vertical ray paths. The interfaces of this "time model" are then chopped up into smaller segments in order to fulfill (3.14), converted into the line drawing format described above and migrated in the same way as reflection segments. The simultaneous depth migration of the structural model of the subsurface and the observed reflectivity can be useful for interpretative migration. Convergence of the migrated reflections and the geometry of the velocity structure is a possible criterion for the correctness of the latter.

3.6 Algorithmic Aspects of Program MIGART

The only fundamental difference between MIGRAY and MIGART lies in the parametrisation of the velocity model. To those who have never gone through the joys and frustrations of developing and coding a numerical modelling algorithm this necessarily seems to be a "picnic". However, it turns out that the opposite is the case: the
model parametrisation is the decisive part of any modelling and migration algorithm. This argument is supported by the fact that a recent PhD thesis (Schrader 1990) was essentially dedicated to the development of a particular velocity model parametrisation scheme for migration.

In contrast to MIGRAY, which requires constant layer velocities, MIGART is more flexible since it allows the velocity to vary both vertically and laterally within each layer; in that case the ray equations have to be solved numerically.

**Model Parametrisation**

MIGART will be primarily used to migrate deep seismic reflection profiles for which the velocity structure is known in great detail. Such detailed velocity information on a crustal scale is provided by seismic wide-angle angle data which are interpreted with a limited number of forward ray tracing programs. It is therefore sensible to choose a parametrisation scheme which is compatible with that of the most popular forward programs such as the wide-spread SEIS83 (Cerveny and Psencik 1984) and its successors such as RAY84/83 (Luetgert 1988) and RAY87 (Siervo 1988), which are currently used at the USGS and ETH Zürich. Such a velocity model is displayed in Figure 3.8 and can be characterised as follows:

- n layers are defined by n+1 interfaces.
- Each interface must be continuous between the left and right margin of the model and must not intersect another interface.
- Velocities are specified at the intersection points of vertical "speed" lines and the interfaces.
- The program divides the model into rectangles by enveloping each interface with two horizontal speed lines. For converging interfaces a certain minimum vertical separation between the horizontal speed lines will be maintained.

To solve the ray equation system (3.6) the velocity and its gradients must be known along the entire ray path. For the velocity model outlined above this can be done as follows:

Find out in which layer and in which rectangle the particular ray point lies; knowledge of the velocities at the corners and the gradients along the edges of the rectangle then allows to determine the velocity and its gradients at any point within this rectangle by 2-dimensional linear interpolation (cf. Press et al.1986).
Figure 3.8. Schematic illustration of the parametrisation of the velocity model for RAY84/83 (Luetgert 1988), RAY87 (Sierro 1988) and MIGART. The user supplies the geometry of the interfaces (heavy lines), the location of the vertical speed lines (solid vertical lines) and the velocities at the intersections of the vertical speed lines with the interfaces; the program then bounds every interface by two horizontal speed lines (horizontal dashed lines) and thus divides the model into a set of rectangular boxes. For a given point its layer and box number are evaluated which then allows the velocity and its spatial derivatives to be determined by interpolation.

Numerical Solution of the Ray Equations

In two dimensions the ray tracing system (3.6) consists of four ordinary differential equations of the first order. These equations are linear and "non-stiff" (i.e. for the range of seismic velocities the eigenvalues of the system matrix are of the same order of magnitude) and consequently can be expected to be well-behaved during numerical integration. Therefore a standard fifth-order Runge-Kutta algorithm with adaptive step-size and error control (cf. Press et al. 1986) was chosen to integrate these equations. Figure 3.9 shows how MIGART migrates a digitised line element defined by N points.
• In agreement with Zelt and Ellis (1988) the integration step-size in time $t_{\text{step}}$ was allowed to vary according to the complexity of the model

$$t_{\text{step}} = \frac{\sigma}{|v_x| + |v_z|}$$  \hspace{1cm} (3.16)

where $\sigma$ is a constant factor and $v_x$ and $v_z$ represent the spatial derivatives of the velocity field. Zelt and Ellis (1988) have found that a $\sigma$ value between 0.05 and 0.15 normally does a good job. I can confirm this.

• If $t_{\text{step}}$ as determined by (3.16) is outside a specified range the corresponding extremal value is taken.

• The maximum step-size must not exceed the minimum one-way travel time through the thinnest layer of the model.

• The desired accuracy for a single Runge-Kutta step can be estimated as follows: the maximum expected number of integration steps with the allowed error must not exceed the average digitising error involved in the line drawing preparation.

• If the initial step-size $t_{\text{step}}$ determined by (3.16) leads to a too big error the corresponding integration step will be repeated with successively smaller step-sizes until the desired precision is achieved.
Figure 3.9. Algorithmic details of the ray trace procedure used in MIGART.
4. ESTIMATION OF CRUSTAL VELOCITIES

4.1 Historical Aspects, Experimental Techniques and Resolving Power

October 8, 1909, the date of an earthquake in Croatia may be considered as the birthday of crustal seismology. It was on the seismograph records of this earthquake that the Yugoslavian seismologist Mohorovicic (Mohorovicic 1909) detected two distinct travel time branches for both P- and S-waves which led him to the conclusion that the crust has a thickness somewhere between 30 and 60 km, a P-wave velocity of approximately 6 km/s and is underlain by solid mantle with a P-wave velocity of around 8 km/s. Before that date it was supposed amongst earth scientists that a silicic crust of unknown thickness was afloat on more mafic liquid substratum feeding the volcanoes. The seismological boundary between the crust and the mantle is now referred to as Mohorovicic discontinuity (M) or just as "Moho" (see e.g. Steinhart 1967).

Although E. Wiechert had already used large explosive sources for the seismic investigation of the deeper crust as early as 1923 (Wiechert 1923), the next important step in our understanding of the physical structure of the crust is generally attributed to the Austrian seismologist Conrad (Conrad 1925), who recognised a further P- and S-wave travel time branch with intermediate velocities of 6.3 and 3.6 km/s on the records of the Tauern earthquake of November 28, 1923. This led to the interpretation that the continental crust consisted of a slow "granitic" upper part and a fast "basaltic" lower part separated by the so-called Conrad discontinuity (C) or for short "Conrad". It was not until the onset of controlled-source seismology after World War II that this seismic model of the continental crust was considered to be generally too simplistic and in many cases not applicable at all.

Until about twenty years ago most crustal refraction profiles were acquired by recording quarry blasts with mobile instruments. Although this technique was handicapped by the rather ill-defined, low frequency source (see e.g. Burkhardt and Vees 1976) and a wide station spacing (generally around 5 km or more) it nevertheless provided a first impression of the complexity and the lateral variability of the velocity structure of the crust and provided substantial technical know-how for the further development of crustal wide-angle seismology, particularly in central Europe (cf.
This technique has now largely given way to a kind of true wide-angle seismic profiling with minimum shot and station spacings in the order of 50 km and 1 km, respectively, and average parameters no more than twice these values (Ansorge 1989). This trend towards still smaller shot and receiver spacings is continuing today and in different kinds of marine experiments it has already been possible to close the gap between the wide-angle and the normal-incidence seismic technique (e.g. Stoffa and Buhl 1979; Behrens et al. 1986). It is in fact from this kind of highly resolving seismic wide-angle experiments, which combine the spatial resolution of the normal-incidence technique with the large offsets necessary for an accurate determination of the velocity, that we may expect substantial progress in our understanding of the physical structure and petrological nature of the continental crust and the upper mantle.

Resolving Power of Seismic Wide-Angle Experiments

If a receiver array with a station spacing $\Delta x$ is approached by a wave front at an angle $\phi$ to the horizontal through a medium of velocity $v$ then elementary geometric considerations show that all seismic energy with instantaneous frequencies higher than

$$v = \frac{v}{2 \Delta x \sin \phi}$$

(4.1)

is spatially aliased. In the case of a horizontally stratified medium the angle $\phi$ approximately corresponds to the angle of reflection. This means that for an average crustal velocity of $v=6.0$ km/s, a receiver spacing of $\Delta x=2$ km and an angle of the wave front to the horizontal of $\phi=45$ degrees, which is a typical value for the critical angle at the crust-mantle boundary, that all temporal frequencies higher than 2 Hz are spatially aliased.

In crustal wide-angle seismology dominant frequencies are typically between 5 and 10 Hz and the angles of the wave fronts to horizontal are generally bigger than 45 degrees (Ansorge 1989). So does this mean that we virtually cannot resolve anything at all and consequently that all crustal models based on seismic wide-angle data are just fiction? The answer is - of course- no, because the spatial aliasing criterion (4.1) strictly applies to seismic pulses with the same phase only. It does, however, imply that - unlike in reflection seismology - we are generally not able to correlate and interpret individual phases and that consequently the resolution criteria for unaliased seismic data such as the
Fresnel zone diameter for the lateral resolution (Berkhout 1984) are not applicable to seismic wide-angle data. So far no metrics have been worked out to put numbers on the resolving power of spatially aliased seismic data but an educated guess seems nevertheless feasible. In order to be correlated as a wave group a seismic pulse must show coherent energy over several, say five to ten, traces. Assuming an average trace spacing in the order of 1 to 3 km it is therefore safe to say a reflector must be horizontally continuous over several kilometres, i.e. several dominant wavelengths to be unambiguously resolved by a typical seismic wide-angle experiment. Given this lateral continuity we may infer that the minimum thickness of a reflecting bed corresponds to one quarter of the dominant apparent wavelength (Berkhout 1984), which typically will be in the range of approximately 0.5 to 2 km.

**Phase versus Group Correlation**

The first and most critical step in the interpretation of seismic wide-angle data is the correlation of seismic events at individual stations that are considered to arise from the same types of refracted or reflected waves (see e.g. Giese 1976a,b; Ansorge et al. 1982). In the German speaking community of explosion seismologists this process is commonly referred to as "phase correlation". From the semantic point of view this is, as we have just seen, evidently not entirely correct and distracts from the eminently important question about what it is that we are actually correlating.

In marked contrast to spatially active numerical algorithms, such as f-k filtering and migration, the human eye has no problems to detect spatial coherence of seismic signals based on their amplitude and their overall character even if they differ in phase. Since the energy of a wave train travels with the so-called "group velocity" and our eye tends to be caught by the biggest amplitudes, it is probably safe to say that - at least in case of old loosely sampled datasets (see e.g. Figure 4.3) - in the case of seismic wide-angle data the so-called "phase correlation" rather corresponds to a "group correlation", and that consequently the velocity we infer thereof is the group rather than the phase velocity. The group velocity $U$ is defined as

$$U = \frac{x}{t} = \frac{d\omega}{dk} = -\frac{dv}{d(1/\Lambda)}$$  \hspace{1cm} (4.2)
and is related to the phase velocity \( c = \frac{\omega}{k} \), which normally corresponds to the so-called medium velocity, by the relation

\[
U = c - \frac{dc}{d\Lambda} = c + k \frac{dc}{dk} = \frac{c}{1 - \frac{\omega}{c} \frac{dc}{d\omega}}
\]  

(4.3)

where \( \omega \) is the instantaneous frequency, \( \Lambda \) the wavelength, \( k \) the wave number and \( \omega \) the angular frequency. The maximum \( U_{\text{max}} \) of the group velocity \( U \) approximately corresponds to the so-called signal (onset) velocity (see e.g. Mueller 1962 and references therein).

An elastic solid does not disperse sound waves as long as the entire frequency spectrum interferes with the internal structure of the medium in the same way. So for the narrow bandwidth of only about 5 Hz in crustal wide-angle seismology (Ansorge 1989) it is probably valid to assume that the error in estimating the phase velocity from the "group correlated" travel time branches due to dispersion is smaller than the one introduced by the inherent problem of the correlation itself. This is further supported by Ward and Hewitt (1977) who measured only negligible dispersion in sediments for frequencies between 35 and 55 Hz, and by Futterman (1962) who showed that dispersion is normally not relevant to seismic body waves.

Assuming the same amount of dispersion per cycle as Ward and Hewitt (1977) and a bandwidth of 5 Hz (Ansorge 1989) the difference between phase and "group correlated" travel times for a typical wide-angle reflection from the Moho is expected to be in order of 25 to 50 ms which corresponds to a velocity error of 0.01 to 0.02 km/s. This is within the range of systematic travel time errors introduced by the commonly used recording instruments (Deichmann 1984).

**Crustal Structure and Profile Orientation**

As outlined above conventional seismic wide-angle data tend to smooth the velocity structure of the medium over one to several wavelengths (i.e. several km) both laterally and vertically. As a consequence the resolution and reliability of seismic wide-angle data increases with decreasing complexity of the imaged medium. In marked contrast to
normal-incidence seismic reflection profiles it is therefore preferable to orient seismic wide-angle profiles parallel and not perpendicular to the dominating structural grain.

Whilst 3-dimensional effects are probably impossible to handle properly, geometric distortions arising from planar reflectors dipping perpendicular or at an angle to the orientation of the profile can be restored by 2- or 2.5-dimensional migration, respectively. To do this properly we do need, however, two parallel profiles to define the unmigrated dip of the interface. The thus defined line segment can then, based on the previously derived velocity information, be converted to TWT and migrated in the same way as a line element of a digital seismic reflection section.

4.2 Travel Time Interpretation of Seismic Wide-Angle Data

In the last five to ten years it has become standard practice to interpret seismic wide-angle data by numerical ray tracing techniques such as the ones described in chapter 3. Every ray tracing method needs a starting model which is then stepwise adjusted by "trial and error" forward modelling - normally by moving from shallower to deeper structures, i.e. by kind of a manual layer stripping - until the calculated travel time branches match the observed ones. Since ray tracing is not unique on its own, i.e. different models may provide the same calculated travel times, the choice of the starting model is, after travel time correlation, the probably most critical stage in the interpretation of seismic wide-angle data. Whilst an educated parametrisation (e.g. based on experience and/or on a simplistic pre-interpretation of the correlated travel times) of the starting model makes ray tracing a highly efficient and elusive interpretational tool, a badly parametrised starting model will not only turn ray tracing into a painstaking and frustrating process but - even worse - may also lead to features in the final model that may neither be justified by the theory nor by the actually observed data. In the following two simple techniques will be described which I found useful and suitable to derive starting models that combine a low degree of complexity with a high initial compatibility with the observed travel times.
Wiechert-Herglotz Inversion

This technique inverts continuous travel time curves originating from diving waves into a 1-dimensional velocity-depth structure. The velocity may vary continuously and discontinuously with depth, but velocity inversions are not allowed. The latter is in fact guaranteed by the required continuity of the travel time branch.

Herglotz (1907) and Wiechert (1910) have shown that under the above conditions the depth and the velocity of a diving ray at its turning point are linked by the following first-order Abel integral equation:

$$z(x_1) = \frac{1}{\pi} \int_0^{x_1} \frac{v(x_1)}{v(x)} \cosh^{-1}(\frac{v(x_1)}{v(x)}) \, dx$$  \hspace{1cm} (4.4)

Because of the reasonably linear behaviour of $\cosh^{-1}$ for a realistic range of values of $\frac{v(x_1)}{v(x)}$ equation (4.4) can be solved either by direct discretisation or by the discrete recursive formulation (e.g. Hinz et al. 1976)

$$z_n = \frac{1}{\pi} \sum_{i=0}^{n} \Delta x_i \ln \frac{v_n}{v_i} + \sqrt{\left(\frac{v_n}{v_i}\right)^2 - 1}$$  \hspace{1cm} (4.5)

Although several authors (e.g. Meissner 1973; Hinz et al. 1976) have shown that the Wiechert-Herglotz inversion technique can be used to interpret entire seismic wide-angle profiles it strictly is correct only for diving waves, i.e. positive velocity gradients. In crustal wide-angle seismology the only wave which is generally considered to be a truly diving one is the so-called Pg-wave, the wave travelling through the top few kilometres of the crystalline basement. In fact some of the most elusive works of the velocity structure on the uppermost crystalline crust have arisen from careful Wiechert-Herglotz inversions of the Pg travel time branch (Giese 1963; Brändli 1981).
$x^2-t^2$ Method

Whereas the Wiechert-Herglotz inversion technique is applicable well to the inversion of the travel time branches of diving waves the $x^2-t^2$ method is an excellent tool to analyse reflected waves in general.

In the case of one single horizontal layer with a velocity $v$ overlying a half-space at a depth $z=v\cdot t(0)/2$ the travel travel time $t$ is related hyperbolically to the offset $x$

$$t^2(x) = t^2(0) + \left(\frac{x}{v}\right)^2$$

(4.6)

In the case of several plane layers overlying a uniform half-space the travel time-offset relation is still largely, though no more exactly, hyperbolic (Cordier 1985) and we may use the following approximation:

$$t^2(x) \approx t^2(0) + \left(\frac{x}{v_{st}}\right)^2$$

(4.7)

Here $v_{st}$ is equivalent to what is referred to as "stacking velocity" in reflection seismology. Its physical significance is case dependent:

- $v_{st}=v$ for a single horizontal reflector
- $v_{st}=v/\cos\phi$ for a single reflector dipping at an angle $\phi$
- $v_{st}=v_{rms}$ for a series of horizontal reflectors
- $v_{st}=v_{rms}/\cos\phi$ for a series a parallel reflectors all dipping at the same angle $\phi$
- For other cases than the ones listed above $v_{st}$ has no true physical meaning!

$v_{rms}$ is the so-called root-mean-square velocity which is defined as follows:

$$v_{rms} = \sqrt{\frac{\sum_{i=1}^{n} v_i^2 \Delta t_i}{\sum_{i=1}^{n} \Delta t_i}}$$

(4.8)

where $v_i$ is the velocity of the $i$th layer through which the vertical two-way travel time is $\Delta t_i$. 
Since this definition assumes straight ray paths $v_{\text{rms}}$ is in general a few percent larger than the corresponding average velocity $v_{av}$

$$v_{av} = \frac{\sum_{i=1}^{n} v_i \Delta t_i}{\sum_{i=1}^{n} \Delta t_i} \quad (4.9)$$

Thus by plotting the observed travel time branches which we consider to arise from reflections in the $x^2-t^2$ space we shall get a series of straight lines with slopes $1/(s_t)^2$ and intercepts at $t^2(0)$ at the ordinate axis. The corresponding interval velocities and depths can then be obtained by subsequently stripping off the upper layers using formula (4.8) or even more conveniently by using the so-called "Dix formula" derived thereof (Dix 1955):

$$v_n^2 = \frac{v_{\text{rms}}^2(L) \sum_{i=1}^{n} \Delta t_i - v_{\text{rms}}^2(U) \sum_{i=1}^{n-1} \Delta t_i}{\Delta t_n} \quad (4.10)$$

where $v_{\text{rms}}(L)$ and $v_{\text{rms}}(U)$ are the rms velocities to the nth and n-1th reflectors, respectively.

**Starting Model and Subsequent Refinement by Ray Tracing**

By applying the Wiechert-Herglotz inversion technique to the Pg travel time branch and the $x^2-t^2$ analysis to the travel times of the reflected waves we shall get a 1-dimensional model for each shot location within a seismic wide-angle profile. By lining-up all these 1-dimensional models we shall then get a pseudo 2-dimensional model of the crustal velocity structure along the entire profile. This model of the crustal structure can serve as a starting model for forward ray tracing, which nowadays commonly represents the final step in the travel time interpretation of crustal wide-angle seismology. In general this starting model will contain already all the crucial features of the final model and only minor modifications to the geometry and the velocity structure will be necessary during
the ray tracing process to achieve a satisfying fit between the observed and calculated travel time branches.

4.3 Typically Observed Compressional Waves in Crustal Wide-Angle Seismology

Although shear waves are routinely recorded in crustal wide-angle seismology they are rarely analysed thoroughly and for the purpose of migration they do in general not provide any complementary information. Hence, this section will be restricted to the discussion of the most commonly observed and interpreted compressional wave trains in crustal wide-angle seismology.

The compressional waves that can be observed on virtually all crustal wide-angle profiles are the diving wave through the upper crystalline basement commonly referred to as Pg (Banda et al. 1982; Deichmann and Ansorge 1983; Deichmann 1984) and the wide-angle reflection from the Moho commonly referred to as PmP.

From what we have worked out in the previous section we may therefore expect that the Wiechert-Herglotz inversion of the Pg travel times will give us a good idea of the average velocity structure of the upper crystalline crust (Giese 1963; Brändli 1981) whereas the \( x^2 - t^2 \) analysis of the PmP travel time branch will allow us a fair estimate of the average crustal velocity and thickness.

The diving wave from the crust-mantle boundary, known as Pn, is an excellent indicator of the average velocity of topmost mantle. Its amplitude, and thus its detectability, however, strongly depends on the velocity gradient of the topmost mantle (Braile and Smith 1975) and on the overall signal-to-noise ratio of the data.

Amongst the locally observed reflections from intra-crustal discontinuities the so-called PiP reflection from the intermediate Conrad discontinuity is not only the oldest but still also the most debated one (see e.g. Rothé and Peterschmitt 1950; Litak and Brown 1989). Additionally, there often is, particularly in Phanerozoic areas, convincing evidence from the amplitude behaviour of the Pg wave and the presence of wide-angle and normal-incidence reflected phases for an upper crustal zone with significantly reduced velocity (e.g. Mueller and Landisman 1966; Mueller 1977; Müller and Mueller 1979).
In general these intra-crustal phases are, if present at all, much less well-defined than the Pg and PmP wave trains and consequently their correlation and interpretation are not unambiguous. Experiences made with both synthetic and real data at the CCSS (Commission on Controlled Source Seismology) workshop at Karlsruhe in 1977 confirm this statement: largely independent of the applied inversion techniques, there is a general agreement on the gross crustal structure such as the average velocity gradient in the topmost basement, the average crustal thickness and velocity and possibly also on the depth of major crustal discontinuities. However, there may be widely differing views on the details of the velocity structure, which primarily arise from differences in correlation (Ansorge et al. 1982).

Thus, we may conclude that the velocity structure of the topmost 5 km and the average crustal velocity and crustal thickness represent rather safe values but that some caution with respect to the details of the crustal structure is in general justified.

4.4 Alpine Wide-Angle Profiles

Figure 4.1 shows a schematic tectonic map of Switzerland with the location of the seismic wide-angle and normal-incidence reflection profiles relevant to this work superimposed. These profiles were chosen because they run more or less parallel to the strike of the major tectonic units of the Alps and thus may be expected to resolve the gross trends of the crustal velocity structure across this part of the Alpine arc.

The Säntis-Jaunpass (SAE-JAU) profile has been interpreted by Maurer (1989) using 2-dimensional ray tracing. The profiles Lago Bianco-Eschenlohe (LB-ES), Lago Bianco-Pustertal (LB-PU), Lago Bianco-Lago Lagorai (LB-LL), Lago Bianco-Verona (LB-VE) and the Swiss part of the Alpine Longitudinal Profile (ALP75) have been interpreted by Egloff (1979). The Swiss part of ALP75 has been previously interpreted by Ottinger (1976). Both, Egloff and Ottinger, used a technique that allowed them to calculate all the reflected and refracted arrivals from a stack of uniformly dipping parallel layers. Yan and Mechie (1989) attempted a complete interpretation of the entire ALP75 profile using 2-dimensional ray tracing. Deichmann et al. (1986) used 2-dimensional ray tracing and amplitude studies to interpret the Southern Alps profile (SUDALP77) profile;
they also provided a ray-traced reinterpretation of LB-VE which is consistent with their interpretation of SUDALP77.

All these interpretations had the aim to unravel as much as possible of the details of the crustal velocity structure and consequently I refer to them as *maximum* models of velocity structure. The purpose of my work was not to check these interpretations and possibly come up with alternative maximum models but rather to obtain a series of vertically smoothed and laterally consistent *minimum* models that could then be combined to a velocity model suitable for the depth migration of the NFP20 deep seismic reflection data.

These minimum models of the velocity structure of the continental crust are obtained as follows:

- Correlate only strong, laterally continuous wave groups.
- Derive a 1-dimensional velocity model of the topmost basement by a Wiechert-Herglotz inversion of the Pg travel time branch.
- Obtain a 1-dimensional velocity structure of the middle and lower crust by $x^2 \cdot t^2$ analysis of the travel time branches associated with intracrustal wide-angle reflections. Hereby each travel time branch results in exactly one interface.
- If a Pn wave group is detectable use it to estimate the velocity of the topmost mantle. Otherwise choose a value between 7.8 and 8.2 km/s.
- Refine the resulting 1-dimensional model by ray tracing. Additional interfaces and 2-dimensional features (dipping or curved interfaces, lateral velocity gradients) must not be introduced if they are not explicitly required by the seismic data. If no Pn wave group could be correlated then the velocity of the topmost mantle is now chosen to improve the fit between the observed and calculated critical distance of the PmP reflection.

For a particular shot location in a seismic wide-angle profile the above procedure leads to a minimum model of the velocity structure of the crust which fulfills the same criteria as the so-called minimum 1D model in local earthquake tomography (Kissling 1988). The only major difference between the tomographic and wide-angle minimum models are that for the latter the coupling between velocity and geometry is weaker and that the wide-angle minimum model may, though does not need to, be extended into the second dimension. Such a minimum model meets all the basic requirements of a migration velocity model described in chapter 2 and essentially corresponds to Berkhout's "macro-model" (e.g. Berkhout 1987). In the case of coincident high-quality seismic
reflection and refraction data, however, substantial refinements of this scheme of a migration velocity model may be possible and justified.

Figures 4.2 to 4.8 show the minimum models resulting from the application of the above recipe to the Alpine seismic wide-angle profiles shown in Figure 4.1. As mentioned above, the purpose of this reinterpretation was to obtain a migration velocity model across the Alpine arc by combining the minimum models obtained from the individual profiles. As a consequence locally weak but laterally continuous wave groups were considered whilst locally strong but laterally discontinuous wave groups had to be discarded. Of the generally quite numerous travel time branches correlated by the previous workers (see e.g. Figures 4.2 and 4.3) only those associated with the Pg, PmP and an intracrustal wave train termed PiP survived this rigorous low-pass filtering process. The travel time correlations for these wave groups made by the previous workers were left unchanged. This simplistic subdivision of the crust into a slower upper part ($v_p \approx 6.0 \text{ km/s}$) and a faster lower part ($v_p \approx 6.5 \text{ km/s}$) is essentially compatible with the "standard" model of the Variscan crust of central Europe derived by Rothe and Peterschmitt (1950) from the seismic records of the Haslach explosions.

The velocity structure of the sedimentary cover of the Säntis-Jaunpass profile (Figure 4.2) was adapted unchanged from Maurer (1989) and therefore represents rather a maximum than a minimum interpretation. The low-pass filtering process leading from Egloff's (1979) maximum model of the crustal velocity structure to the corresponding minimum model relevant to this work is illustrated in Figure 4.3 on the example of the Lago Bianco-Eschenlohe profile. The numerical values of the velocity-depth functions of the "minimum models" displayed in Figures 4.3 to 4.8 are given in Table 4.1.
Figure 4.1. Schematic tectonic map of Switzerland with the location of the seismic wide-angle (wiggly lines) and normal incidence reflection (solid lines) profiles superimposed. The dotted line represents the Aar Massif profile shot in 1988 which is not yet interpreted.


Figure 4.2. Minimum model of the Säntis-Jaunpass profile (SAE-JAU) showing the original travel time correlations and the detailed structure of the sedimentary cover used by Maurer (1989) for the derivation of his maximum model.
Figure 4.3. a) Original seismic record section of the Lago Bianco-Eschenlohe profile (LB-ES) as compiled by Behnke (1969).

b) "Phase" correlations made by Egloff (1979) for the derivation of his maximum model (see below).
Figure 4.3 (continued). c) Travel time branches adopted from Egloff (1979) for the derivation of the minimum model relevant to this work. d) Comparison of the maximum model for the profile Lago Bianco-Eschenlohe (LB-ES) resulting from the correlations shown in b) and the corresponding minimum model resulting from the correlations shown in c).
Figure 4.4. Maximum (Yan and Mechie 1989) and minimum models of the profile ALP75 at the intersection with the EGT.
LAGO BIANCO-PUSTERTAL

![Graph showing depth vs. velocity for Egloff (1979) and This work models at the intersection with the EGT.]

**Figure 4.5.** Maximum (Egloff 1979) and minimum models of the profile LB-PU at the intersection with the EGT.
Figure 4.6. Maximum (Egloff 1979) and minimum models of the profile LB-LL at the intersection with the EGT.
Figure 4.7. Maximum (Deichmann et al. 1986) and minimum models of the profile LB-VE at the intersection with the EGT.
Figure 4.8. Maximum (Deichmann et al. 1986) and minimum models of the profile SUDALP77 at the intersection with the EGT.
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**Table 4.1.** Velocity-depth functions of the "minimum models" displayed in Figures 4.3 to 4.8. For location see Figure 4.1.
5. CASE STUDY: NFP20 EASTERN AND SOUTHERN TRAVERSES

5.1 Deep Seismic Reflection Data

In this chapter the practical problems involved in the preparation of digital seismic line drawings from the original seismic reflection data and the subsequent ray-theoretical depth migration are addressed.

As an illustrative and challenging example the NFP20 eastern and southern deep seismic reflection traverses across the Swiss Alps (Frei et al. 1989) have been chosen. These traverses run subparallel to the Alpine segment of the European Geotraverse (EGT) (cf. Mueller and Banda 1983) and consist of four major seismic reflection profiles (ET, S1, S3, S5) crossing the entire arc of the Central Alps from the sub-Alpine Molasse in the north to the edge of the Po Basin in the south (see Figure 4.1). The individual profiles range from 5 (line S5) to 95 km (line ET) in length. Although logistic problems did not allow to shoot this Alpine reflection traverse in the form of one single profile, the ultimate goal of the NFP20 deep seismic reflection profiling campaign was to obtain a detailed unified acoustic image of the crust below the central Alps. The problem to be solved is therefore not only the imaging process sensu strictu, i.e. the depth migration of the observed reflectivity, but also the combination of the individual seismic lines into one single profile.

*Digital Seismic Line Drawings of the NFP20 Eastern and Southern Traverses*

The acquisition and processing strategies applied to the NFP20 deep seismic reflection data have been discussed in detail by Valasek et al. (1990) and consequently only those aspects relevant to line drawing preparation are recapitulated here.

The entire eastern traverse (line ET) and line S3 of the southern traverse were acquired using both Vibroseis and explosive sources. The Vibroseis source has a controlled source spectrum and provides a high resolution at upper crustal levels (down to approximately 5 s TWT) but, at the high ambient noise levels common in Switzerland, generally fails to penetrate the deeper parts of the crust. This energy deficiency at lower crustal levels was compensated by shooting strong but irregularly spaced explosive
charges into the same receiver spreads (charge sizes vary roughly between 100 and 1000 kg; shot spacing is approximately 10 km on average).

In order to portray the reflectivity in the subsurface of lines ET and S3 as accurately as possible the line drawings were prepared from the Vibroseis data in the upper part (i.e. down to about 5 s TWT) and from the dynamite data in the lower part. The process of abstracting the processed seismic data into a line drawing of the observed reflectivity is discussed in chapter 3 (see also Holliger and Klemperer (1989), Valasek and Holliger (1990) and Warner (1986, 1987)) and is illustrated in Figure 5.1 with the example of the northernmost 15 km of the eastern traverse (line ET). Every effort was made to conserve as much as possible of the internal characteristics of the various reflectivity patterns observed in the different tectonic units (Valasek et al. in press).

In contrast to lines ET and S3, lines S1 and S5 were acquired using irregularly spaced explosions only and for that reason their resolution in the topmost 2 to 3 s TWT is much lower than the one of the lines ET and S3. Figures 5.2 a, b, c and d show the unmigrated line drawings digitised from lines ET, S1, S3 and S5.

5.2 Combination of the NFP20 Eastern and Southern Traverses

Migration Before or After Profile Combination?

In order to obtain a unified acoustic image of the crust across the entire Alpine arc the individual seismic lines of the NFP20 eastern and southern traverses must be combined into one single profile. Now that we have all the prerequisites for a successful depth migration, i.e. migration velocity structure (Figure 5.3) and digital line drawings (Figures 5.2a, b, c, d) of the seismic reflection data, the obvious question that arises is whether to do this combination before or after migration. Unlike migration algorithms based on the wave equation ray-theoretical migration does not suffer from edge effects, i.e. discontinuities in the boundary/initial values. Hence one might be tempted to argue for a combination of the individual profiles after migration. Since our velocity model is derived from seismic wide-angle profiles perpendicular to the seismic reflection traverses the contours of this velocity model (Figure 5.3) need migration as well and the reliability of the migration of such a sparsely sampled dataset critically depends on the number of data points and thus on the lateral extent of the model.
Figure 5.1. Schematic illustration of the line drawing preparation process for the northernmost 15 km of the NFP20 eastern traverse. Horizontal and vertical scales are approximately 1:1 for an average crustal velocity of 6.0 km/s.

Top left: Topmost 4 s TWT of the processed, multifold Vibroseis data used for line drawing preparation in the upper crust.

Bottom left: 7 to 14 s TWT processed, single-fold dynamite data used for line drawing preparation in the middle and lower crust.

Right: line drawing derived from the seismic data on the left-hand side.
Figure 5.2. Unmigrated digitised line drawings prepared from the NFP20 eastern and southern traverses. Horizontal (distance in km) and vertical (TWT in s) scales are 1:1 for an average crustal velocity of 6.0 km/s. For location see Figure 4.1.

a) Unmigrated line drawing of line ET (Vibroseis and dynamite combined)
b) Unmigrated line drawing of line S1 (dynamite only)
c) Unmigrated line drawing of line S3 (Vibroseis and dynamite combined)
d) Unmigrated line drawing of line S5 (dynamite only)
In our case the velocity model would have to be chopped up into segments ranging from 5 (line S5) to 95 km (line ET) in length. On some of these segments the geometry of the velocity model would have to be interpolated which necessarily would result in considerable uncertainties of the migration velocity model and consequently in inconsistencies and errors in the subsequent combination process. According to the criteria developed in chapter 3 profiles S1 and S5 (Figure 4.1) are too short to be migrated independently because in such a case we do not only lose information by reflectivity migrating out of the profile but necessarily we also lack reflectivity migrating in from outside the profile; this results in a biased acoustic image of the covered subsurface. The migration of short profiles combined through one consistent velocity model can at least partially overcome this problem of insufficient profile length. Provided that the data can be projected into one single profile in a suitable way, the simultaneous depth migration of the reflection data and the velocity structure after projection and combination therefore seems to be preferable.

Finally, one may argue that migration after projection and combination is the more independent and therefore more objective approach because the temptation of an interpretative combination of the seismic data - such as to match the known geological structure or to reduce the amount of conflicting dips - is not given.

*How to Combine the Individual Profiles?*

Since the NFP20 eastern and southern traverses were acquired subparallel to the Alpine segment of the EGT which runs perpendicular to the major tectonic units (Figure 4.1), it is sensible to project the individual seismic lines into this profile. But how?

Laubscher (1990) connected the eastern traverse (line ET) and the northern end of the southern traverse (line S1) by projecting along the strike of the Aar Massif. Because of the dramatic changes of the structural style and strike direction across the Alpine arc, particularly south of the Aar Massif, this approach is problematic and, not unexpectedly, led him to find a pronounced discrepancy between the geologically mapped location of the Insubric Line and its interpreted expression in the seismic data. Moreover, the reflections considered to originate from the Moho along lines ET and S1 became unaligned (Laubscher 1990).

For the purposes of this work the combination of the individual profiles of the NFP20 eastern and southern traverses aims at obtaining a unified acoustic image which is
Figure 5.3. Unmigrated migration velocity model (cf. Rothe and Peterschmitt 1950) in two-way travel time (TWT) along the EGT derived from the velocity-depth functions of the seismic wide-angle profiles at their intersection points with the EGT (cf. Figures 4.1 to 4.9). Numbers are migration velocities in km/s. Thin solid line: base of sediments; heavy dashed line: "Conrad"; heavy solid line: Moho. Zero distance corresponds to the northernmost Vibroseis-CMP of the NFP20 eastern traverse (Swiss coordinates: x=743830, y=231530). Arrows indicate the intersection points with the seismic wide-angle profiles (Figures 4.1, 5.5 and 5.6); the intersection with the Säntis-Jaunpass profile (SAE-JAU) is located at a distance of approximately 6 km. Refraction profiles: SAE-JAU: Säntis-Jaunpass (Maurer, 1989); LB-ES: Lago Bianco-Eschenlohe (Egloff 1979); Alp75: Alpine longitudinal profile (Yan and Mechie 1989); LB-PU: Lago Bianco-Pustertal (Egloff 1979); LB-LL: Lago Bianco-Lago Lagorai (Egloff 1979); LB-VE: Lago Bianco-Verona (Deichmann et al. 1986); SUDALP77: Southern Alps profile (Deichmann et al. 1986). For location see Figure 4.1.
Figure 5.4. Bouguer gravity map of Switzerland with the gravity effects of the Molasse Basin and the Ivrea Body subtracted (from Kissling 1980, 1982) showing schematically the location of the EGT (wavy line) and the seismic reflection profiles relevant to this work. Lowercase letters (a,b,c,d,e) refer to particular features of the gravity field discussed by Kissling (1980, 1982).
representative at a crustal scale. As a consequence any approach relying on surface geology only - even when integrated on a regional scale - is problematic because - in general - the upper crust of the Alps must be expected to be detached from the middle and lower crust due to flake tectonics. On the other hand Bouguer gravity anomalies primarily do reflect the gross structure of the crust as a whole (Kissling 1980, 1984; Kissling et al. 1983) and hence their regional trend is considered to be a suitable projection direction for the purposes of this work. The gravity effect of the Ivrea Body must be subtracted from the original Bouguer gravity anomalies because it is assumed to be not present in the subsurface imaged by the NFP20 eastern and southern traverses (Kissling 1980, 1984; Schwendener 1984; Schwendener and Mueller 1990). As demonstrated by Kissling (1980, 1984) the trend of the thus "Ivrea-corrected" Bouguer gravity anomalies closely follows an east-west orientation for the entire region relevant to this study, i.e. from the southern end of the eastern traverse to the edge of the Po Plain near Chiasso (Figure 5.4).

5.3 Depth Migration

Migration Velocity Model

All seismic wide-angle profiles relevant to this study do intersect the EGT and with the exception of the Säntis-Jaunpass profile all of them do image the entire crust below the EGT (Figure 4.1). Therefore, the migration velocity model can be constructed without projection (again with the exception of the Säntis-Jaunpass profile) directly from the velocity-depth functions of the seismic wide-angle profiles at their intersection points with the EGT (Figures 4.2 to 4.8). As evident from Figures 4.2 to 4.8 lateral velocity variations are rather mild on a large scale. Therefore a migration velocity model with the following constant interval velocities was considered to be appropriate (cf. Rothé and Peterschmitt 1950):

- Sediments (sub-Alpine Molasse, Southern Alps): 5.0 km/s (Stäuble 1990; Deichmann et al. 1986).
- Upper crust: 6.1 km/s
- Lower crust: 6.5 km/s
- Upper mantle: 8.1 km/s
The geometry of the sediments of the sub-Alpine Molasse and the Southern Alps was inferred directly from the unmigrated seismic reflection data. Figure 5.3 shows the resulting *unmigrated* velocity model along the EGT. This velocity model is, as already mentioned before in chapter 4, a minimum or "macro-model" (Berkhout 1987) of the velocity structure of the subsurface and its only purpose and justification is to allow an accurate depth migration of the observed reflectivity.

**Depth Migration, i.e. Image Reconstruction**

Figure 5.5 shows the unmigrated composite time section resulting from the projection of the NFP20 eastern (Figure 5.2 a) and southern (Figure 5.2 b,c,d) traverses along the east-west running trend of the "Ivrea-corrected" Bouguer gravity anomalies (Figure 5.4) onto the EGT (Figure 4.1). Though this profile already shows some aspects of the integrated fine structure of the crust below the central Alps, it still represents a severely distorted acoustic image (as e.g. illustrated by the numerous conflicting reflector dips) because migration and depth conversion have not yet been performed. Figure 5.6 shows the image resulting from the ray-theoretical depth migration of the time section shown in Figure 5.5 using the velocity structure sketched in Figure 5.3.

The most crucial test for any projection and migration method is the correlation between the resulting acoustic image and known or interpretable parts of the geological structure of the subsurface. In the case of the NFP20 eastern and southern traverses the mylonite zone of the Insubric Line is the only well-documented geologic feature that is believed to be unambiguously imaged by the seismic data. As can be seen in Figure 5.6 the migrated reflections from the Insubric Line neatly project into its mapped outcrop location. Moreover, a correctly performed projection and migration procedure does result in a spatial deconvolution of the dataset, i.e. increases its spatial resolution by removing the effects of geometric spreading and reducing "cross dip" effects (Berkhout and Van Wulfoten Palthe 1979). Ray theory assumes that the input data have perfect resolution and therefore does not remove the effects of geometric spreading. However, the amount of conflicting dips has been drastically reduced and the "crispness" - and thus the interpretability - of the migrated section have been increased (compare Figures 5.5 and 5.6). Finally, the simultaneous depth migration of the seismic reflection and refraction data leads to an amazingly good agreement between the refraction Moho and the base of the reflective lower crust, i.e. the "reflection Moho" (cf. Klemperer et al. 1986; Braile
Figure 5.5. Unmigrated time section resulting from the projection of the NFP20 eastern and southern traverses onto the EGT along the east-west running trend of the "Ivrea-corrected" Bouguer gravity anomalies (Kissling 1984). Zero distance corresponds to the northernmost CMP; EGT shot D (Säntis) corresponds to the location of the Säntis-Jaunpass profile (SAE-JAU). Horizontal and vertical scales are 1:1 for an average crustal velocity of 6.0 km/s.
Figure 5.6. Depth migration of Figure 5.5. Velocities used for migration: sediments 5.0 km/s (Molasse Basin and Southern Alps), upper crust, 6.1 km/s, lower crust 6.5 km/s, upper mantle 8.1 km/s (Figure 5.3). Clear, unambiguous wide-angle reflections are represented by stars, doubtful ones by crosses. Please note that the interfaces of the velocity model (depicted by stars and crosses for the 'Conrad' and the Moho) have been migrated as well. Horizontal and vertical scales are 1:1. For location see Figure 4.1.
and Chiang 1986). On the other hand the generally good agreement between the "Conrad" discontinuity interpreted from the seismic refraction data and the top of the reflective lower crust (cf. Mooney and Brocher 1987; Wever 1989) suggested by Figure 5.6 should not be over-emphasized because overall both features are ill-defined.

Although the points made above do by no means represent a formal proof of the correctness of the projection and the migration approach taken in this work they nevertheless significantly increase the confidence one may have in the resulting depth-migrated acoustic image of the Alpine crust as shown in Figure 5.6.

5.4 Interpretation

Basic Problems

Whilst based on surface geology and/or well information seismic reflections from the sedimentary cover can be correlated with distinct lithologies (e.g. Holliger and Klemperer 1989), the interpretation of reflections from within the crystalline basement is much more difficult. Deep seismic reflection profiles represent acoustic snapshots of the present physical state of the crust, which almost everywhere is the result of a complicated, multiphase tectonic history. Moreover, every tectonic event can have different impacts at different crustal levels (e.g. Furlong and Fountain 1986; McKenzie and Bickle 1988), simultaneously create new reflectivity patterns and destroy older ones or find no expression at all in the crustal reflectivity. Therefore, in addition to the uncertainties on the physical origin of deep seismic reflections (e.g. Matthews 1986; Warner 1990) there often is a considerable uncertainty on the age of their formation (cf. Klemperer et al. 1990). As a consequence the tectonic interpretation of deep seismic reflection data is - even after proper processing and migration - anything but straightforward and every effort must be made to substantiate it by using complementary geological and geophysical information.

The surface geology of the Alps has been extensively studied and at least for the most recent deformation phases has led to essentially consistent large-scale tectonic models (cf. Trümpy 1980; Coward and Dietrich 1989). Moreover, the eastward axial plunge of the tectonic grain allows to project geologic structures observed at the surface further to the west to depths of 5 to 10 km in the range of the NFP20 eastern and
southern traverses (Pfiffner et al. 1990). In the central Alps the amount of surface
geology that can be more or less directly correlated with the imaged reflectivity (such as
the Penninic nappes and the Insubric Line) is therefore probably unique. In addition to
these geological constraints for the upper crust there is a wealth of complementary
geophysical information which is particularly relevant at middle and lower crustal levels
(Mueller et al. 1976, 1980). In this context the seismic wide-angle data discussed in
chapter 4 (see Figures 4.1 to 4.8) are of upmost importance because their orientation
parallel to the structural trend as well as the relatively dense lateral spacing of the
individual profiles (approximately 20 km on average) make them a formidable source of
information for the large-scale velocity structure of the Alpine crust (Mueller et al. 1980)
and as such also an ideal database for the derivation of a suitable migration velocity
model.

Review of Some Models of Alpine Tectonics
The following review is by no means intended to be complete. Its purpose is to present
some corner stones which in my opinion have fruitfully contributed to the still on-going
debate on the tectonic and geodynamic evolution of the Alps.

Based on tectonic balancing considerations already Ampferer (1906) came to the
conclusion that substantial amounts of crustal material had disappeared during Alpine
compression. Although our present ideas of plate tectonics were not to be born until some
sixty years later and not even the distinction between crust and mantle had yet been
worked out, Ampferer (1906) quite clearly depicted a subduction process, which he
descriptively called "Verschluckung" and "Unterströmung", to account for the observed
surface geology. Amongst other early workers also Argand (1916, 1924) and Staub
(1928) have developed large-scale tectonic models for the Alpine orogeny. Despite the
fact that these models were only based on the courageous extrapolation of surface
geology and brilliant intuition - and therefore led to rather rough sketches of the deep
orogenic structure - many of the crucial features still have their validity by today's

Another big step towards a unifying tectonic model was made by Laubscher (1970,
1971) who, still largely based on geological evidence, tried to explain the evolution of the
Alps in the light of modern plate tectonics. This model suggests the subduction of the
European lithosphere beneath the African plate at a rather low angle until the onset of
continental collision in the Eocene. The subsequent near-vertical orientation of the subduction zone and the simultaneous bivergent subduction of both the European and the African plates, including unspecified amounts of crustal material, is then postulated to account for the steepening of the root zone, the backthrusting of the Penninic and Austro-Alpine nappes and the late neo-Alpine crustal shortening inferred for the Southern Alps (Laubscher 1970, 1971, 1985).

The presence of such a near-vertical "lithospheric root" of relatively high velocity, high density and low temperature beneath the axial zone of the Alpine arc was found to be compatible with the observed dispersion behaviour of seismic surface waves (Panza and Mueller 1979; Panza et al. 1980), teleseismic P-wave delay times (Baer 1979, 1980) and the positive residual gravity anomalies after having stripped off the gravity effect of the crust (Kissling 1980, 1982; Kissling et al. 1983; Schwendener 1984; Schwendener and Mueller 1990). Based on the assumption of a vertical orientation of the subduction zone Kissling et al. (1983) and Werner and Kissling (1985, 1988) have developed kinematic and dynamic models that cannot only account for the positive residual gravity anomaly but also for the inferred neo-Alpine uplift rates. The dynamic model of Werner and Kissling (1985, 1988) is similar to the more general considerations of Fleitout and Froidevaux (1982) who have shown that the density anomalies induced by lithospheric subduction and thickening may act as a self-sustaining driving force during the process of mountain building. For the special case of the Alps Fleitout and Froidevaux's (1982) general dynamic model is very satisfying in that it not only offers an attractive explanation for the on-going compression in the Alpine arc (cf. Mueller 1984) after continental collision but also naturally incorporates major tectonic features in the foreland of the Alps such as a flexural Molasse Basin and rift zones. Moreover, Fleitout and Froidevaux (1982) showed that this dynamic model does not critically depend on the geometry of the subduction zone and the associated thickening of the lithosphere. Karner and Watts (1983) have modelled the observed gravity anomalies and Moho topography (Mueller et al. 1980) by loading an elastic lithosphere with the known load of the topography and unknown subsurface loads. Although it turns out that substantial subsurface loads must be present, which is essentially a reproduction of the results of previous isostatic studies by Kissling (1980) and Klingelé and Kissling (1982), this approach does not consider the effect of a lithospheric root sensu strictu.

Hsü (1979) presented a model for the post-Eocene Alpine tectonics in which no lithospheric subduction occurs and the entire neo-Alpine shortening of at least 100 km
(Trümpy 1980) is taken up by the internal deformation of the European plate. This hypothetical reconstruction results in a steep contact zone between the Aar Massif and the Penninic Front and avoids the problems arising from the subduction of large amounts of crustal material. Butler (1986) instead favoured - based on rather detailed tectonic balancing in the western Alps - the subduction of substantial amounts of lower crustal material at a shallow angle during the neo-Alpine phase. In contrast to Hsü (1979) this tectonic model requires the frontal Penninic nappes being thrust onto the southern flank of the external massifs at a rather low angle. To avoid isostatic problems in the still subsiding Po Basin and to account for the postulated continuity of the refraction Moho across the Alps (Mueller et al. 1980) Butler (1986) suggested that eclogite metamorphism below 50 km raised the density and the velocity of the subducted crust to the corresponding values of the upper mantle. Based on tectonic balancing considerations and the preliminary interpretation of the unmigrated NFP20 eastern and western traverses (Schweizerische Arbeitsgruppe für Reflexionsseismik 1988) and the unmigrated ECORS-CROP profile across the western Alps (Bayer et al. 1987) Laubscher (1988) came again to the conclusion that substantial amounts of lithospheric - including crustal - material must have been subducted. In contrast to Butler (1986) and Spakman (1990) Laubscher (1988), however, still favours a vertical orientation of the lithospheric slab at depth.

Nowadays most geologists agree on a post-Eocene tectonic model involving substantial - albeit unknown in volume and geometry - southward subduction of European lower crust. The huge stack of upper crustal rocks recognized particularly in the Penninic domain simply requires the "disappearance" of lower crustal rocks from which it was detached (cf. Trümpy 1980; Laubscher and Bernoulli 1982; Coward and Dietrich 1989). On the other hand estimates of post-Eocene shortening of the European crust are rather ill-constrained. Most authors, however, agree that it must be in the order of 100 km, possibly much more but certainly not much less (cf. Trümpy 1980; Coward and Dietrich 1989). Following the above arguments then - in analogy to the Penninic domain - also the Southern Alps, i.e. the Adriatic plate, requires the northward shortening of some 50 km in order to accommodate the observed late neo-Alpine southvergent thrusting of the upper crust (cf. Laubscher 1985, 1989; Roeder 1989).
Tentative Interpretation of the Combined Depth-Migrated NFP20 Eastern and Southern Traverses

At the northern end of the depth-migrated profile shown in Figure 5.6 there is an almost perfect agreement between the refraction and the reflection Moho (cf. Klemperer et al. 1986; Braile and Chiang 1986). After migration they both dip south at an angle of approximately 15 degrees. From the fact that the basement north of the Aar Massif has seen no or only little internal deformation during the Alpine orogeny (Pfiffner et al. 1990) we may conclude that the origin of the reflections from the European Moho and probably also of the sporadic "Conrad" reflectors are of pre-Alpine age. With the onset of the upper crustal Penninic reflectivity (at a distance around 40 to 50 km) the lower crust and the Moho abruptly cease to be reflective. The strong, multicyclic and laterally discontinuous Penninic reflectivity reaches down to a depth of 15 to 20 km. It is considered to arise from a "subhorizontal" stack of crystalline nappes separated by thin slivers of sediments (Pfiffner et al. 1990).

Just north of the Insubric Line at a distance of 70 to 90 km and a depth range of 50 to 60 km (Figure 5.6) there is a deep reflection package whose overall character, dip and extrapolated depth match those of the European reflection Moho lost some 30 km further north. If this feature in fact represents the southern continuation of the European reflection Moho lost with the onset of the strong upper-crustal reflectivity in the Penninic domain then our acoustic image clearly depicts the subduction of at least some 10 km of the lowermost European crust below the Adriatic promontory of the African plate at a low angle (15 to 20 degrees). As evident from Figure 5.6 this interpretation is compatible with the migrated Moho points inferred from the seismic wide-angle data. In fact an amazingly similar topography of the Moho and the lower crust emerges from seismic reflection and refraction data in the western Alps (Bayer et al. 1987; ECORS-CROP Deep Seismic Sounding Group 1989) and in the Pyrenees (Daignières et al. 1982; ECORS Pyrenees Team 1988; Choukroune and ECORS Team 1989). Similar to the southward downdip of the European Moho and lower crust and in agreement with the strongly asymmetric Alpine "Moho trough" interpreted by Mueller et al. (1980), Figure 5.6 supports a northward downbending the Adriatic Moho and lower crust.

Therefore assuming a pre-Alpine age for the origin of the reflections from the Moho and the lower crust and a meso- or neo-Alpine age for the origin of the reflections from the middle and upper crust the following large-scale tectonic model is quite naturally
advocated by the combined depth-migrated NFP20 eastern and southern traverses shown in Figure 5.6:

With the onset of continental collision in late Eocene times subduction of European lower crust continued - possibly due to a self-sustaining driving mechanism such as the one suggested by Fleitout and Froidevaux (1982) - though at a rather low rate (<0.5 cm/y for an estimated neo-Alpine shortening of 100 to 200 km (Trümpy 1980; Butler 1986; Coward and Dietrich 1989)). The fact that the crustal thickness of the European and Adriatic continental margins had been previously dramatically reduced as a result of Triassic and Jurassic rifting may have facilitated early crustal shortening and subduction. The meeting of the two "grown-up" crusts in the Oligocene (cf. Trümpy 1980) led to a space problem which in turn resulted in a downbending of the Adriatic Moho and lower crust, wedging at mid-crustal levels (Mueller et al. 1980; Kahle et al. 1980; Mueller 1990), backthrusting and vertical movements at upper crustal levels (Heitzmann 1987; Pfiffner et al. 1990) as well as in the uplift of the external massifs and the progression of nappe formation and subhorizontal thrusting into the Alpine forelands (Trümpy 1980; Coward and Dietrich 1989). The downbending of the Adriatic Moho and lower crust suggested by Mueller et al. (1980) and by Figure 5.6 can in fact account for most of the interpreted late neo-Alpine shortening of the upper crust in the Southern Alps (Laubscher 1985, 1989; Roeder 1989). However, the simultaneous bivergent near-vertical subduction or "Verschluckung" of both the European and the African lithospheres as postulated by Laubscher (1971) cannot be substantiated on the basis of the available seismic reflection data. Wedging at mid-crustal levels has been previously inferred by Kahle et al. (1980) and Kissling (1980, 1982) from the presence of a distinct local gravity high in the Ticino and by Mueller et al. (1980) based on the detailed interpretation of seismic wide-angle data. In the low-pass filtered reinterpretation of some of these profiles used for the purposes of this work (cf. Figures 4.1 to 4.8) the thickening of the slow upper crust relative to the fast lower crust (see Figures 5.3, 5.5 and 5.6) in the axial zone of the Alps may be considered as an indication of mid-crustal wedging in conjunction with lower crustal subduction (cf. Mueller 1990). The concentration of upper crustal vertical movements in the contact zone of the European and Adriatic plates resulted in the development of the highly reflective mylonite zone of the Insubric Line (Fountain et al. 1984; Schmid et al. 1987, 1989). The fact that after migration the reflections from the Insubric Line are confined to the upper crust, i.e. 15 to 20 km depth, (Figure 5.6) may be taken as further indirect evidence for subhorizontal wedging in the middle crust.
The topography of the refraction and reflection Mohos suggests that at least the recent parts of the neo-Alpine subduction of the European lower crust and uppermost mantle have occurred at a low angle of 15 to 20 degrees (Figure 5.6). Since the lower crust cannot be expected to have been coupled with the tectonic evolution of the middle and upper crust and since the original positions of the observed Moho reflections are unknown the total amount of this low-angle subduction can hardly be quantified.

The above tectonic of course model just represents one possible explanation of the reflectivity distribution within the Alpine crust shown in Figure 5.6. And again one has to bear in mind that this image is nothing but an acoustic snapshot of the present-day physical state of the crust below the central Alps, in which certain important tectonic features may have found no expression at all whereas some other rather unimportant aspects may well be over-emphasized. However, with all its crudeness and limitations I do consider the above interpretation to be an attractive and appealing one because it is in good qualitative agreement with the relatively well-established post-Eocene tectonic evolution of the Alps and the mass balance considerations derived therefrom as well as with the most pertinent complementary geophysical information such as the pronounced asymmetry of the Moho "trough", the low average crustal velocity in the Penninic domain (Mueller et al. 1980), the local gravity high in the Ticino (Kahle et al. 1980; Kissling 1980, 1982) and the presence of a lithospheric root required by gravity (Kissling 1980, 1982; Kissling et al. 1983; Schwendener 1984; Schwendener and Mueller 1990), surface wave dispersion analysis (Panza and Mueller 1979; Panza et al. 1980), teleseismic delay times (Baer 1979, 1980), teleseismic tomography (Spakman 1990) and the elevated velocities in the upper mantle beneath the ALP75 seismic wide-angle profile (Yan and Mechie 1989).

The geometric inconsistency between the gently south-dipping subduction of the European lower crust and uppermost mantle shown in Figure 5.6 and the near-vertical orientation of the lithospheric root inferred from the analysis of surface wave dispersion (Panza and Mueller 1979; Panza et al. 1980) and teleseismic P-wave delay times (Baer 1979, 1980) is only a crude approximation of the actual situation. In this context three important arguments have to be noted. First, the acoustic image derived in the course of this work (Figure 5.6) reaches to a depth of only some 60 km and necessarily leaves everything below open to speculation; second, the geophysical methods (surface wave dispersion, teleseismic P-wave delays) used to interpret the near-vertical orientation of the lithospheric slab at greater depth have a rather low resolution even on a lithospheric scale;
third, post-Eocene Alpine tectonics can be as readily explained by the gently south-dipping subduction of the European plate (cf. Butler 1986) as with the simultaneous bivergent near-vertical subduction of both the European and the Adriatic plates (cf. Laubscher 1970, 1971).
6. DISCUSSION

As pointed out by Klemperer (1989) and confirmed by the 1990 CCSS (Commission on Controlled Source Seismology in the International Association of Seismology and Physics of the Earth's Interior; see e.g. Ansorge et al. 1982) workshop at Fellhorst, Germany, the conventional processing of deep seismic reflection data including stack and post-stack S/N enhancement has reached a high degree of maturity and standardization. This is not true for migration. Although it is commonly acknowledged that all deep seismic data ought to be migrated in order to yield a less distorted acoustic image of the subsurface it is often done either in a lukewarm way or not at all. In my opinion the reason for this stepmotherly behaviour towards migration is to be attributed to the fact that "tinmed" seismic processing software works well as far as S/N enhancement and CMP-stacking of deep data are concerned but standard migration algorithms are generally ill-behaved.

In the methodological part of this work (chapters 2 and 3) I have shown how important the migration of deep seismic reflection data is and how sensitive this process is to an accurate parametrisation of the velocity field. Short, laterally discontinuous reflection segments and high ambient noise levels characteristic of deep seismic reflection data represent fundamental violations of the boundary/initial value problem of the scalar wave equation upon which conventional migration is based. As a consequence the more accurate an algorithm models the scalar wave equation the more disastrous are the effects of these data insufficiencies. Attempts to overcome at least the noise problem by harsh enhancement of coherent energy either before (e.g. Valasek et al. 1990) or during (Milkereit 1987) migration have led to certain improvements. However, the sensitive parametrisation of the coherency filter and the gain function results in the loss of much of the original amplitude and phase information and makes this approach prone to the introduction of artefacts. Moreover, the result still does have a "smiley", overmigrated look due to the lateral discontinuity of the reflection segments and "lacking" diffractions at their ends (cf. Warner 1987).

I therefore argue that in the case of deep seismic reflection data ray-theoretical migration of the observed travel times, i.e. of a digitised line drawing of the observed primary reflections and diffractions, represents the vital alternative to migrating the "full" observed wave field using algorithms based on the scalar wave equation. Ray theory represents a crude first-order approximation to the solutions of the wave equation for
high frequencies and for this reason does not suffer from the data insufficiencies hampering wave equation migration. Ray-theoretical migration is geometrically accurate and numerically efficient, can handle very complex velocity structures and is able to account for geometric focusing and defocusing effects; as a high-frequency approximation it does, however, assume perfect resolution of the stacked input data and therefore, unlike wave equation migration, does not increase the spatial resolution of the migrated section by collapsing the Fresnel zone. I think that the major objection against ray-theoretical migration which concerns the inherent subjectivity of the hand-made line drawings and their lack of amplitude and phase information is largely invalidated by the fact that - as mentioned above - computer-generated line drawings used to trick conventional algorithms (cf. Valasek et al. 1990; Milkereit 1987) are subjective as well and are unlikely to contain much original amplitude and phase information. Finally, one must not forget that until the early 1970s virtually all the oil discoveries - amongst them some of the world's largest hydrocarbon reserves - were made based on ray migration of "picked" time sections and that ray theoretical depth migration of interpreted time horizons is still standardly used by the oil industry (Sattlegger 1982).

I have shown that at the travel times and velocities relevant for deep seismic reflection data the parametrisation of the velocity field is of considerable importance. If one assumes that the observed reflectivity on normal-incidence seismic data represents random deviations from a so-called macro-velocity structure (e.g. determined by the most prominent wide-angle reflections) then it is sensible to separate the large-scale components of the velocity field from its small-scale components for the purpose of migration. A smoothed velocity structure such as obtained by the minimum interpretation of seismic wide-angle data may thus be considered not only as a sufficient but as an ideal migration velocity model. This approach assumes that wave propagation is governed by the large-scale velocity structure of the medium and is independent of reflectivity and, therefore, it is strictly valid only if first-order discontinuities of the large-scale velocity structure do not coincide with continuous reflected phases in the seismic reflection data. Unlike reflectors from within the sedimentary cover individual deep crustal reflectors are rarely laterally continuous and therefore cannot be expected to show a one-to-one correlation with the discontinuities of the laterally smoothed large-scale velocity field. On the other hand changes in crustal reflectivity, such as the "reflection Moho" (cf. Klemperer et al. 1986), may well correspond to changes in lithology and as such to first- or second-order velocity discontinuities. Therefore the convergence of changes in
reflectivity with the contours of the macro-velocity structure after migration is one possible indication of the correctness of the used migration velocity model.

The experience gained in the course of this work allows some comments on the peculiarities of the acquisition of the NFP20 Alpine reflection traverses (cf. Valasek et al. 1990). The strategy to achieve, wherever possible, a redundant coverage of highly resolving Vibroseis data and strong single-fold dynamite data was proven to be successful. A crustal study at this scale would not have been possible on the basis of one dataset alone: e.g. only a fraction of the upper crustal Penninic and Helvetic reflectivity is imaged by the dynamite data whilst the Vibroseis data do in general not provide reliable information at lower crustal levels (Pfiffner et al. 1988; Pfiffner et al. 1990; Valasek et al. in press). While combining the two datasets during line drawing preparation, i.e. Vibroseis data for the upper and middle crust and dynamite data for the middle and lower crust, I generally found - despite the crooked line geometry and the relatively large offsets of the dynamite data - a one-to-one correlation of reflectors that were imaged by both datasets. The most likely explanation for this phenomenon is a considerable lateral continuity of the most prominent reflectors as well as the robustness of the NMO-correction at mid and lower crustal levels (at 10 s TWT and a maximum offset of 10 km using a NMO-velocity of 6.0 km/s instead of one of 6.5 km/s results in a maximum travel time difference of only 0.02 s, which is barely recognizable).

For the purpose of a unifying study the fact that unlike the ECORS Pyrenees line (ECORS Pyrenees Team 1988; Choukroune and ECORS Team 1989) and the ECORS-CROP line across the western Alps (Bayer et al. 1987; Nicolas et al. 1990) the NFP20 Alpine traverses had to be acquired in the form of several individual segments (cf. Frei et al. 1989; Valasek et al. 1990, in press) is unfortunate. In order to gain a unified acoustic image across the central Alpine arc covered by the NFP20 traverses the individual profiles have to be combined. With the strong lateral variations of surface geology in the Penninic domain any such combination is bound to be a subjective and in certain aspects an insufficient compromise for the upper crust. The approach taken in this work aimed at the consistency of complementary geophysical data on a crustal scale as well as on the actual topography. Hence projecting horizontally along the relatively smooth east-west trend of the "Ivrea-corrected" Bouguer gravity isolines (Kissling 1980, 1984) is considered to be sensible for the middle and lower crust. Since upper crustal effects are not adequately represented in this trend of the corrected Bouguer gravity anomalies this approach leads to inconsistencies in the upper crustal reflectivity patterns and their
geological interpretation: although it is not entirely clear how to correlate the Penninic gneiss nappes of the Ticino with those of the Grisons (cf. Trümpy 1980), the base of the upper crustal reflectivity on line S1 at 2 to 3 s TWT is unlikely to correspond to the base of the Tambo nappe on line ET (cf. Pfiffner et al. 1990) as suggested by this projection technique.

Whilst in areas of mild tectonic variations coincident parallel seismic reflection and refraction profiles (Mooney and Brocher 1987) may be an ideal combination, in more complicated environments seismic wide-angle profiles ought to be oriented parallel to the structural trend in order to reduce the negative effects of spatial aliasing. On the other hand seismic reflection profiles should always be oriented perpendicular to the tectonic trend in order to permit an unbiased and complete sampling of a 2-dimensional cross section through the actual 3-dimensional wave field. Therefore the availability of relatively densely-spaced (approximately 20 km on average) seismic wide-angle profiles oriented parallel to the structural grain of the central Alpine arc (cf. Figure 4.1), which is the result of long-term careful scientific planning and coordination at an international level (cf. Mueller et al. 1980 and references therein), to my knowledge represents an absolutely unique situation.

The migration velocity model for the combined NFP20 deep seismic reflection profiles was derived from reinterpreting these profiles in a low-pass filtered way (cf. Figure 5.3). Since such an approach strongly emphasizes lateral continuity locally clear but laterally discontinuous wave groups had to be discarded in some places whereas in other places weak but otherwise laterally continuous wave groups were considered. By no means is the resulting velocity model (cf. Figure 5.3) intended to be an alternative or even an improvement on any previous or future detailed velocity model of the crustal structure. Its only purpose and justification is to provide a macro-model that allows accurate migration over the entire length of the combined NFP20 eastern and southern traverses.

Since the interfaces of the macro-model are not horizontal, the reflection points are not located vertically beneath the refraction profiles and consequently have to be migrated as well before serving as input for the migration of the digitised line drawings. This was achieved by converting the velocity contours into two-way travel time and then migrating them in the same way as the reflection data. Assuming a pre-Alpine age of origin for the reflections from the lower crust and the Moho and meso- to neo-Alpine ages of origin for the reflections from the middle and upper crust the resulting acoustic image (cf. Figure
is interpreted as originating from the post-Eocene low-angle subduction of the European lower crust and topmost mantle below the Adriatic promontory of the African plate. This is interpreted to have resulted in the northward downbending of Adriatic lower crust and Moho (cf. Valasek et al. in press), wedging at mid-crustal levels (cf. Kahle et al. 1980; Mueller et al. 1980; Mueller 1990; ETH Working Group 1990) and sub-horizontal nappe emplacements as well as vertical escape movements at upper crustal levels (cf. Trümpy 1980; Coward and Dietrich 1989; Pfiffner et al. 1990). This tectonic concept is schematically illustrated in Figure 6.1. Similar - albeit not identical - interpretations have been proposed for the Pyrenees profile (Choukroune and ECORS Team 1989) and the deep seismic reflection profile in the western Alps (Nicolas et al. 1990).

Figure 6.1. Cartoon illustrating the present-day lithospheric framework of the central Alps as interpreted by Pfiffner (1990) from a precursor of Figure 5.6.
The gently south-dipping subduction zone inferred from the depth-migrated reflection and refraction data shown in Figure 5.6 implies substantial differential motions between the base of the crust and the base of the lithosphere, at present not included in the most prominent tectonic (Laubscher 1970, 1971) and geophysical (Panza and Mueller 1979; Panza et al. 1980) models, both of which favour a near-vertical "Verschluckung". In addition, Figure 5.6 places clear constraints on the crustal geometry of those models and implies a very narrow zone beneath the Alps in which any possible near-vertical downdipping of the crust could be accommodated. Evidently the proposed large-scale tectonic model does not require Butler's (1986) eclogite transition below 50 km to make the refraction Moho continuous across the Alpine arc. But what about the buoyancy of the subducted crustal material? Now that we have a better understanding of the shallow geometry of the Alpine subduction zone it should be subject to future gravimetric studies to put constraints on how much crustal material in which density range can be subducted. The additional a priori information on the geometry of the deeper parts of the subducted lithosphere and on the lithosphere/asthenosphere density contrast required for such a gravimetric study may in turn be provided by detailed teleseismic imaging of the Alpine system. Based on the results of this work I am of the opinion that thoroughly planned and coordinated future research concentrating on the geometry of the Alpine subduction zone as well as on the mechanisms of crustal shortening and subduction will be scientifically highly rewarding. A potential sequence of research topics might be:

- Local tomographic studies - possibly involving active sources - to derive detailed images from the middle and upper crust (cf. Kissling 1988; Kradolfer 1989).
- Detailed teleseismic tomography (cf. Dziewonski and Anderson 1984) to constrain the geometry of the subducted lithosphere at depths not accessed by this study, i.e. below 60 km.

One of the most fiercely debated topics ever since the NFP20 data became available was the apparent discrepancy between the refraction Moho, which was interpreted to form a continuous asymmetric "trough" across the Alpine arc (Mueller et al. 1976; 1980), and the reflection Moho which evidently is discontinuous. There are two different aspects to this discontinuity of the reflection Moho: first, the rather abrupt loss of the reflection Moho south of the Aar Massif in the Penninic domain (approximately from 35 to 65 km distance on migrated combined data shown in Figure 5.6); second, the reappearance of
the reflection Moho just south of the Insubric Line at a substantially greater depth than the one inferred for the Adriatic refraction Moho from seismic wide-angle data (cf. Figures 5.5 and 5.6).

The simultaneous migration of the interfaces interpreted from the seismic refraction data and the line drawings of the seismic reflection data has removed this contradiction in that there is no longer any solid evidence upon which to postulate a continuous refraction Moho. On the contrary, both the reflection and the refraction Mohos, where defined, now nicely depict the frame of a gently (10 to 20 degrees) south-dipping subduction of the uppermost mantle and parts of the lower crust of the European plate beneath the Adriatic promontory of the African plate. This in turn suggests that the discontinuity of the reflection Moho below the Penninic nappes of the Grisons most likely represents a seismic imaging problem. Reflections from the Moho and the lower crust are also lacking in the axial parts of the western Alps profile (Nicolas et al. 1990) and are strongly "washed out" on the Pyrenees profile (Choukroune and ECORS Team 1989). From all these "transparent" areas doubtless wide-angle reflections from the Moho have been reported (cf. Figures 5.5 and 5.6; ECORS-CROP Deep Seismic Sounding Group 1989; Daignières et al. 1982).

In the case of the NFP20 data the fact that there are wide-angle reflections but no normal-incidence reflections from the Moho may be attributed to insufficient source energy which could penetrate the deep parts of the thickened orogenic crust, defocusing of the reflected energy out of the used receiver spread (Valasek et al. 1990) or destruction of the reflectors by intense deformation (Pfiffner et al. 1990). The energy argument seems unlikely because similar or weaker charges have imaged the Moho at substantially greater depth further to the south. Furthermore, the excellent imaging of the upper 5 to 10 s TWT (cf. Pfiffner et al. 1990) indicates that the S/N is not worse than elsewhere along the traverse. Defocusing effects seem unlikely because the slope of the Moho constrained by seismic wide-angle reflections (Figure 5.6) does not change in the Penninic domain (Valasek et al. in press). The deformation argument is unlikely because then no Moho reflections would be expected further south where the deformation is still more intense. It is noteworthy that the loss of the Moho reflections coincides with the onset of upper crustal reflectivity in the Penninic domain and the reappearance of the Moho reflections with the disappearance of the Penninic reflectivity (Figures 5.5 and 5.6). Hill and Levander (1984) and Carbonell and Smithson (1990) have shown by finite difference modelling of the full wave field that the imaging of highly reflective zones at depth may
not be possible if the wave front is broken up by overlying scatterers. Such scattering phenomena represent source-generated noise and hence for the same frequency spectrum become more severe with increasing source strength. If a broken-up wave front is "reflected" from an interface the recorded seismic section does not show any lateral coherence in terms of phases but only in terms of wave groups, i.e. energy accumulations. Therefore the presence of the reflecting interface is only detectable on spatially aliased seismic wide-angle data but not on normal-incidence seismic reflection data. Scattering due to the complex upper crustal geology in the Penninic domain of the Alps and the axial zone of the Pyrenees therefore represents an plausible qualitative explanation for the apparent discontinuity of the underlying reflection Mohos.

As illustrated in Figure 2.1 seismic migration and modelling are closely related; in fact they both represent solutions of the wave equation for different boundary/initial conditions: whilst migration tries to unravel the reflectivity from observed seismic data, modelling produces seismic data from the given reflectivity. Migration and modelling should therefore be treated as kind of a positive iterative feedback process to reveal the distribution of reflectivity in the imaged subsurface and at the same time to put constraints on its physical origin. In this study I have tried to work out the problems and limitations of migration of deep seismic reflection data, and the results suggest that the shortcomings of migration of deep seismic reflection data can largely be overcome by simultaneous forward modelling: whereas ray migration of deep seismic reflection data allows us to reconstruct parts of the real wave field and thus gives us an idea of the spatial distribution of reflectors in the subsurface, modelling allows us to construct the corresponding full - albeit synthetic - wave field which may then be compared with the original seismic data e.g. in terms of their phase and amplitude attributes. In the future, therefore, I think that substantial contributions to our understanding of wave propagation in general and of the detailed physical nature of the crust and the upper mantle in particular are to be achieved by simultaneous iterative migration and modelling of deep seismic reflection data.
7. REFERENCES


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CURRICULUM VITAE OF KLAUS HOLLIGER

June 2, 1962: Born in Aarau, Switzerland, as the first of two sons of the theatre director Erich Holliger and the primary school teacher Ruth Zulauf

1969 to 1973: Primary school in Langenthal, Switzerland
1973 to 1978: Secondary school in Langenthal, Switzerland
1978 to 1981: Grammar school in Langenthal, Switzerland
September 1981: Maturity exam type B (classical/linguistic direction)
October 1982 to April 1987: Diploma in natural sciences at the Swiss Federal Institute of Technology (ETH) in Zürich, Switzerland; diploma with distinction and ETH medal for diploma thesis (Title: "Crustal structure of the North Sea based on the analysis of SEASAT altimetry, deep seismic reflection data and shipborne Bouguer gravity anomalies")

July to October 1984: Work in a seismic field crew in Yugoslavia
April 1985 to October 1986: Part time research assistant with the Petrological Institute of ETH Zürich

October 1986 to April 1987: Exchange student at Imperial College in London, U.K., and visiting scientist at the University of Cambridge, U.K.

May to December 1987: Work at the seismic processing centre of ETH Zürich
April to August 1988: Visiting fellow of the British Natural Environmental Research Council (NERC) at the British Institutions' Reflection Profiling Syndicate (BIRPS) at the University of Cambridge, U.K.

Since July 1989: Teaching assistant in the Department of Environmental Sciences of ETH Zürich

January 1988 to December 1990: PhD student at the Institute of Geophysics of ETH Zürich
APPENDIX A: Description of MIGRAY

MIGRAY is a modular FORTRAN77 program to perform ray theoretical depth migration of digitised line drawings of seismic reflection profiles through a two dimensional velocity structure. The velocities within the individual layers are defined to be constant. The theoretical justification of the method, the algorithm and the original version of the program were developed by Bernard Raynaud while studying for a PhD at Cambridge University (Raynaud 1988a,b). Subsequent verifiable modifications were made by Dave Smythe (Glasgow) and Richard Hobbs (Cambridge). Rather significant modifications of the graphics part led to the current interactive VAX VMS "Zürich version" of the program which can now be linked to TERMPILTLIB, an extended CALCOMP online library (cf. description by Edi Kissling).

The program consists of three major modules: MAIN, ANALYS and GRAPH. MAIN organises the large scale process flow; ANALYS performs essentially all the calculation work; GRAPH handles the graphic display of the results of a previous run of ANALYS (see Figures A.1 and A.2). Currently there are three versions of the program:

MIGRAY
- allows you to plot interactively on a Tektronix compatible screen
- linked to TERMPILTLIB (cf. description by Edi Kissling)
- installed on AEOLUS, JANUS and REFLEX

MIGRAYHP
- allows you to make A3 size hardcopies on the HP7550 pen plotter
- linked to NHPLIB (cf. description by Nicolas Sierra)
- installed on AEOLUS and JANUS

MIGRAYVT
- allows you to make hardcopies of "arbitrary size" on the 36 inch Versatec electrostatic plotter
- linked to the SSL graphics library (cf. description by SSL); creates a *.gfx file which can be rasterised subsequently
- installed on REFLEX only
Apart from lacking the CALCOMP routines CHANGEPLANE and CLEAR the modular structure of MIGRAYHP and MIGRAYVT is virtually identical to that of MIGRAY.

**Figure A.1.** Flow of control in MIGRAY.
Figure A.2. Large scale structure of MIGRAY.
MODULES
The structure of the program at a modular level and the flow of control is illustrated in Figures A.2, A.3 and A.4.

MAIN
Opens and closes the data files, initialises and deactivates the plot software and directs the process flow to either run ANALYS, GRAPH or exit the program. After each run of ANALYS or GRAPH control always returns to MAIN.

OPENFL
Opens units 8 to 11, 13 to 15 and 17 for writing (analysis run) or reading (graphics run).

CLOSFL
Closes units 8-16 opened either for analysis or graphics. Remember: units 8 to 16 were opened for analysis; only units 8 to 11, 13 to 15 were opened for graphics.

OFIL
Opens units 12 (*.dig) and 16 (*.gst) for an analysis run. Not used in a graphics run.

ANALYS
Directs the process flow for an analysis run. Interactive input: x- and t-window, pretty factor, angle between line and dip, 1- or 2-D structure.

STRUCT
Reads in the parameter file (*.gst, unit 16).

INREAD
Auxiliary subroutine to read in data that are supplied at ten values per line.

CHUCK
Auxiliary subroutine to produce output with ten values per line.
PICVEL  
Reads in the velocities and time or depth values from unit 17 provided in a "velan" (velocity analysis) format and constructs isovelocity contours. Velocity inversions are not allowed.

RINTER  
Linear interpolation function.

RDVEL  
In this case the velocity field is defined in an isovelocity layer format. Reads in times or depth to the base of each layer. Velocity inversions are allowed.

CONTRS  
Outputs the velocity contours in depth used by the program.

EXTEND  
Adds one more extra column onto either side of the model in order to prevent ray loss.

MIGRAT  
Reads in digitised line segments from unit 12 (*.dig) and performs depth migration.

MIGSTR  
Converts structural contours into a form in which they resemble the segments of a line drawing in preparation for migrating the structure.

PRESEG  
Produces a single smoothed segment of the line drawing.

DISCRD  
Throws away a point that is out of range and reads the next one from unit 17 (*.dig).

DPOINT  
Reads in a line segment, cuts it into smaller pieces if its curvature exceeds a certain threshold and smooths each section.
SMTH3
Three point smoothing of line segments on both x- and t-values by applying weights of 0.25 0.5 0.25 and summing.

SMOOTH
Performs triangular smoothing.

WEIGHT
Auxiliary weighting function for SMOOTH.

QUAD1 and QUAD2
Auxiliary routines for quadratic smoothing of line segment ends.

PRETTY
Reduces the curvature of the line segments in order to suppress "smiling" effects in the migrated output section.

APPOINT
Reads in the corners of the digitised section.

PRERAY
Starts off a new ray which will migrate one line element.

STEP
Organises the tracing of one ray through the model.

SEARCH
Searches for the next velocity column to the left.

PLANMT
Determines the length of the ray in a particular layer.
REFRACT
Finds the refracted or reflected wave vector of the ray as it enters a new zone of constant velocity.

DOT
Calculates the dot (scalar) product of two vectors.

NORM
Normalises a vector.

CHECK
Prints out some data for user checking.

ENDRAY
Finishes a ray and outputs data and messages.

WINDUP
Puts end markers on the output files when the data are all read. After each call of
WINDUP control returns to MAIN.

GRAPH
Organises the flow of control for a graphics run.

RDSTR
Reads in the structure data from unit 14 (*.cont).

CHosen
Statement of options chosen for plotting.

OPTION
Offers menu for graphic display of the results. After each plot control always returns to
here via GRAPH.
READIN
Reads in the velocity contours data from unit 10.

CHANGE
Allows the following changes to the graphic display of the results:
• Squash factor (>1 squeezes the plot in the x-direction)
• Change the x-range of the viewing window
• Smile filter on or off
• Steepest segment to be drawn, minimum amplitude to plot, minimum amplitude for emboldening

PLOTSETUP
Gets everything ready for plotting. The scaling factor is determined so that the plot will fill the screen.

PLTNOW
Calls PLOTSETUP and the plotting routines corresponding the choice made in OPTION.

NXTPLT
Prepares the next plot. Flushes the plot buffer, restores alphanumeric mode and determines whether the new plot is to overlay the old one or the graphics plane is to be cleared.

AXES
Draws a frame around the section and labels it.

AXE
Draws an axis an labels it.

DRAWT
Draws a raw or smoothed unmigrated time section.

DRAWZ
Draws a migrated depth section.
DRAW
Scales a given number of coordinates and draws a line through them.

PLTSTR
Plots the velocity-depth structure.

PLTRAY
Plots the normal incidence rays used for migration in the requested density. A ray density of 0 will plot no rays at all, one of 1 will plot all the rays, one of 10 will plot every tenth ray.

PROFIL
Draws scaled velocity profiles in either time or depth.

XPLOT and ZPLOT
Auxiliary functions to transform the original x (km) and z (km or secs) values into plot coordinates.
Figure A.3. Modular structure of the analysis part of MIGRAY.
Figure A.4. Modular structure of the graphics part of MIGRAY.
INPUT
Interactive Input
See "User's Guide to MIGRAY".

Noninteractive Input
Input to MIGRAY are a parameter or "ghost" file (*.gst, unit 16) and a file containing the digitised line drawing (*.dig, unit 12). For more details see "User's Guide to MIGRAY".

*.GST FILE (UNIT 16)
This file contains all the crucial information for the definition of the velocity field used for migration. The comment line which precedes each data field makes its handling largely self-explaining. The data are given in free format. This file is read by STRUCT and INREAD. For more details see "User's Guide to MIGRAY".

*.DIG FILE (UNIT 12)
This file contains the coordinates of the digitised line segments preceded by those of the corners and by a title. Title, corners and individual line segments are separated by flags specified in the second data field of *.GST. The current format of both coordinates and flags is 2I7. The data are read by MIGRAT, PRESEG, DSCRD, APOINT, CORNERS, DPOINT.

OUTPUT
Interactive Output
This essentially consists of the graphic display of unmigrated time section, migrated depth section, velocity structure, ray paths and velocity profiles in time or depth.

Noninteractive Output
This consists of eight files (units 8-11,13-15,17) which are briefly described below:

*.STR (UNIT 8)
This is essentially a copy of *.GST.
*.RAW (UNIT 9)
Contains the coordinates and the corresponding amplitude attributes of the original
digitised line segments. The amplitude attribute is always 1.0 since no migration and
hence no amplitude changes due to geometric focusing or defocusing have occurred yet.

*.CONT (UNIT 10)
Last two data fields of *.GST: type of velocity data and depths to base of layers.

*.RAY (UNIT 11)
Each column contains top to bottom: x-, z-, t-coordinates of a particular ray.

*.ZSE (UNIT 13)
Coordinates and amplitudes of line segments after depth migration.

*.COR (UNIT 14)
Corners in km and secs of the data window, angle between line and dip.

*.SMT (UNIT 15)
Coordinates of smoothed line segments (vertical axis is time), and corresponding
amplitude attributes (always 1 since the data are unmigrated).

*.PRI (UNIT 17)
Run time documentation.

COMMON BLOCKS
The overwhelming part of the parameter transfer between the individual modules is
organised by COMMON blocks. This may be a practicable solution for the first moment
man, it does, however, anything but help the understanding of what is happening where
in the program for latecomers like me. Based on this experience a detailed description of
all the COMMON blocks is given here with the advice to avoid them as much as possible
whenever you start from scratch with a new program.
COMMON Blocks used in an Analysis Run

COMMON / IOTERM /
SCRN  unit 6 (defined in MAIN for interactive output to the screen)
KBRD  unit 5 (defined in MAIN for interactive input from the keyboard)

COMMON / IOFILE /
I     analysis(1), graphics(2) or stop(0)?

COMMON / DIMENS /
NLAY number of layers (*.gst)
NCOL number of columns (*.gst)
TMAX maximum time to be considered (*.gst)
ZMAX maximum depth to be considered (*.gst)
VMAX velocity below bottom layer (*.gst)
SMTH length of horizontal smoothing (*.gst)
SIDIP sinus of maximum interface dip (*.gst)
ILOOK true 2-D structure? (interactive input)
DIPCPT cosinus of angle between line and dip (interactive input)

COMMON / CTRL /
IRAY
IRAYL
IRAYH (first data field of *.gsf)
ISTOP
IBUG
IFTIME velocity in time or depth? (*.gst)
IFPROF "velans" or layers of constant velocity? (*.gst)

COMMON / STUFF /
IX x-coordinate of digitised line segment (*.dig)
IY y-coordinate of digitised line segment (*.dig)
XSC x-scale of digitised section (*.gst)
TSC time scale of digitised section (*.gst)
W segment width (*.gst)
ANG maximum bending of segments in degrees (*.gst)
VLMIN minimum segment length (*.gst)
VLMAX minimum segment length (*.gst)
PRET smile reduction factor (interactive input)

COMMON / FLAGS / (second data field of *.gst)
IXEN end flag
IXBR break flag
IXCO corner flag
IXRF reflection flag
IXDOT dotted flag

COMMON / EDGES / (attention these parameters may be redefined during a graphics run!)
XL & XLW left corner of viewing window (as defined by interactive input)
XR & XRW right corner of viewing window (as defined by interactive input)
ZT & ZTW top depth window (as defined by interactive input)
ZB & ZBW bottom of depth window (as defined by interactive input)

COMMON / VELMOD / (read from *.gst)
VEL(20) velocity values (*.gst)
XCOL(50) x-values of velocity columns (*.gst)
ZED(50,20) z-values of velocity columns (*.gst)

COMMON / LINSEG / coordinates of a particular line segment before migration, i.e. as read in from *. dig
X0(500), Z0 (500): smoothed
X(500), Z(500): unsmoothed.

COMMON / RESULT / coordinates after depth migration
XMIG(1000)
ZMIG(1000)
COMMON / RAYS /
XLOOK x-coordinate to start off a ray
XC(4) current x-,y-,z-,t-values
XP(4) previous x-,y-,z-,t-values
RK(3) wave vector components

COMMON / PATH / ray coordinates
TM(21)
XM(21)
ZM(21)

COMMON Blocks used in a Graphics Run

COMMON / IOTERM /
SCRN unit 6 (defined in MAIN for interactive output to the screen)
KBRD unit 5 (defined in MAIN for interactive input from the keyboard)

COMMON / TITLES /
ITITLE read in from unit 14 (*.cor); corresponds to the group path name which is interactively entered

COMMON / XNRX /
NRAY ray density
IRAY first parameter of first data line of *.gst

COMMON / CORNS /
XEN end flag (*.gst)
XBR break flag (*.gst)
XCO corner flag (*.gst)
XRF reflection flag (*.gst)
XDOT dotted flag (*.gst)
X1, X2 original x-window
XL, XR extended x-window used for migration
ZBT bottom of time window
ZBZ: bottom of z-window (km)
ZT: top of z-window (km or secs)
ZB: bottom of depth window (km or secs)

COMMON / PSEUDO / (interactive input)
- AMPMIN: minimum amplitude to draw
- AMPBLD: minimum amplitude for emboldening
- DEGMAX: maximum dip of a line segment to consider
- DIPMAX: tangens of DEGMAX
- ISMI: smile filter off (0) or on (1)?
- XPSL; XPSR: new window limits assigned in CHANGE
- SQUASH: squash factor

COMMON / GRAPHS /
- NLAY: number of layers (*.gst)
- NCOL: number of columns (*.gst)
- DIPC: angle between line and dip in degrees (interactive input)
- ZMAX: maximum model depth (km or secs) (*.gst)
- VMAX: velocity below bottom layer (*.gst)

COMMON / FRAME /
- XLW, XRW, ZTW, ZBW: limits of data window considered

COMMON / TCURTS /
- VEL(20): velocity values (*.gst)
- XCOL(50): x-values of velocity columns (*.gst)
- ZED(50,20): z-values of velocity columns (*.gst)
COMMON / SCAL /

NX, NZ  number of intervals
DX, DZ  inches/10 km or inches/sec
X0W, Z0W values to label origin with
DXW;DZW increment of these values
DXSCAL  inches/sec
DZSCAL  inches/km
MIGRAY is an interactive program to migrate digitised line drawings of reflection seismic data through a 2-D velocity structure. It consists of two major collections of modules to perform the calculation work related to migration (analysis part) and to handle the graphic display of the results (graphics part). Each run of the graphics part must be preceded by at least one analysis run for a particular data set. Currently there are three versions of the program:

**MIGRAY**
- allows you to plot interactively on a Tektronix compatible screen
- installed on JANUS and REFLEX

**MIGRAYHP**
- allows you to make A3 size hardcopies on the HP7550 pen plotter
- installed on JANUS only

**MIGRAYVT**
- allows you to make hardcopies of "arbitrary size" on the 36 inch Versatec electrostatic plotter
- installed on REFLEX only

The non-interactive input to MIGRAYHP and MIGRAYVT is identical to that of MIGRAY.

**MIGRAY NON-INTERACTIVE INPUT TO MIGRAY**
Although the program is largely interactive, the two input files containing the model parameters and the digitised line elements cannot be accessed interactively at present.
*.gst ("Ghost") file
This file contains all the information on the velocity structure to be used for migration. Since each data field is preceded by an instructive comment line no further information is given here (Figure B.1). The data can all be supplied in free format.

```
* Print one ray in ---- from ---- to ---- and end at segment -----. Debug? (1/0)
  4 1 1000 10000 0
Flags: end, break, corner, reflection, dotted
-999999 -666666 -777777 -444444 -888888
xscale and tscale (1/1000N where N is no. of cm/km or cm/s)
  0.000333 0.000178
For segments width, min length, max length(all in km), degrees bending
  0.4  0.05  0.7  20.0
Now supply the structure data
  First the number of constant velocity bands (<60)
    4
  Now the velocities (in km/sec : 10 per line)
    2.1  5.2  4.3  5.5
  The number of columns of vel/depth to be given
    5
  the x- coordinates of the columns (in km)
    0.  4.7  4.8  5.1  7.
  Maximum time (s) & depth (km) to be considered & velocity under lowest layer
    3.  5.  8.
  Length of horizontal smoothing ? maximum interface dip ? true 2-dim model ?
    0.0  90  1
  Are the velocity profiles in time (0) or depth (1)? type of data? "velans"(0) or constant velocity layers(1)
    0  1
  Depths to base of interval velocities
    0.74  1.90  1.91  1.83  1.36
    0.74  1.90  1.91  1.99  2.00
    1.94  1.99  1.99  1.99  2.00
    2.29  2.78  2.77  2.70  2.41
```

Figure B.1. Example of a "ghost" file.

*.dig file
This file contains the coordinates of the digitised line segments preceded by those of the corners and by a title (Figure B.2). Title, corners and individual line segments are separated by flags specified in the second data field of *.gst. The origin is set at the top left corner, corners are digitised clockwise starting at the bottom left corner, line segments from left to right. The current format of both coordinates is 217, that of the flags 216.
INTERACTIVE INPUT TO MIGRAY

In the following a description of the interactive input to MIGRAY is given. For a more
detailed explanation of some important parameters and the methodological background
see Raynaud (1988b).

Once you have set up the *.gst and *.dig files you may now start to run the program by
typing MIGRAY. Messages prompted on the terminal are displayed italic.

Figure B.2. Example of a *.dig file.

<table>
<thead>
<tr>
<th>WESTEST</th>
<th>-999999-999999 end flag</th>
</tr>
</thead>
<tbody>
<tr>
<td>-999999-999999 corner flag</td>
<td></td>
</tr>
<tr>
<td>+000000+000000 top left corner</td>
<td></td>
</tr>
<tr>
<td>+020790-000030 top right corner</td>
<td></td>
</tr>
<tr>
<td>+020857+016672 bottom right corner</td>
<td></td>
</tr>
<tr>
<td>+00025+016675 bottom left corner</td>
<td></td>
</tr>
<tr>
<td>-999999-999999 end flag</td>
<td></td>
</tr>
<tr>
<td>+001967+006020 start of first line segment</td>
<td></td>
</tr>
<tr>
<td>+0002117+005915</td>
<td></td>
</tr>
<tr>
<td>+002305+005837</td>
<td></td>
</tr>
<tr>
<td>+002450+005772</td>
<td></td>
</tr>
<tr>
<td>+002615+005727</td>
<td></td>
</tr>
</tbody>
</table>

+007997+008370 end of first line segment
+008070+008450
+008127+008507
+008177+008557 .999999-999999
+006572+007995 start of second line segment
+006685+007955
+006822+007917

+013807+011585 end of second line segment
+013852+011640 .999999-999999
+011160+010147

+020837+013545 .999999-999999
-999999-999999 end of *.dig file
The first two questions you will be asked concern the device specification of your graphics terminal followed by:

Option required; 1 for analysis, 2 for graphics, 0 to stop

Interactive Input to an Analysis Run (Option 1)

Group pathname for data files
This string precedes the attributes of the data files created during a run of MIGRAY (filename = group pathname. attribute).

File pathname ghost structure data
Name of *.gst file.

File pathname digitised line drawing data
Name of *.dig file.

x- and t-limits of window
Note: only line segments lying within this x-t-window will be considered for migration and display.

Angle (in degrees) between line and dip direction
If the original seismic line was not shot perpendicular to the structural grain, this option allows a corresponding geometrical correction to be applied to the normal incidence ray paths. This "2.5-D" option will perform correctly up to angles of about 75° between line and dip. For bigger angles it becomes numerically unstable since the ray vector depends on the reciprocal value of the cosinus of this angle.

Pretty factor; p=1 no effect, 0<p<1 makes smiles worse, p>1 reduces smiles
This allows to reduce smiling by removing strongly curved parts from the depth migrated line elements.

The main reasons for smiles in line drawing depth migration are somewhat similar to those in wave equation migration: noisy or coarse digitised data set, incorrect velocity structure, incorrect angle between line and dip directions.
Other possibilities to reduce smiling are: increasing the length of horizontal smoothing in the *.gst file, smoothing of the velocity field either in the *.gst file or by choosing the "quasi 1-D" option (i.e. ray theoretical time migration) in the analysis part (see there), a smile filter in the graphics part (see there) or reducing the minimum amplitude to be plotted in the graphics part (see there).

**Full 2-D (1) or quasi 1-D (0) structure**

(1) Velocity structure as defined in *.gst file.

(0) Ray theoretical time migration. The original 2-D velocity structure is approximated by a sequence of 1-D velocity functions, i.e. all interfaces are being treated as being locally horizontal in the ray tracing. This causes a smoothing of the velocity structure and reduction of smile generation by short-wavelength variations of the velocity field.

**Stacked (0) or time migrated (1) section**

(0) Migration and depth conversion by inverse normal incidence ray-tracing.

(1) Depth conversion along the image rays.

0 to see only the structure data migrated, 1 to see only the line drawing migrated, 2 to see both the structure and the line drawing migrated, 3 to see the image rays of the structure, 4 to see both the image rays and the migration

Options 3 and 4 of course make only sense, if the line drawings were digitised from a time migrated section.

**Interactive Input to a Graphics Run**

*group pathname for data files*

This string precedes the attributes of the data files created during a run of MIGRAY (filename = group pathname. attribute).

(1) time section, (2) depth sections, (3) structure and ray paths, (4) profiles, (5) change plot characteristics, (6) quit from graphics run return to MAIN

(1) --> original digitised time section (0) or smoothed time section (1);
1 to clear screen or 0 to overlay plots

(2) --> 1 to clear screen or 0 to overlay plots
(3) --> ray density; one in every ... (0 no rays, 1 all the rays, 10 every tenth ray)
    1 to clear screen or 0 to overlay plots

(4) --> profiles in time (0) or depth (1)
    1 to clear screen or 0 to overlay plots

(5) --> squash factor
    >1 compresses the plot in the x-direction, <1 extends the plot in the x-direction
    corners of viewing area

    smile filter on or off
    If the smile filter is switched on, segments with crossing normal incidence ray will not be shown in the depth migrated section.
    (i) steepest line segment to be drawn, (ii) minimum amplitude to be drawn,
    (iii) minimum amplitude for emboldening
    (i) depends somewhat on the velocity structure; normally dips up to 70° cause no problems.
    (ii) Can be used to reduce smiling; a value between 0.1 and 0.5 is then generally appropriate.
    (iii) If only amplitudes that got stronger during migration are to be emboldened: >1.

(6) --> Option required; 1 for analysis, 2 for graphics, 0 to stop
MIGRAYHP

NON-INTERACTIVE INPUT TO MIGRAYHP
Same as for MIGRAY.

INTERACTIVE INPUT TO MIGRAYHP
(To produce A3 size plots on the HP7550 plotter; installed on JANUS and AEOLUS)

Filename for 7550 plots (A80):

Enter entry number (0-22): Normally 0

Enter option number required; 1 for analysis, 2 for graphics, 0 to stop

Interactive Input to an Analysis Run (Option 1):
Same as for MIGRAY.

Interactive Input to a Graphics Run (Option 2):

Enter group pathname (i.e. prefix) for data files: Same as for MIGRAY.
Enter option number: Same as for MIGRAY.

Automatic (0) or variable (1) scaling:
(0): The plot will be scaled to fill the available paper size (A3).
(1): You will be asked the scaling factors in cm/km and cm/sec TWT. If the plot does exceed the available paper size, this will be reported and you have to input more appropriate scaling factors.

To submit a plot to the HP7550 pen plotter:
• Exit from MIGRAYHP
• Type SUBP <name of plot file>
MIGRAYVT

NON-INTERACTIVE INPUT TO MIGRAYVT
Same as for MIGRAY.

INTERACTIVE INPUT TO MIGRAYVT
To produce "arbitrary size" plots on the 36 inch Versatec plotter; installed on REFLEX only.

Interactive Input to an Analysis Run (Option 1):
Same as for MIGRAY.

Interactive Input to a Graphics Run (Option 2):
Same as for MIGRAYVT.

To submit a plot to the 36 inch Versatec plotter:
• Exit from MIGRAVT
• Type QRAST
APPENDIX C: Description of MIGART

MIGART is a modular FORTRAN77 program to perform ray theoretical depth migration of digitised line drawings of seismic reflection profiles through an arbitrary two dimensional velocity structure.

The major difference between MIGRAY and MIGART is the definition of the velocity structure of the medium. Whereas MIGRAY requires the velocity to be constant within each layer, MIGART allows it to vary both vertically and laterally. This makes MIGART computationally more expensive than MIGRAY but also more versatile, particularly for the simultaneous inversion of seismic reflection and refraction data. The model definition is essentially compatible with RAY84 (Luetgert 1988), RAY87 (Sierro 1988) and SEIS83 (Cerveny and Psencik 1984), three of the most frequently used forward ray tracers.

The program consists of three major modules: MAIN, ANALYS and GRAPH. MAIN organises the large scale process flow; ANALYS performs essentially all the calculation work; GRAPH handles the graphic display of the results of a previous run of ANALYS (see Figures C.1 and C.2).

Currently there is just an interactive screen version of the program which is functional and tested on the large but still subject to numerous modifications on the small:

- allows you to plot interactively on a Tektronix compatible screen
- linked to TERMPLTLIB (cf. description by Edi Kissling);
- installed on AEOLUS and JANUS
CALL PLOTS
Initialise plot software

CALL PLOT (....,...,-3)
Set origin

MENU:
0 to stop
1 for an analysis run
2 for a graphics run

1 or 2

CALL GRAPH
Graphics run

CALL OPENFL
Open data files

CALL ANALYS
Analysis run

CALL CLOSFL
Close data files

CALL PLOT (0..0..+999)
Deactivate plot software

STOP

Figure C.1. Flow of control in MIGART.
Figure C.2. Large scale structure of MIGART.
MODULES

MAIN
Opens and closes the data files, initialises and deactivates the plot software and directs the process flow to either run ANALYS, GRAPH or exit the program. After each run of ANALYS or GRAPH control always returns to MAIN.

OPENFL
Opens units 8 to 11, 13 to 15 and 17 for writing (analysis run) or reading (graphics run).

CLOSFL
Closes units 8 - 16 opened either for analysis or graphics. Remember: units 8 to 16 were opened for analysis; only units 8 to 11, 13 to 15 were opened for graphics.

OFIL
Opens units 12 (*.dig) and 16 (*.gst) for an analysis run. Not used in a graphics run.

ANALYS
Directs the process flow for an analysis run. Interactive input: x-and t-window, angle between line and dip.

STRUCT
Reads in the parameter file (*.gst, unit 16).

INREAD
Auxiliary subroutine to read in real data that are supplied at ten values per line in free format.

READINT
Auxiliary subroutine to read in integer data that are supplied at ten values per line in free format.

CHUCK
Auxiliary subroutine to produce output with ten real values per line in the format 10F7.2.
EXTEND
Adds one more extra vertical speed line on either side of the model in order to prevent ray loss.

EXTRAP
Bounds the top and bottom of each layer by horizontal speed lines. Together with the vertical speed lines defined in the original input model this results in a velocity grid defined by rectangles of variable size.

SPLIN1
Performs linear interpolation through a line defined by more than two points. Output are the coefficients of the individual line segments.

MIGRAT
Organises the migration of a reflection segment.

READSEG
Reads in a reflection segment from unit 12 (*.dig), converts it into physical units and corrects for the angle between the line and strike direction.

SMTH3
Three point smoothing of line segments on both x- and t-values by applying weights of 0.25, 0.5, 0.25 and subsequent summing.

PRERAY
Determines the take off angle of the ray for a particular line element.

RAY
Traces one ray through the model.

RKGS
Solves the ray equations using a fifth order Runge-Kutta method with adaptive stepsize control.
FCTK2
Provides the right-hand side of the ray equations as input to RKGS.

VELIN2
Determines the layer and the rectangle number of a particular ray point and then obtains
the velocity and its vertical and lateral gradients by two-dimensional linear interpolation
of the gridded velocity model produced by EXTRAP.

INTERSECT
Checks whether the ray has hit an interface and if so adjusts the coordinates and the travel
time of the ray.

REFRACT
Applies Snell's law whenever a ray has hit an interface. In the case of overcritical
reflection the ray and the corresponding line element are discarded.

ENDRAY
Writes out successful rays to unit 11 (*.ray).

ENDSEG
Writes out successfully migrated reflection elements to unit 13 (*.zse).

GRAPH
Organises the flow of control for a graphics run.

RDSTR
Reads in the extended and dip-corrected structure data from unit 14 (*.cont).

OPTION
Offers menu for graphic display of the results. After each plot control always returns to
here via GRAPH.
Allows the following changes to the graphic display of the results:
• Squash factor (>1 squeezes the plot in the x-direction)
• Change x-window

PLOTSETUP
Gets everything ready for plotting.
In MIGART the scaling factor is determined so that the plot will fill the screen.

PLTNOW
Calls PLOTSETUP and the plotting routines corresponding to the choice made in OPTION.

NXTPLT
Prepares the next plot. Flushes the plot buffer, restores alphanumeric mode and determines whether the new plot is to overlay the old one or the graphics plane is to be cleared.

AXES
Draws a frame around the section and labels it.

AXE
Draws an axis and labels it.

DRAW
Converts a given number of physical coordinates into plotting coordinates and draws a line through them.

DRAWT
Draws a raw or smoothed unmigrated time section.

DRAWZ
Draws a migrated depth section.

PLTSTR
Plots the velocity-depth structure.

PLTRAY
Plots the normal incidence or image rays rays used for migration in the requested density. A ray density of 0 will plot no rays at all, one of 1 will plot all the rays, one of 10 will plot every tenth ray.

XPLOT and ZPLOT
Auxiliary functions to transform the original x (km) and z (km or secs) values into plot coordinates.

Figure C.3. Modular structure of the analysis part of MIGART.
Figure C.4. Modular structure of the graphics part of MIGART.
INPUT
Interactive Input
See "User's Guide to MIGART".

Noninteractive Input
Input to MIGART are a parameter or "ghost" file (*.gst, unit 16) and a file containing the digitised line drawing (*.dig, unit 12). For more details see "User's Guide to MIGART".

*.GST FILE (UNIT 16)
This file contains all the crucial information for the definition of the velocity structure used for migration. The comment line which precedes each data field makes its handling largely self-explaining. The data are given in free format. This file is read by STRUCT. For more details see "User's Guide to MIGART".

*.DIG FILE (UNIT 12)
This file contains the coordinates of the digitised line segments preceded by a title. The title and the individual line segments are separated by flags  -10'000.

OUTPUT
Interactive Output
This essentially consists of the graphic display of the unmigrated time section, migrated depth section, velocity structure, ray paths and velocity profiles in time or depth.

Noninteractive Output
This consists of eight files (units 8-11,13-15,17) which are briefly described below:

*.STR (UNIT 8)
This is essentially a copy of *.GST with the structure data extended by one vertical speed line to the left and right and corrected for the angle between the line and the dip direction.

*.RAW (UNIT 9)
Contains the coordinates of the original digitised line segments in free format.

*.CONT (UNIT 10)
Coordinates of the vertical speed lines and the gridded velocity model.

*.RAY (UNIT 11)
x- and z-coordinates of the rays each in the format 10F7.2.

*.ZSE (UNIT 13)
Coordinates of line segments after depth migration in free format.

*.COR (UNIT 14)
x- and t-values of the corners of the viewing window.

*.SMT (UNIT 15)
Coordinates of smoothed line segments (vertical axis is time) in free format.

*.PRI (UNIT 17)
Run time documentation.

**COMMON BLOCKS**

**COMMON Blocks used in an Analysis Run**

**COMMON / IOTERM/**

SCRN unit 6 (defined in MAIN for interactive output to the screen)

KBRD unit 5 (defined in MAIN for interactive input from the keyboard)

**COMMON / IOFILE/**

I analysis(1), graphics(2) or stop(0)?

**COMMON / CONTROL/**

ISTOP set to 0 when input error in STRUCT was encountered

FLAG flag used in *.zse and *.ray
XSCL, TSCL: scaling factors to convert the digitised coordinates into physical ones
XLFT, XRG T: x-range of the model
DPTH: maximum depth to consider

COMMON / ANALYS /
XRW, XLW: x-range of the viewing window
TTW, TBW: time range of the viewing window
DIOC OR: factor to correct for crossdip

COMMON / ITTLE /
TTITLE: title of model file

COMMON / BOUND /
NBND: number of interfaces
NPOINT(100): number points defining a particular interface
XCOR(100,100): x-coordinates of a particular interface
ZCOR(100,100): z-coordinates of a particular interface

COMMON / LAYVEL /
RVEL1(100,100): velocity at the top of a particular layer
RVEL2(100,100): velocity at the bottom of a particular layer

COMMON / VELMAT /
SPEEDZ(100,2): horizontal speed lines bounding each layer
ALFA(100,100,2): velocity matrix of the gridded model

COMMON / SPEEDLN /
NSPEED: number of vertical speed lines
SPEED(100): ranges of the vertical speed lines

COMMON / RUKU /
STEPMIN: minimum time step for integration
STEPMAX: maximum time step for integration
SIGMA used to determine the length of the time step (0.0015 to 0.05)
EPS maximum error in seconds allowed for one integration step

COMMON Blocks used in a Graphics Run

COMMON / IOTERM /
SCRN unit 6 (defined in MAIN for interactive output to the screen)
KBRD unit 5 (defined in MAIN for interactive input from the keyboard)

COMMON / PLOTOPT /
IFI plot option
INDT 0:original, 1:smoothed time section
NRAY plot every NRAYth ray

COMMON / CHANGEPLOT /
SQUASH squash factor
XPSL,XPSR new x-range of the viewing window

COMMON / WINDOW /
XRW,XLW x-range of the viewing window
TTW,TBW time range of the viewing window

COMMON / LIMITS /
XLFT,XRGT x-range of the model
DPTH maximum depth to consider

COMMON / INTERFACES /
NBND number of interfaces
NPOINT(100) number points defining a particular interface
XCOR(100,100) x-coordinates of a particular interface
ZCOR(100,100) z-coordinates of a particular interface
COMMON /SPEEDNET /
NSPEED number of vertical speed lines
SPEED(100) ranges of the vertical speed lines
SPEEDZ(100,2) horizontal speed lines bounding each layer

COMMON /SETUP /
XL,XR x-range used for plotting
ZT,ZB time or depth range used for plotting
XLENGTH length of the x-axes
ZLENGTH length of the z-axes
XINT,ZINT tick intervals
DXSCAL scaling factor for plotting
DZSCAL scaling factor for plotting
APPENDIX D: User's Guide to MIGART

MIGART is an interactive program to migrate digitised line drawings of reflection seismic data through an arbitrary 2-D velocity structure containing vertical and lateral velocity gradients. The velocity structure of the model is defined by velocity values supplied along a number of vertical speed lines at the top and bottom of each layer. At any point in a particular layer the velocity and its vertical and lateral derivatives can thus be obtained by two-dimensional linear interpolation.

The program consists of two major collections of modules to perform the calculation work related to migration (analysis part) and to handle the graphic display of the results (graphics part). Each run of the graphics part must be preceded by at least one analysis run for a particular data set, though not necessarily within the same session.

The model definition is essentially compatible with RAY84 (Luetgert 1988), RAY87 (Siervo 1988) and SEIS83 (Cerveny and Psencik 1984), three of the most frequently used forward ray tracers.

Currently there is just an interactive screen version of the program which is functional and tested on the large but still subject to numerous modifications on the small:

MIGART
- allows you to plot interactively on a Tektronix compatible screen
- installed on AEOLUS and JANUS

NON-INTERACTIVE INPUT TO MIGART
Although the program is largely interactive, the two input files containing the model parameters and the digitised line elements cannot be accessed interactively at present.

*.gst ("Ghost") file
This file contains all the information on the velocity structure to be used for migration. The data can all be supplied in free format (Figure D.1).
The comment field which precedes each data field makes the handling of this file largely self-explaining. Only the last data field (STEPMIN, STEPMAX, SIGMA, EPS) needs some extra consideration.

- **STEPMIN**: minimum stepsize in seconds; generally values between 0.01 and 0.001 are appropriate. This parameter is not really crucial for a correct functioning of the algorithm since the Runge-Kutta integration will, if necessary, adjust it to achieve the required precision (see description of EPS below).
- **STEPMAX**: maximum stepsize in seconds; this value should not be bigger than the minimum one-way travel time of a ray through the thinnest layer otherwise the program adjusts STEPMAX to the corresponding value. This may drastically reduce the overall numerical efficiency of MIGART (see definition of TSTEP below) and hence the user should make every effort to avoid thin layers - let alone layers shrinking to zero thickness - but rather use velocity gradients to accommodate complicated structures.
- **SIGMA**: MIGART uses variable time steps for the integration of the ray equations. The time step TSTEP (STEPMIN<TSTEP<STEPMAX) is evaluated as follows (Zelt and Ellis 1988):

\[
TSTEP = \frac{\text{SIGMA}}{|V_x| + |V_z|}
\]

The SIGMA-value to be chosen depends somewhat on the desired accuracy EPS (see below); the higher the desired accuracy, the smaller the value for SIGMA. Zelt and Ellis (1988) have shown that SIGMA should be in the range from 0.0015 to 0.05.

- **EPS**: desired accuracy in seconds for each integration step. Values between 0.0001 and 0.001 are generally appropriate in crustal reflection seismology.
S5 Test
Maximum number of rays to be traced (<32'000)
10000
Flag to indicate end of line segment (<-10'000)
-.999999
xscale [km/digitising unit] and tscale [s/digitising unit]
0.00041 0.000249
Now supply the structure data:
Left model boundary,right model boundary,maximum depth (all in km)
0 20 60
Total number of boundaries
5
Number points per boundary (top to bottom)
2 2 2 2 2
Now supply the x-&z-coordinates in km of each layer from left to right (10 values per line):
1st boundary
0 20
0 0
2nd boundary
0 20
12 12
3rd boundary
0 20
24 24
4th boundary
0 20
38 38
5th boundary
0 20
60 60
Number of vertical speed lines
2
x-coordinates of the speed lines
0 20
Now supply the velocities along the speed lines at the top and the bottom of each layer (10 values per line):
1st layer
4.5 4.5
5.5 5.5
2nd layer
6.0 6.0
6.2 6.2
3rd layer
6.5 6.5
6.5 6.5
4th layer
8.1 8.1
8.1 8.1
STEPMIN,STEPMAX,_SIGMA(0.0015 - 0.05),EPS
0.001 0.5 0.03 0.0001

Figure D.1. Example of a "ghost" file.
*.dig file
Column containing the x- and t-coordinates of the digitised line drawing. The individual line segments are separated by flags (<-10'000).
Both coordinates and flags are in free format.

**INTERACTIVE INPUT TO MIGART**
In the following a description of the interactive input to MIGART is given. Once you have set up the *.gst and *.dig files you may start to run the program by typing MIGART. Messages prompted to the screen are displayed italic.

The first two questions you will be asked concern the device specification of your graphics terminal followed by:

Option required; 1 for analysis, 2 for graphics, 0 to stop

**Interactive Input to an Analysis Run (Option 1)**

*Group pathname for data files*
This string precedes the attributes of the data files created during a run of MIGART (filename = group pathname. attribute).

*File pathname ghost structure data*
Name of parameter (*.gst) file.

*File pathname digitised line drawing data*
Name of *.dig file.

*x- and t-limits of window*
Note: only line segments lying within this x-t-window will be considered for migration and display.

*Angle (in degrees) between line and dip direction*
If the original seismic line was not shot perpendicular to the structural grain, this option allows a corresponding geometrical correction to be applied to the normal incidence ray
paths. This "2.5-D" option will perform correctly up to angles of about $75^\circ$ between line and dip. For bigger angles it becomes numerically unstable since the ray vector depends on the reciprocal value of the cosinus of this angle.

Stacked (1) or time migrated (2) section
(1) Migration and depth conversion by inverse normal incidence ray-tracing.
(2) Depth conversion along the image rays.

Interactive Input to a Graphics Run

group pathname for data files
This string precedes the attributes of the data files created during a run of MIGART (filename = group pathname. attribute).

(1) time section, (2) depth section, (3) structure and ray paths, (4) change plot characteristics, (5) quit from graphics run and return to main menu

(1) ---> original digitised time section (0) or smoothed time section (1);
     1 to clear screen or 0 to overlay plots
(2) ---> 1 to clear screen or 0 to overlay plots
(3) ---> ray density; one in every ... (0 no rays, 1 all the rays, 10 every tenth ray)
     1 to clear screen or 0 to overlay plots
(4) ---> squash factor
     >1 compresses the plot in the x-direction, <1 extends the plot in the x-
     direction
corners of viewing area
(5) ---> Option required; 1 for analysis, 2 for graphics, 0 to stop
APPENDIX E: Outlook - Ray Tracing and Beyond

Musings on Velocity Model Parametrisation in Seismology

The velocity models of today's most popular ray tracers for seismic forward modelling are parametrised on the basis of laterally continuous layers in which the velocity is allowed to vary both vertically and laterally (e.g. Cerveny and Psencik 1984; Luetgert 1988; Sierra 1988; Zelt and Ellis 1988). Due to the combination of velocity gradients and first-order discontinuities this parametrisation scheme is very flexible and versatile and allows the user to efficiently construct and modify highly complicated models. There are, however, also drawbacks to this concept upon which I wish to comment in the following based on my - necessarily subjective - experience as a programmer:

• This scheme not only allows but rather invites the use of velocity models that violate the basic assumptions of the ray approximation in seismology and thus implicitly questions the usefulness of ray amplitudes.

• The presence of complicated structures, such as layers shrinking to zero thickness, in a model drastically reduces the numerical efficiency of the ray tracer. The reason for this is that either the maximum ray step length becomes very small or that the program first has to search the model for such pathological regions and afterwards correspondingly check every ray passing through the model. Given the power of today's computers this argument is not really a vital one in the case of forward modelling where one generally considers a few tens of rays per run. In the case of migration, however, each run typically consists of thousands of rays and hence numerical efficiency still is of critical importance.

• The versatility of a layer-based model parametrisation scheme may not only make the comparison or combination of identical or nearby profiles interpreted by different workers difficult or impossible it also makes the practical extension of this concept into the third dimension anything but straightforward.

• Coordinate transformations, such as flat-to-round earth or vice versa, and two-point ray tracing (so-called bending methods) are difficult or impossible.

• The layer-based parametrisation used in seismic ray tracing is potentially incompatible with most other seismic forward modelling or inverse schemes. This hampers the basic intentions of the ray method in seismology to serve as an efficient first-order approximation to derive suitable starting models for further complementary and/or more sophisticated investigations.

To sum up: the layer-based velocity parametrisation for ray tracing is a high-level description of the velocity model. As such it is versatile, efficient and user-friendly but also prone for abuse and hard to be transported and used outside its mother program. Evidently the solution is to lower the level of the model description. A possible low-level description scheme of the velocity model which largely eliminates the above problems without creating too many new ones might consist of discrete individual velocity values distributed either randomly or according to simple rules such as user-defined vertical speed lines. First order velocity discontinuities *sensu strictu* are not allowed but can be decently approximated by sharp gradients. The velocity and its spatial derivatives at an arbitrary point in the model are obtained by suitable nearest-neighbour interpolation. Such a model parametrisation is relatively easy to implement and ideal for two-point ray tracing, can be extended to the third dimension without too many complications, supports the enhancement of numerical efficiency by varying the integration step-size according to the complexity of the model, is easy to transform into another coordinate system or into another model parametrisation scheme, makes it difficult to violate ray theory fundamentally and thus makes ray amplitudes potentially meaningful. A possible drawback is that this description of the velocity model is more abstract and hence not as user-friendly as the versatile layer-based one. This sacrifice is, however, easy to make for whoever agrees that ray tracing can and must be just the basis for further complementary and more complete seismic investigations.