CHARACTERIZATION OF THE FRASER FAULT, SOUTHWESTERN BRITISH COLUMBIA, AND SURROUNDING GEOLOGY THROUGH REPROCESSING OF SEISMIC REFLECTION DATA

by

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ABSTRACT

Seismic reflection data were acquired in 1988 along a 70 km profile crossing the Fraser fault, a major dextral strike-slip boundary between the Eastern Coast and Intermontane belts of the southern Canadian Cordillera. Preliminary processing by industry contract and subsequent interpretation revealed several interesting tectonic features; however, the subsurface position and depth extent of the fault were ambiguous. The present study involves reprocessing of these data in an effort to provide an improved subsurface image, and hence a better understanding of the local tectonic regime.

Severe crookedness of the line necessitated the implementation of unconventional processing techniques, including creation of several “mini-profiles” which cut through the scatter of source-receiver midpoints near the fault zone, and application of a first-order correction for effects of reflector crossdip (i.e., the crossline component of reflector dip). A method for estimating the optimum crossdip correction parameter was developed. Although it demonstrated promise when applied to synthetic data, disappointing results were obtained with the field data. Refraction-based statics were computed using a two-step procedure consisting of initial identification and correction of systematic errors in the picked first break travel times, and subsequent application of a conventional 2-D statics algorithm. Dip moveout correction (DMO) was applied to the upper 5.0 s of data to enhance the image of steeply dipping features.

Significant aspects of the interpretation of the reprocessed dataset include: (i) evidence of deep crustal extent (or possible crustal penetration) for the Fraser fault; (ii) correlation of two northeast-dipping reflectors (not visible on the contract-processed sections) with southwest-directed thrusting along the Pasayten fault; and (iii) correlation of two east-dipping events near the western edge of the profile with deep roots of the Coast Belt Thrust System.
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To my grandparents . . .

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Chapter 1
INTRODUCTION

1.1 Background and research motivation

The Lithoprobe Southern Cordillera transect was established in an effort to provide a
detailed characterization of the Cordilleran orogeny, and to thereby elucidate the tectonic
processes which contributed to the westward growth of North America in the past, as
well as those which continue today to shape the western edge of the continent. As part
of the transect, several seismic reflection profiles were run in 1988 across southwest and
south-central British Columbia (Figure 1.1). Profile SBC18 crosses the Fraser fault, a
major Eocene strike-slip feature that has accommodated more than 100 km of right-lateral
displacement. The fault marks the boundary between the Eastern Coast and Intermontane
belts of the southern Canadian Cordillera. Restricted road access has resulted in a severely
crooked survey line; although the profile is nominally oriented perpendicular to the
northwest-southeast direction of regional strike, its easternmost half trends principally
north-south.

Contract processing of the SBC18 dataset by Western Geophysical and subsequent
interpretation (Varsek et al., 1993; Monger and Journeay, 1992) revealed, in the words
of Varsek et al. (1993), “one of the most complex patterns of dip domains observed
on any Lithoprobe seismic data across the southern Canadian Cordillera”. In particular,
the subsurface position and depth extent of the Fraser fault were ambiguous. These
considerations provided the original motivation for both careful reprocessing of the dataset
and subsequent interpretation, which together constitute the essence of this thesis.

The thesis is founded upon two principal objectives: (i) a more detailed subsurface
characterization along the entire profile, placing particular emphasis on structure near
Figure 1.1. Location map showing Lithoprobe reflection profiles in southwestern British Columbia (see legend), Consortium for Continental Reflection Profiling (COCORP) reflection profile W7 in northwestern Washington state, and Southern Cordillera Refraction Experiment (SCoRE) lines 3 and 10 (boldface dotted lines) (following page). Figure is adapted from Varsek et al. (1993). Major tectonic units are also displayed in the figure. The five morphogeological belts of the Cordillera are displayed in the inset. Abbreviations: Cascade metamorphic core (CMC), Central Nicola horst (CNH), Mount Lytton complex (MLC), Northwest Cascade system (NWCS), Okanogan plutonic complex (OPC), Ashlu Creek fault (ACF), Ashcroft fault system (AF), Bralorne fault system (BF), Beaufort Range fault (BRF), Central Coast Belt detachment (CCBF), Cowichan Lake fault (CF), Castle Pass fault (CPF), Chuwanter fault (ChwF), Crescent fault (CrF), Coldwater fault (CWF), Duffy Lake fault (DLF), Fraser fault (FF), Hozameen fault (HF), Harrison Creek fault (HCF), Jack Mountain fault (JMF), Kwoiek Creek fault (KF), Marshall Creek fault (MCF), Miller Creek fault (MIF), Mocassin Lake fault (MLF), Mission Ridge fault (MRF), Methow River fault (MtF), Okanagan Valley fault (OVF), Owl Lake fault (OLF), Pasayten fault (PF), Phair Creek fault (PCF), Quilchena fault (QF), Ross Lake fault (RLF), Red Shirt fault (RSF), Shuskan fault (SF), Thomas Lake fault (TLF), Tyaughton Creek fault (TyF), West Coast fault (WCF), Yalakom fault (YF).
FT Overlap Assembly/Successor Basins
Northwest Cascades System / Cascade Metam.
Core (Deformed Rocks from Adjacent Belts)

LITHOPROBE SOUTHERN CORDILLERA TRANSECT CORRIDOR

Volcanic Center (Neogene-Recent)

Plutons (Mesoz., Cenoz.)
North American Metasediments
Intermontane Belt
Insular Belt
Cascadia Belt
Overlap Assembly/Successor Basins

Northwest Cascades System / Cascade Metam.
Core (Deformed Rocks from Adjacent Belts)

FAULTS

Contraction
Extension
Strike Slip

OTHER

Seismic Profile
Exploration Well
Volcanic Center (Neogene-Recent)
the Fraser fault and, equally important, (ii) the establishment of an effective processing stream for future reprocessing of crooked line deep crustal datasets.

The first objective stems not only from the complexity of the dip patterns observed on the result produced by the contract processing, but also from the fact that new geological mapping and a recently-acquired better understanding of surface geological relations in the region together provide an excellent opportunity for relating the seismic section to surface features.

The second objective arises because our work represents the first-ever effort at reprocessing a southern Cordillera reflection dataset at the University of British Columbia, where several other reprocessing projects are either in initial stages of progress (SBC13 and SBC15 from Figure 1.1) or have been proposed (SBC14 and SBC17). Deep crustal datasets differ from their shallower counterparts encountered in the oil industry in several important ways. For instance, they are characterized by a lower signal-to-noise ratio (S/N) and the acquisition typically takes place along crooked lines. As a result, while several of the processing techniques successfully employed in the oil industry are readily adapted to the deep crustal environment (e.g., deconvolution, refraction statics), others require specialized modifications and/or yield disappointing results (e.g., DMO and 2-D migration). Moreover, the significant offline scatter of source-receiver midpoints necessitates the implementation of non-standard techniques in order to account for reflector crossdip (i.e., that component of dip perpendicular to the survey line).

In light of the multidisciplinary nature of the Lithoprobe project, the integration of our interpretation into the existing geophysical/geological framework necessarily figures among our objectives. However this third objective is ancillary to the above two in the sense that it represents a preliminary effort. A final synthesis encompassing results from all facets of the scientific program awaits the reprocessing of other reflection datasets and
the completion of refraction studies.

1.2 Tectonic setting

SBC18 straddles the boundary between the Intermontane and Coast belts, two of the five sub-parallel morphogeological zones into which the entire Canadian Cordillera is divided (Figure 1.1, inset). The formation of these two belts, and more generally, the evolution of the Cordillera as a whole, reflect a protracted history of Mesozoic and Cenozoic collision and deformation along the western margin of the North American continent.

The concept of a “terrane” is commonly invoked in describing the Cordilleran orogenesis. Yorath (1990) defines a terrane as “a part of the earth’s crust which preserves geological rocks different from those of neighbouring terranes”. Implicit in his definition is that no assumption is made about the location of origin of the rocks. While some terranes were formed in proximity to the ancient North American continent (“autochthonous” terranes), others travelled great distances before reaching their present locations (“allochthonous” terranes). The dominant mechanism responsible for the creation of the Cordillera is that of terrane “accretion”. Accretion refers to the process of attachment of the terranes onto the North American continent. The various terranes of the Canadian Cordillera are shown in Figure 1.2.

Various models have been proposed that attempt to explain the formation of the Coast belt. The generally accepted view is that of Monger et al. (1982), who postulate that the Coast belt contains the suture that resulted from the mid-Cretaceous collision between an exotic Insular “superterrane”, which is a composite terrane formed by pre-collisional amalgamation of Wrangellia and the Alexander terrane (Figure 1.2), and the North American continent, whose western margin at that time consisted of the previously accreted Intermontane superterrane, itself comprising three terranes—Stikinia, Cache
Figure 1.2. Terrane map of the Canadian Cordillera (adapted from Clowes, 1990). The various terranes are depicted using different shading patterns and are labelled with circled letters. Asterisks are used to denote terranes which are thought to have formed near the western margin of ancestral North America (autochthonous terranes). All other terranes are probably exotic (allochthonous terranes), having been accreted to the continent at various times in the Mesozoic.
Creek, and Quesnellia. The collision followed the subduction-related closure of an intervening ancient ocean basin, vestiges of which are presumably contained in the rocks of the Bridge River and Methow terranes. The suture has since been overprinted by a magmatic arc related to a post-collisional east-dipping subduction zone. Recently however, van der Heyden (1992) has suggested that the Coast Belt magmatic intrusions are due to a prolonged period of east-dipping subduction ongoing since the mid Jurassic, at which time a single “megaterrane” consisting of Wrangellia, Alexander and Stikinia was accreted to ancestral North America along a suture which is today represented by rocks of the Cache Creek and Bridge River terranes.

Between latitudes 49° and 51° N, Journeay and Friedman (1993) have divided the southern Coast belt into western, central and eastern parts, the easternmost two of which are crossed by SBC18 (Figure 1.3). The western Coast belt (WCB), which consists of voluminous mid-Jurassic to mid-Cretaceous plutons that have intruded mid-Triassic to early Cretaceous arc sequences, is of low metamorphic grade. Except in the vicinity of its eastern flank, where it is imbricated along a series of east-dipping thrusts, the WCB is presumed to have acted as a rigid crustal block during two stages of vigorous late Cretaceous shortening that took place along a west-vergent contractional belt, the Coast Belt Thrust System (CBTS) (Figure 1.3, inset). Journeay and Friedman (1993) interpret this shortening to be a reflection of (i) late phases of the northeastward underplating of the Insular superterrane, postdating the main collisional event and/or (ii) westward ramping of the central Coast belt (CCB) over the WCB. In the study area, the CBTS extends across strike for approximately 35 km, encompassing both the eastern flank of the WCB as well as the western portion of the central Coast belt (CCB), where it is manifest in a stack of east-dipping thrust sheets which carry highly metamorphosed deep levels of the eastern Coast belt (ECB) in their hanging walls. The CCB and ECB feature metamorphosed oceanic rocks of the Bridge River and Methow terranes, as well as island
arc sequences of Cadwallader terrane affinity. Of particular significance to our study is the fact the basal detachment associated with the later stage of late Cretaceous shortening is thought to root in the middle to lower crust to the northeast of the CBTS (Journeay and Friedman, 1993), near the western edge of the SBC18 profile (Figure 1.3).

In the vicinity of the study area, the Intermontane belt consists primarily of Quesnel terrane (Quesnellia) rocks, specifically Triassic and Jurassic granodiorites of the Mount Lytton Complex (Figure 1.4). The Mount Lytton pluton, which was probably emplaced at deep crustal levels within a Late Triassic-Early Jurassic Quesnellian crustal arc, was uplifted in late Early Cretaceous time together with the Late Jurassic-Early Cretaceous Eagle Plutonic Complex which it abuts to the south (Monger and Journeay, 1992). The uplift was apparently accommodated by thrusting on the southwest flanking, northeast-dipping Pasayten fault (Grieg, 1992), and was coeval with the extrusion of the Spences Bridge volcanic group (Monger and Journeay, 1992), which lies immediately to the northeast of the line (Figure 1.4).

The SBC18 profile is effectively divided into eastern and western halves by its intersection with the Fraser fault (Figure 1.1, notation “FF”), a major dextral strike-slip fault which in combination with its southern extension in Washington state, the Straight Creek Fault, stretches for at least 500 km in a north–south direction, cutting acutely across the north-northwest trending regional grain. The fault, or more properly the associated fault system (in the vicinity of our study area, the system is represented by the Cantilever fault—see Figure 1.4), was active in mid to late Eocene time (35–47 Ma, Monger and Journeay, 1992). Over the last fifteen years, various estimates of dextral offset across the fault have ranged from 80–190 km (Misch, 1977; Mathews and Rouse, 1984; Kleinspehn, 1985; Monger, 1985; Coleman and Parrish, 1991). The currently-accepted figure is 130 km (Parrish and Monger, 1992).
Figure 1.3. Geological map of the southeastern Coast belt and adjacent Intermontane belt, adapted from Journeay and Friedman (1993) (following page). Inset shows a composite geological cross-section of the Coast Belt Thrust System taken from profiles AA and BB displayed in the figure (modified from Monger and Journeay, 1992). Abbreviations: Thomas Lake fault (TLF), Fitzsimmons Range fault (FRF), Terrarosa thrust (TT), Fire Creek fault (FCF), Ascent Creek fault (ACF), Harrison Lake shear zone (HLSZ), Breakenridge fault (BF), Siollicum River fault (SRF), Central Coast Belt Detachment (CCBD), Castle Towers pluton (CTP), Ascent Creek pluton (ACP), Mt. Manson pluton (MMP), Breakenridge plutonic complex (BPC), western Coast belt (WCB), central Coast belt (CCB), and eastern Coast belt (ECB).
Figure 1.4. SBC18 profile (nomenclature "88-18") superimposed atop local geology maps of Monger (1989a, 1989b) (following page). Station numbers are identified in the figure. Extreme geological complexity of the region provides motivation for the detailed reprocessing of the SBC18 dataset. With reference to Section 1.2, note that in the vicinity of the profile, the Intermontane belt consists primarily of Quesnellian rocks (see stations 900-1500 on figure). Note also that the Fraser fault assumes the name of "Cantilever fault" on these maps.
Transcurrent motion along this fault and several other coeval strike-slip features found throughout the Cordillera may be related to an episode of Eocene extension, and was possibly a relaxation response to the transpressional nature of accretion of the Insular superterrane to the North American continent (Coleman and Parrish, 1991). In an alternate yet related interpretation, the right-lateral motion is attributed to shearing induced by the interaction between the northeast-bound Kula oceanic plate and the westward-moving continental plate (Yorath, 1990).

1.3 Thesis outline

In performing a detailed reprocessing of the SBC18 dataset, we attempted a number of different procedures. While some of these were successful, others yielded disappointing results. After considerable testing, we arrived at a final processing stream which embodies those more effective processes. We present the sequence, whose details and underlying theoretical considerations are described in Chapters 2–7, in Figure 1.5. In this section, we describe the organization of the thesis with reference to this figure.

In Chapter 2 we describe the data acquisition and the initial steps undertaken in reprocessing. Deconvolution is addressed within the framework of the deep crustal vibroseis survey.

In Chapters 3–5 we present details of our main processing stream together with relevant theoretical discussion. In Chapter 3 we address the statics computation, which is essentially a two-step procedure consisting of (i) identification of systematic error in the first break picks and subsequent error correction, followed by (ii) application of a conventional 2-D refraction statics algorithm. In Chapter 4 we describe the procedure we adopted in performing dip moveout correction (DMO) as well as our velocity analysis. In Chapter 5 we address the crossdip correction. We derive a 3-D reflection traveltime expression which allows us to verify the validity of the application of a first order
correction proposed by Larner et al. (1979). We also develop a method for estimating the magnitude of this correction.

In Chapter 6 we describe the additional processing of the portion of the dataset near the Fraser Canyon, and in Chapter 7 we describe the three final stages of processing—data merging, migration and coherency filtering.
In Chapter 8 we present our revised interpretation which is based on the results of our reprocessing; in Chapter 9 we provide a summary of the key aspects of our work together with a concluding discussion.
Chapter 2
ACQUISITION AND INITIAL PROCESSING

In this chapter, we provide details on the data acquisition and describe the initial processing steps. We present a thorough, tutorial-style discussion of the deconvolution problem in the context of vibroseis deep crustal processing. We address wavelet phase considerations, present synthetic tests on a mixed-phase wavelet which embodies anticipated reverberatory effects, and describe the deconvolution procedure used for SBC18. Finally, we discuss the spatial resolution of the dataset.

2.1 Acquisition details

Sonix Exploration Ltd. performed the data acquisition. The asymmetric split spread recording geometry is illustrated in Figure 2.1. The acquisition parameters are listed in Table 2.1. The data were recorded in “master–slave” configuration using two DFS-V recording units. The master unit recorded channels 121–240, while the slave recorded channels 1–120. Each receiver group consisted of 12 geophones which simultaneously measured the signal (Fig. 2.1, inset 1). For further redundancy in acquisition, four vibrator trucks vibrated simultaneously, performing eight separate frequency sweeps at a single shot station or “vibrator point” (VP). The trucks “moved up” a distance of 7.14 m between sweeps (Fig. 2.1, inset 2). The details of the source sweep are given in Table 2.1. Cross-correlation with the sweep signal was performed in the field.

Geophones were deployed at every station, so that the “group interval” was 50 m. Shooting took place at every second station (i.e., VP interval of 100 m), resulting in a nominal common midpoint (CMP) fold of 60 and a nominal CMP interval of 25 m (the “nominal” CMP fold is the number of coincident source-receiver midpoints that would correspond to our VP and group intervals in the case of a straight line survey;
Figure 2.1. Acquisition geometry for SBC18.

- **Ch#1** to **Ch#80**: Slave channels 1-80
- **Ch#81** to **Ch#120**: Slave channels 81-120
- **Ch#121** to **Ch#240**: Master channels 121-240
- **Slave recorder truck**
- **Master recorder truck**
- **Sweep 1** to **Sweep 8**
- **Station**
- **Geophone**
- **Vibrator truck**
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<tr>
<th>Recording filter</th>
<th>out-64 Hz; 72db/oct slope</th>
</tr>
</thead>
<tbody>
<tr>
<td>Record length (uncorrelated)</td>
<td>32 s</td>
</tr>
<tr>
<td>Record length (correlated)</td>
<td>18 s</td>
</tr>
<tr>
<td>Sampling interval</td>
<td>4 ms</td>
</tr>
<tr>
<td>Energy source</td>
<td>4 vibrators</td>
</tr>
<tr>
<td>Vibrator model</td>
<td>L. R. S. 315</td>
</tr>
<tr>
<td>Sweep frequency</td>
<td>10-56 Hz.; linear</td>
</tr>
<tr>
<td>sweep length</td>
<td>14 s</td>
</tr>
<tr>
<td>sweep taper</td>
<td>0.5 s</td>
</tr>
<tr>
<td>geophone type</td>
<td>Mark L10a</td>
</tr>
</tbody>
</table>

Table 2.1 Acquisition parameters (from Hostetler et al., 1989).

an analogous definition exists for the nominal CMP interval). At both ends of the line, the recording geometry was modified to allow a gradual tapering of the CMP fold, a procedure usually referred to as “roll-on” (“roll-off”).

Midway along the line, the SBC18 profile crosses the Fraser Canyon. While the rugged topography of the canyon walls prevented continuous recording, subsurface coverage beneath the canyon was obtained by means of the “undershooting” acquisition scheme that is illustrated schematically in Figure 2.2.

2.2 Processing system

Hardware used for the prestack processing consisted of a MicroVAX II minicomputer, a METHEUS Ω500 graphics processor and a Sony monitor. Poststack processing (migration and coherency filtering) was performed on a SUN SPARCstation 10. A Versatec 36” electrostatic plotter was used to plot the final sections presented in Chapter 7. We used the Inverse Theory and Applications (ITA) INSIGHT software for most processing tasks.
Figure 2.2. Acquisition near the canyon. (a) gap closes when cable reaches west edge of canyon. Trucks vibrate between stations 517 and 641; (b) undershooting. Master portion of cable moves to the east side of the canyon. Trucks back up and vibrate between stations 601 and 681; (c) undershooting. Cable remains fixed and trucks move to east side of canyon. Trucks vibrate between stations 682 and 763; (d) Trucks back up and vibrate between stations 721 and 767. Gap then reopens and normal roll-along shooting is resumed.
2.3 Initial processing steps

We obtained the unprocessed, demultiplexed common shot gathers (CSGs) in SEG-Y format from the Lithoprobe Seismic Processing Facility (LSPF) in Calgary. The LSPF also provided us with the Sonix Exploration “observer notes” which provide detailed information concerning the acquisition geometry, as well as a set of “survey notes” which contain the Universal Transverse Mercator Grid coordinates for every station along the line. We converted the data to ITA internal (binary file) format, and entered the line geometry information from the observer and survey notes into the trace headers.

Next, we applied a spherical spreading correction based on a single regional velocity-time function that we derived from the velocity analysis information available from Western’s preliminary processing. We performed first break picking for our refraction statics analysis using the ITA interactive editor “VAQ2”. This editor was also used for deleting noisy traces and correcting polarity reversals.

The next processing step was the creation of CMP “bins”. In a crooked line survey, source-receiver midpoints are not colinear. In order to process the dataset using two-dimensional (straight line) techniques, neighbouring midpoints are grouped into bins and each bin is treated as a single CMP in the subsequent processing. Such a procedure respresents a trade-off between spatial resolution and any S/N enhancement that we might hope to gain through an increased redundancy of information. Related to this limitation in subsurface resolution is the crossdip problem, which we describe in detail in Chapter 5.

We performed the binning using the ITA interactive module “CDPBIN”. The technique entails first choosing an effective profile line or “slalom” line which ideally cuts through the centre-of-mass of the midpoint scatter, then specifying the (rectangular) bin dimensions (Figure 2.3). The bin “width” is measured in the direction of the slalom line while the bin “length” runs perpendicular to the line (Figure 2.3, inset). We constructed
a "main" slalom line extending across the entire length of the profile (in Chapter 6, we describe some additional binning schemes that we used in the vicinity of the Fraser Canyon). We chose a bin width equal to the nominal CMP interval of 25 m. We chose a large bin length (2000 m) in order to encompass the majority of the midpoints. The CMP bin map for SBC18 is presented in Figure 2.3.

2.4 Data quality

From our initial processing, we observed that the data quality varied quite dramatically over the line. Near the western end of SBC18 and again just east of the Fraser Canyon, the signal-to-noise ratio was relatively high. On the other hand, shot gathers in the vicinity of station 500 were very noisy; according to the observer notes, this was due to the proximity of the fast-flowing Nahatlach River. However as a general statement, the data quality was good.

A periodic noise pattern was observed on many traces. It was associated with the proximity of receivers to the recording trucks. On average, 10 traces per shot gather were affected by this systematic ringing. Analysis of the amplitude spectra revealed the presence of spurious Fourier components between 40 and 50 Hz. Although a simple zero-phase notch filter succeeded in removing a significant portion of the noise for some traces, it did not improve the result for others. While this observation might suggest a selective application of the filter based on individual trace characteristics, such a scheme would be prohibitively time-consuming. Moreover, any wavelet alteration between filtered and non-filtered traces would adversely affect subsequent processing. On the other hand, a universal application of the filter (i.e., applying the filter indiscriminately to the entire dataset) would remove useful signal from the uncontaminated traces. With these considerations in mind, we decided to simply delete all of the contaminated traces.
Figure 2.3. The CMP binning process. The SBC18 profile (black dots) is displayed in the main figure together with scattered source-receiver midpoints (blue dots). Station numbers are indicated in black. The slalom line (red line) is defined by a series of points (red dots). Inset provides an expanded view of midpoint scatter near station 300 together with CMP bins. Because each bin acts as a single CMP in the subsequent processing, the algorithm attempts to tie bin numbers (red) to station numbers (black) to within a scale factor of ten. Additional slalom lines were run through the significant midpoint scatter near the Fraser Canyon.
A significant number of traces were contaminated by the direct shear wave arrival and/or ground roll energy both of which are manifest on the shot gathers as a linear pattern of coherent noise (Figure 2.4). Unfortunately, the typical time dips (slopes in \(x-t\) space) associated with these events were sufficiently steep to cause spatial aliasing, a phenomenon we discuss in Section 4.2.2. As a result, a standard \(f-k\) dip-filtering procedure (see, for example, Yilmaz (1987)) could not be employed. Considering that the majority of the spurious energy was confined to early times (0–2 s) together with the fact that our analysis was based on a crustal scale, we chose a simple approach to removing the coherent noise, namely the deletion of heavily contaminated portions of the dataset.

2.5 Deconvolution

Although deconvolution was performed after the refraction statics analysis (Chapter 3), conceptually it represents a preliminary processing step and so we describe it in this chapter. To appreciate the objective of the popular processing technique let us consider a noise-free, zero-offset (coincident source-receiver) experiment. From the viewpoint of the interpreter, the ideal output would consist of a “reflectivity spike series”, each spike being proportional to a reflection coefficient at depth. In reality, the output section deviates significantly from this ideal. Reverberations, earth attenuation effects, the intrinsic source signature, the nature of the near-surface coupling for both source and receiver, as well as the response of the recording instrument all conspire to hinder the interpretation of the recorded seismogram.

As a first order approximation, we may represent each of the above “undesirable” effects by a linear, time-invariant filter and model the output seismogram as the convolutional product of these filters together with the reflectivity spike series. The act
of deconvolution, then, is to remove the various filter effects in order to recover the reflectivity spike series.

In its various forms, deconvolution enjoys widespread use by seismic data processors, including those addressing the crustal scale problem. Yet despite its popularity, it remains one of the least understood and most often "abused" of all processing techniques. Fortunately, the abusers are afforded a degree of impunity, because the method is generally robust in character. We admit to a certain amount of exploitation of this robustness, for after considerable testing, we opted to use a standard predictive deconvolution algorithm.
In this section we describe our synthetic testing as well as the procedure undertaken in performing the actual deconvolution.

### 2.5.1 Deconvolution theory

#### 2.5.1.1 The convolutional model

We may write the mathematical form of the convolutional model as

\[ x(t) = r(t) \ast a(t) \ast k(t) \ast s_c(t) \ast r_c(t) \ast i(t) \ast e(t) \quad , \]

where the "\( \ast \)" symbol denotes convolution, and \( r(t) \), \( a(t) \), \( k(t) \), \( s_c(t) \), \( r_c(t) \), and \( i(t) \) are the linear, time-invariant filters which attempt to describe the reverberation, attenuation, source signature, coupling between source and near-surface, coupling between receiver and near-surface, and recording instrument response, respectively. \( x(t) \) is the recorded seismogram and \( e(t) \) is the reflectivity series. Yilmaz (1987) discusses the approximations made in forming a convolutional model. At this point in the analysis, we are assuming a zero-offset, noise-free, compressional wave dataset.

The recording filter impulse response, \( i(t) \), is known, since the recording filter parameters are documented (Table 1.1). Because we know the exact impulse response, in principle we could remove the effect of the recording filter deterministically by, say, performing a spectral division. However, for reasons discussed below, we are not interested in any enhancement in temporal resolution that such an exercise might afford, so conceptually, we will group the recorder response as part of the source signature term. The "new" source signature term is just the convolutional product of \( k(t) \) defined above and the recorder impulse response \( i(t) \).

The nature of the coupling between the vibrator baseplate and the earth surface is poorly understood at the present time, and as such is the topic of much active research (e.g., Baeten et al., 1988). This effect, though quite significant for a shallow
high-resolution survey, may be viewed as second-order for a deep crustal study and consequently neglected. Similarly, we ignore the complex coupling between the geophone array and the earth surface. Under these simplifying approximations, we can rewrite (2.1):

$$x(t) = r(t) \otimes a(t) \otimes k(t) \otimes e(t),$$  

where the "new" $k(t)$ incorporates the effects of both the source signature and the recording instrument response.

### 2.5.1.2 The wavelet

By definition, a wavelet is a signal that is both one-sided (i.e., it has a definite origin time before which it is identically zero) and stable (i.e., it has finite energy and therefore must die out as time progresses) (Robinson and Treitel, 1980). There is much ambiguity in the literature surrounding what constitutes "the wavelet" in the context of deconvolution. "The wavelet" $w(t)$ in our present discussion is given by

$$w(t) = r(t) \otimes a(t) \otimes k(t),$$

and as such it embodies both earth effects (i.e., reverberation and attenuation) and the source signature. This is the wavelet that, ideally, we would like to deconvolve from the seismic trace, since from equations (2.2) and (2.3)

$$x(t) = w(t) \otimes e(t).$$

### 2.5.1.3 Statistical deconvolution

At this point, it may appear that the deconvolution problem is solved, that we may simply calculate the wavelet inverse and use this as the operator in a deterministic
deconvolution. Unfortunately, such a procedure is predicated on the assumption of an *a priori* knowledge of the wavelet form. In practice, it is impossible to acquire knowledge of the reverberation and attenuation components of the wavelet, at least not without performing extensive field tests at considerable expense.

We may design a "statistical" deconvolution operator without knowing the shape of the wavelet if we make an assumption about the statistical properties of the reflectivity sequence. Specifically, we might assume that this sequence has a flat or "white" amplitude spectrum, \( A_\varepsilon(f) \), where \( f \) is the temporal frequency:

\[ A_\varepsilon(f) = A_o = \text{constant} \]  

(2.5)

(e.g., Yilmaz (1987)). Under this assumption, *the amplitude spectrum of the unknown wavelet is equal to the amplitude spectrum of the seismic trace* to within a constant scale factor (this is easily seen by applying the convolution theorem to equation (2.4)). Stated alternatively, the autocorrelation of the wavelet is equal, again to within a scale factor, to the autocorrelation of the input seismic trace.

### 2.5.1.4 Phase considerations

By way of the white reflectivity assumption (equation 2.5), we have a means of estimating the wavelet amplitude spectrum. As for the wavelet phase spectrum, standard statistical deconvolution algorithms make the assumption of *minimum phase*. As in the case of the wavelet defined in Section 2.5.1.2, a minimum-phase signal is well-behaved in that it is causal (i.e., it is identically zero before time zero) and stable. Moreover, minimum phase is preserved under inversion; this bodes well for our present situation since the wavelet inverse is the ideal deconvolution operator and we desire that this operator be well-behaved in the above sense. The question remains, however, as to the validity of the minimum phase assumption for our wavelet (equation (2.3)).
Robinson and Treitel (1980) consider a perfectly elastic earth model consisting of an arbitrary number of horizontal, homogeneous surface reverberating layers overlying a deep reflector. Ignoring the time delay due to the two-way propagation time down to the deep reflector and assuming that the magnitude of all reflection coefficients is less than unity, they show that the impulse response at zero-offset or “reverberation spike train” is minimum phase. In the context of our convolutional model (equation (2.2)), this impulse response may be identified with the filter $r(t)$. The deep reflector propagation time is taken into account by the timing of the spike arrival in the reflectivity series.

We may infer from their analysis that the reverberating spike train encountered in many physical situations ought to be approximately minimum-phase. Even if the reverberatory effects observed on SBC18 conform reasonably to this simple model, our situation differs in two important respects: we do not have zero-offset data and the incidence is generally non-vertical. This means that strictly speaking, the periodicity of the multiples is not preserved (a lack of periodicity may adversely affect the deconvolution process). To rectify this situation, Yilmaz (1987) suggests performing a predictive deconvolution in the slant-stack domain (periodicity of multiples is preserved for non-zero offsets in this domain). We do not feel that the reverberatory problem for SBC18 is sufficiently severe to warrant such a complex analysis.

It may be argued that the impulse response of a homogeneous attenuating (i.e., inelastic) earth is approximately minimum phase. The justification arises from the consideration that the energy of the wavelet associated with the impulse tends to be concentrated near its onset or “front end”. Of the entire suite of signals with a given amplitude spectrum, the one whose energy is maximally concentrated near its front end is that signal having minimum phase (Robinson and Treitel, 1980).

Despite its apparent minimum phase character, it is not appropriate to directly
associate this impulse response with the time-invariant filter $a(t)$ in equation (2.2). The problem is due to the "non-stationarity" of the attenuated wavelet. Attenuation modifies the wavelet as it propagates through the subsurface, broadening it and removing high-frequency content. The true filter describing the attenuating earth is time-variant; yet filter time-invariance is one of the two fundamental assumptions underlying any convolutional model (the other being linearity). In our analysis, we have attempted to compensate for the non-stationarity by performing a time-variant deconvolution, designing different data-dependent operators for different time windows, then merging the output results. Within each window, we assume that the attenuation effect may be adequately represented by a minimum phase filter $a(t)$ in the convolutional model of equation (2.2).

We have seen that both reverberation and attenuation effects may be plausibly modelled using minimum-phase filters. Minimum phase is preserved under convolution. From equation (2.3) then, we see that we would have some reasonable representation of a minimum phase wavelet, if only the source signature $k(t)$ were minimum phase. Unfortunately for the present situation, it is zero-phase. The vibrator trucks input a signal to the ground that is "swept" through a range of frequencies over a time interval on the order of ten seconds. To increase the temporal resolution of the trace (i.e. to decrease the effective width of the basic source wavelet), the output seismogram is then cross-correlated with the source sweep signal. The source signature is the autocorrelation of the input sweep signal and is referred to as the "Klauder wavelet" (e.g., Yilmaz (1987)). Because it is an autocorrelation, the Klauder wavelet is zero-phase. The Klauder wavelet (modified by the zero-phase recording filter) for SBC18 is shown in Figure 2.5.

Gibson and Larner (1984) describe the phase correction that must be applied to seismic data generated by the Vibroseis source. The operator that performs the correction is that operator which transforms the Klauder wavelet, $k(t)$, to its minimum phase equivalent. Denoting the amplitude spectrum of the Klauder wavelet by $A_k(f)$, the
minimum phase equivalent of $k(t)$ has an identical amplitude spectrum and a phase spectrum $\phi_{\text{min}}(f)$ given by

$$\phi_{\text{min}} (f) = \frac{1}{2\pi} \int_{-\infty}^{\infty} \frac{\log_{e} A_k (f')}{{f'} - f} df'.$$

(2.6)

Unfortunately, the Klauder wavelet is a band-limited signal. For SBC18, the only meaningful energy content lies between 10 and 56 Hz. Moreover, the high-cut recording filter explicitly removes any high frequency noise. The implication relevant to the present discussion is that, strictly speaking, there is no minimum-phase equivalent for the Klauder wavelet because the natural logarithm is singular when $A_k(f)$ is equal to zero. To circumvent this, a small amount of "white noise" is added to all frequency components of the Klauder wavelet in creating the minimum-phase conversion operator.
The theoretically "rigorous" procedure for the deconvolution of our dataset may now be set forth: we find the (approximate) minimum phase equivalent of the Klauder wavelet using equation (2.6). By taking the ratio of the Fourier transform of the minimum phase signal to that of the original Klauder wavelet (the addition of a small amount of white noise is required in performing this spectral division due again to the band-limited character of the Klauder wavelet) and inverse transforming, we find the minimum phase conversion operator $m(t)$. By convolving $m(t)$ with the seismic trace $x(t)$, we see from equations (2.2) and (2.3) that we obtain a "new" wavelet $w'(t)$:

$$w'(t) = m(t) \otimes k(t) \otimes r(t) \otimes a(t) .$$

(2.7)

To within the approximations outlined in the discussion above, this new wavelet is minimum phase. Hence, the new seismic trace $x'(t)$ given by

$$x'(t) = m(t) \otimes x(t) ,$$

(2.8)

is in a suitable form for the application of a standard minimum-phase deconvolution.

There is one subtle point that remains to be addressed. The band-limited nature of the source signature implies that we cannot expect the output to be a true reflectivity spike series having energy in all frequency components. Any energy in the output seismogram at frequencies outside the source pass-band is, by necessity, noise. Hence, we must apply a bandpass filter or "interpreter's wavelet" to the deconvolved output. Recalling the band-limited character of the Klauder wavelet, it stands as a natural choice for such a filter. We should not consider the ideal output to be the reflectivity series alone, but rather the convolutional product of the reflectivity series and the Klauder wavelet. Because of the commutative and associative properties of the convolution operation, the method we have here outlined is identical to the phase-correction method of Ristow and Jurczyk (1975), which consists of applying standard "spiking" deconvolution followed by convolution with the Klauder wavelet minimum phase equivalent.
2.5.1.5 Predictive vs spiking deconvolution

In Section 2.5.1.4, we stated that the wavelet inverse is the ideal deconvolution operator. We referred to the process of convolving the seismic trace with an estimate of the wavelet inverse as "spiking" deconvolution, because the operator attempts to convert the wavelet to a spike. Although this would appear to be the process of choice, in practice noise amplification at higher frequencies becomes a problem. Even after convolution with an interpreter’s wavelet, we may find noise contamination in frequency components of the output that are "high", yet still resident within the source passband. We find in the process of "predictive deconvolution" a means of trading off between the resolution of the output wavelet and the high-frequency noise. The two parameters $\alpha$ and $n$, the prediction distance (also called the "prediction gap length") and prediction operator length, respectively, determine the action of the filter. The length of the prediction operator is not to be confused with the length of the predictive deconvolution operator, although these two quantities are related in a straightforward manner. Robinson and Treitel (1980) show that predictive deconvolution attempts to convert an input wavelet of length $\alpha+n$ into an output pulse of length $\alpha$. We will refer to this as the "pulse-shaping" property of predictive deconvolution in the discussion that follows.

Robinson and Treitel (1980) show that spiking deconvolution is a special case of predictive deconvolution with $\alpha$ set equal to unity. Hence, predictive deconvolution may be viewed as a generalized approach to deconvolution. It is hardly surprising then, that the minimum phase assumption is required for predictive deconvolution as well. Ulrych and Matsuoka (1991) show that the output of predictive deconvolution applied to a minimum phase wavelet is identical to the input wavelet for times up to the prediction distance, and is identically zero for times greater than this prediction gap length. In other words, for a minimum phase input, predictive deconvolution enjoys perfect success in
its pulse-shaping. If, on the other hand, the wavelet exhibits mixed phase characteristics, the output again preserves the wavelet shape up to the prediction gap, but beyond this time it may display a ringy "tail", whose nature is dictated by the phase characteristics of the input wavelet. As we shall see, the spurious effects of the tail may not be too severe for a suitably chosen prediction distance, given the zero-phase properties of our source signature.

2.5.2 Synthetic tests

The likely source of reverberation or "ringing" on the SBC18 dataset is the near-surface weathering layer, whose low velocity results in sharp acoustic impedance contrasts at both the free surface and the weathering layer-bedrock interface. In the following example, we consider the reverberation response associated with a one-dimensional, constant density model consisting of a constant velocity weathering layer overlying a bedrock refractor. We choose a weathering layer thickness of 100 m, a weathering layer velocity of 2500 m/s, and a refractor velocity of 5600 m/s. All these estimates are based on the refraction statics analysis performed in the following chapter. Because we want to identify this response with the filter $r(t)$ that appears in equations (2.2) and (2.3), we must account for the totality of the reverberations due to an initial downgoing source ray, as well as all ringing associated with a reflected upward travelling ray (the reflection being associated with an interface at depth) that impinges on the reverberating layer from below. The calculation of this "ringy" spike train is based on equation (12-18) of Robinson and Treitel (1980), who show that it is minimum phase. We display it in Figure 2.6a.

In Figure 2.6b, we have convolved the ringy spike train with the Klauder wavelet shown in Figure 2.5. For display purposes, we have time-shifted the output wavelet by +410 ms so that in effect, the output represents the result of a zero-offset, noise-free Vibroseis experiment conducted over a perfectly elastic, constant density one-dimensional
Figure 2.6. Comparision of predictive deconvolution and spiking deconvolution followed by phase correction: (a) the ringy spike train; (b) the convolutional product of the ringy spike train and the Klauder wavelet (output has been time-shifted downward by +410 ms for display purposes); (c) predictive deconvolution ($\alpha=33$ ms, $n=400$ ms) applied to the ringy wavelet in (b); (d) spiking deconvolution of the ringy wavelet (b) followed by convolution with the minimum phase equivalent of the Klauder wavelet. Abscissa is trace amplitude normalized to one and ordinate is two-way traveltime (s).
earth that consists of a single deep reflector (two-way traveltime 500 ms) embedded beneath a reverberating constant velocity weathering layer.

Figure 2.6c shows a predictive deconvolution applied "blindly" (i.e., without consideration of the wavelet phase) to the data trace in Figure 2.6b. Acknowledging the characteristics of the ideal output (i.e., a single Klauder wavelet centred on t=500 ms), and further recalling the pulse-shaping properties of predictive deconvolution described above, it makes no sense to choose a prediction distance $\alpha$ that is shorter than the effective length of the Klauder wavelet. Following the suggestion of Gibson and Larner (1984), we have used a prediction distance of 33 ms, this being the effective width of the central peak of the Klauder wavelet (Figure 2.5). Much of the ringiness has been removed. The spurious "tail" due to the mixed phase character of the wavelet is actually fairly innocuous, although this result is arguably less desirable from an interpretation viewpoint than the ideal output.

Figure 2.6d is the output of a deconvolution using the phase-correction method described in Section 2.5.1.4. Note that minimal improvement, if any, is afforded by accounting for the wavelet phase. This observation spurred us to apply and compare the two methods on a test portion of SBC18. Again, very little benefit was gained through application of the phase-correction procedure. Ulrych and Matsuoka (1991) show that the two methods (i.e., predictive deconvolution and the phase-correction method) are mathematically identical in the limit of infinite prediction distance. We conclude that a prediction distance judiciously chosen so as to bypass the effective width of the Klauder wavelet is sufficiently large that predictive deconvolution has a similar action to the more rigorous method that attends to the wavelet phase.

The results of this synthetic experiment suggest that a simple predictive deconvolution without minimum phase conversion suffices to remove the reverberatory effects of the
weathering layer from the SBC18 dataset. Having hereby waived the requirement of a minimum phase wavelet, we remain optimistic about the success of the deconvolution in removing both the ringiness due to the near-surface, as well as attenuation effects, to the degree that the other assumptions underlying the convolutional model (i.e., random reflectivity sequence, zero-offset data, the absence of noise, no complex source and/or receiver coupling with the near-surface) are not too severely violated.

2.5.3 Deconvolution in practice

We used the ITA prestack module GAP2 to perform a time-variant predictive deconvolution in the common shot domain. The module calculates the autocorrelation of the input traces in the shot gather and uses this information to estimate the wavelet and thereby design the predictive deconvolution or “prediction error” operator. We recall from Section 2.5.1.3 that the autocorrelation of the wavelet is identical to the autocorrelation of the input seismic trace under the white reflectivity assumption. Naturally then, an analysis of the autocorrelation of the input CSGs serves as a useful tool for estimating the characteristics of the input wavelet, and ultimately, for choosing the prediction operator length and prediction distance. Figure 2.7 is an “autocorrelogram” of CSG 773, that is, a plot of the autocorrelation of the traces of the input shot gather. The autocorrelation was computed over a time window from 9.0–12.0 seconds. The first isolated burst of energy that is common to all the traces should represent the autocorrelation of the wavelet (Yilmaz, 1987); hence, the length of the wavelet may be estimated by the length of the first transient zone on the autocorrelogram. We estimated the wavelet length here to be 160 ms.

In Section 2.5.2, we explained our rationale in choosing a prediction gap of 33 ms, but we have yet to choose the prediction operator length $n$. The pulse-shaping property of predictive deconvolution is such that an input wavelet of length $a+n$ is
shortened to an output pulse of length $\alpha$. Thus, the prediction operator length should be chosen so that $\alpha + n$ is equal to the estimated length of the wavelet. Yilmaz (1987) describes the degradation of the output result when the prediction operator length far exceeds the wavelet length. Our present analysis suggests a prediction operator length of $160 - 33 \approx 120$ ms.

The autocorrelogram analysis was performed on (roughly) every 40th CSG along the line. To account for wavelet non-stationarity, we partitioned the input CSGs into 4 time windows, computed separate autocorrelograms for each, and chose different prediction distances and prediction operator lengths, based on the individual autocorrelogram analyses. The deconvolution was performed separately over each window in the sense that the algorithm designed and applied a separate prediction error operator for each, based on our selection of input parameters $\alpha$ and $n$, then merged the results together on output. The top window extended from just below the first break arrival (we excluded the first
break energy from the analysis) down to 3.5 seconds. The second window extended from 3.5–7.0 s, the third from 7.0–10.0 s, the fourth from 10.0–13.0 s, and the fifth from 13.0–16.0 s. At certain points along the line, it was possible to see a slight trend toward lower frequency content and a more “dispersed” appearance in the autocorrelogram character as the time window increased with depth. This, of course, is a testament to the wavelet non-stationarity. However as a general statement, the variation was not particularly significant.

Figure 2.8 displays the first 4 seconds of CSG 776 before deconvolution. Figure 2.9a displays the results after a predictive deconvolution based on our parameter selection. Figure 2.9b displays the result after a time-variant deconvolution using a prediction operator length of 180 ms and a prediction gap length of 24 ms (these parameters were selected by Western in the preliminary processing). Clearly both deconvolution runs are “successful”, since they remove a great deal of ringiness. Comparing Figures 2.9a and 2.9b, it is also evident that there is not much difference between the two deconvolved datasets. The Western result has a slightly higher frequency content. This may be directly attributed to the discrepancy in the choice of prediction lags (24 ms vs 33 ms). However, it is unclear from this example whether one result is “better” than the other.

As a general statement, our final stacked section reveals enhanced coherency of events at later times compared with its Western counterpart. While some of the improvement is likely due to our statics application and crossdip correction, we believe that deconvolution has also figured significantly in the success. Recalling our discussion in Section 2.5.1.5, we expect a narrower prediction gap to boost high frequency noise. As well, we are unsure if the 24 ms gap is sufficiently large to avoid problems with wavelet phase. In short, we are content to trade off a small degree of temporal resolution in the interest of a slight enhancement of the S/N.
2.6 Spatial resolution

Beyond making the qualitative observation that deconvolution has succeeded in removing reverberation effects, thereby compressing the basic wavelet (equation 2.3) (or alternatively stated, broadening the wavelet spectrum), we may address the issue of vertical resolution in a quantitative manner by considering the dominant wavelength of the seismic waves. The effective vertical resolution is given by (Yilmaz, 1987) \( \frac{\lambda_e}{4} = \frac{v}{4f_e} \), where \( \lambda_e \) is the effective wavelength which may be computed from the velocity \( v \) and the dominant frequency \( f_e \). Taking \( v = 6000 \text{ m/s} \) and \( f_e = 30 \text{ Hz} \), we estimate a vertical
Figure 2.9. Comparison of predictive deconvolution runs applied to CSG 773: (a) our parameter selection; (b) Western's parameter selection. In both cases a spherical spreading correction and a 1000 ms AGC gain have been applied.
resolution of 50 m. In other words, we do not expect to be able to distinguish two reflectors which are separated by a depth interval of less than 50 m.

The lateral resolution may be quantified in terms of the Fresnel zone radius \( r(t) \) which is given by \( r(t) \approx \left( \frac{5.5}{f} \right) \left( \frac{1}{f_n} \right)^{1/2} \) (Yilmaz, 1987). Synthetic tests have indicated that a lateral dimension of \( r(t)/4 \) represents the threshold for event detection (Guy Cross, personal communication, 1993). Thus, at 2 s two-way traveltime, a corresponding Fresnel radius of 750 m implies that our experiment is capable of detecting a feature having a lateral extent of 200 m. At 10 s, the minimum detectable lateral dimension increases to roughly 500 m. Interpretation of the final stacked sections must be governed by the recognition that our processing efforts provide a subsurface representation which is distorted by significant lateral blurring.
Chapter 3
STATIC CORRECTIONS

The “statics” problem has received much attention over the last twenty years and today remains a primary focus of seismic processing research. Static corrections are applied to compensate for traveltime anomalies introduced by variations in near-surface structure. As such, they comprise the conventional elevation adjustments as well as corrections for heterogeneity in the shallow, unconsolidated “weathering” layer. Static effects whose spatial variation is on the order of several cablelengths (“long wavelength statics”) can lead to the erroneous positioning of deep reflectors on the stacked section. On the other hand, near-surface traveltime anomalies which vary significantly on the order of a single cablelength (“short wavelength statics”) can seriously impair stack quality since they introduce differential time shifts among traces in a single CMP gather (Farrell and Euwema, 1984).

We adopted a two-step procedure for our refraction-based statics computation, first identifying and correcting systematic errors in our picked first break traveltimes, then proceeding with the application of a 2-D statics algorithm.

3.1 The shortcomings of the residual statics method

We performed a preliminary assessment of the severity of the statics problem by constructing a stack that was essentially identical to the Western section, except for the application of statics. Whereas Western computed statics by means of a surface-consistent residual statics algorithm, we corrected only for elevation effects. The significance of near-surface heterogeneity was clearly illustrated by degradation in the quality of this preliminary stack.
Despite improvements gained through Western’s application of residual statics, it is important to note the method has shortcomings. Residual statics solutions are unable to reliably estimate the long wavelength component of the statics solution (Wiggins et al, 1976). Moreover, Western’s in-house residual statics algorithm “MISER” requires trace-to-trace static deviations as input. These are computed by a separate algorithm which measures relative time shifts across a reflection horizon from cross-correlations. In the deep crustal environment, the lack of a clear, continuous horizon can impair the performance of correlation-based picking routines. For instance, they may suffer from “cycle-skipping” problems due to low S/N. A cycle-skip occurs when the peak corresponding to the actual horizon is obscured by relatively high amplitude noise and “missed” in the picking analysis. Moreover, standard residual algorithms require a modification for crooked line geometries to account for the effects of reflector crossdip (Larner et al, 1979). Crossdip is the component of reflector dip perpendicular to the seismic line. Crossdip effects are discussed in detail in Chapter 5. This imposes an additional degree of solution non-uniqueness. The “MISER” routine did not correct for crossdip effects (James Hostetler, personal communication, 1992).

3.2 Refraction statics

In principle, an analysis of refraction traveltimes can provide a model-based statics solution that is complete in the sense that all orders of wavelength are correctly computed. Moreover, the “refraction statics” method is robust in the presence of noise, since for Vibroseis data, first breaks are easy to pick on all but the noisiest shot gathers. This is because the onset of first break energy is effectively located at the dominant peak of the first break event that appears on the gather (a consequence of the zero-phase character of the Klauder wavelet) and therefore usually affords unambiguous detection.
The practical reality is that restrictive assumptions made about the near-surface in the automated implementation of the refraction statics analysis as well as a sparse spatial distribution of shots and receivers may adversely affect the success of the final statics correction (Hampson and Russell, 1990). However, the ITA module "REFSTAT2" is a relatively sophisticated algorithm, and the multifold acquisition geometry of SBC18 ensures a reasonable sampling of the near-surface.

3.2.1 The REFSTAT2 model

REFSTAT2 assumes a two-dimensional model which consists of a constant-velocity weathering layer (velocity $V_w$), of variable thickness $h(x)$ and variable surface topography $e(x)$, overlying a bedrock refractor. $V(x)$, the velocity at the top of the bedrock, is constrained to be slowly varying. The variable $x$ here denotes position along the seismic line. The algorithm models the first breaks as head waves which propagate along the interface between the weathering layer and bedrock at the bedrock velocity (Figure 3.1). It should be noted that the algorithm can honour a multi-layer model, but that the character of the first break picks along the entire line suggested that there was little, if any, first break energy associated with deeper layers.

Because this near-surface model does not honour any abrupt geological changes such as those associated with the Fraser Fault, the input dataset was divided into two discrete segments before being passed into the REFSTAT2 algorithm. The statics were determined separately for each segment. The first segment contained all first break picks for shots and receivers on the west side of the canyon, and the second contained all picks for shots and receivers on the east side of the canyon. No picks associated with the undershoot portion of the line (i.e., shot and receiver on opposite sides of the canyon) were used in the statics computation, although they were considered in the systematic error analysis described in Section 3.4.
3.2.2 The REFSTAT2 algorithm

We embarked on a thorough investigation of the REFSTAT2 algorithm for two main reasons. First, the procedure we have adopted for identifying systematic errors in the first break picks (Section 3.4) requires an understanding of the algorithm. Second, we noted that the ITA documentation for this popular package is cryptic and includes significant errors of statement. As a result of the latter point, we presented our findings to the Lithoprobe processing community (Perz and Wang, 1992).

The first step in the statics computation is the calculation of the refractor velocity \( V(x) \). REFSTAT2 uses the “abc” method (Coppens, 1985). The calculation of the velocity at the \( k \)th station (position \( x_k \)) involves forming a linear combination of three first break travel times that correspond to certain shot-receiver pairs in the vicinity of \( x_k \). Several different first break “triplets” are considered in the analysis, thereby providing independent velocity estimates. These are then averaged and smoothed to yield a slowly varying final result, \( V(x_k) \).
It is appropriate at this point to introduce the concept of a "delay time". The shot delay time at station $j$, $t_{sj}^d$, is defined as the raypath time between the source and the refractor minus the travel time measured along the normal projection of this raypath onto the refractor (for example see Barry, 1967). A corresponding definition exists for the receiver delay time, $t_{rk}^d$. These definitions assume that the interface between the weathering layer and bedrock is locally planar. With reference to the raypath lengths given in Figure 3.1 these delay times are given by

$$t_{sj}^d = \frac{ac}{V_w} - \frac{bc}{V(x_j)}$$

(3.1)

and

$$t_{rk}^d = \frac{df}{V_w} - \frac{ef}{V(x_k)}$$

(3.2)

By reciprocity, we have the condition that

$$t_{si}^d = t_{ri}^d$$

(3.3)

recalling that there is no uphole time associated with the Vibroseis source. Note that the delay times are "surface-consistent" in the sense that the same station delay time is attributed to any first break arrival associated with a particular site, independent of whether the site contained a shot or receiver, and furthermore, independent of the other site (shot or receiver) to which the first break is coupled.

From the above discussion, $T_{jk}$, the first break traveltime observed at receiver $k$ due to shot $j$ may be approximated by the sum of its associated shot and receiver delay times, plus a term which accounts for the propagation time along the refractor:

$$T_{jk} = p * X_{jk} + t_{sj}^d + t_{rk}^d$$

(3.4)

where $X_{jk}$ is the source-receiver offset, and $p$ is the reciprocal of the spatially-averaged refractor velocity given by

$$p = \frac{1}{\frac{1}{x_k - x_j} \int_{x_j}^{x_k} V(x) \, dx}^{-1}.$$  

(3.5)
Assuming that a sufficiently large number of first break picks is properly distributed over the shot and receiver spaces, equations (3.3) and (3.4) represent an overdetermined full-rank system. The algorithm uses the method of least-squares to solve for the delay times.

If we assume that the shots are located at the surface and the weathering layer-bedrock interface is horizontal, it is easy to show that the delay time at the kth station is related to the local weathering layer thickness \( h(x_k) \) and bedrock velocity \( V(x_k) \) by

\[
t_s k^d = t_r k^d = \frac{h(x_k) \sqrt{1 - \left( \frac{V(x_k)}{V_w} \right)^2}}{V_w}.
\]

Once calculated, the delay times are separated into high and low frequency spatial components by means of a nine-point median filtering operation followed by three-point smoothing. REFSTAT2 then simply inverts (3.6) using the low-frequency component of the delay time to obtain the layer thicknesses at all stations. Note that the filtering operation effectively imposes a smoothness constraint on the layer thickness estimates.

The final step in the process is the actual application of the statics correction. REFSTAT2 explicitly distinguishes between the high and low frequency statics. \( S_{lk} \), the low frequency static component at station \( k \), is given by

\[
S_{lk} = \frac{h(x_k)}{V_{repl}} - \frac{h(x_k)}{V_w},
\]

where \( V_{repl} \) is the user-specified replacement velocity.

As we stated previously, the model described in Figure 3.1 is restricted by smoothness constraints, both in \( V(x) \) and in \( h(x) \). While it is reasonable to expect the bedrock velocity to be a slowly varying quantity, the weathering layer thickness can change significantly.
over the order of only a few group intervals (Hampson and Russell, 1990). Thus, there is a basic error in the model specification. Nonetheless, the short wavelength information is contained in the high frequency component of the delay time, and it is this latter quantity that is considered to be the high frequency component of the refraction statics solution. The total static at a given station is simply the sum of its long and short wavelength constituents.

3.3 Systematic errors

In practice, the first break picks suffer from the usual random observational errors. More significantly, they may be tainted by a component of systematic error. Given the error in the model specification, as well as the presence of both systematic and random errors in the first break traveltime picks, the least squares solution of equation (3.4) with constraints (3.3) is properly considered an estimate of the delay times (the validity of the very concept of the delay time becomes questionable for a true 3–D statics problem). Nevertheless, we shall see that this estimate may be used to identify and remove the systematic errors associated with the first break picks. Using the delay times calculated through a preliminary pass of REFSTAT2, we may model the systematic error by performing a time term decomposition of the error-contaminated first break picks, $T_{jk}^e$:

$$T_{jk}^e = p \cdot X_{jk} + t_j^d + t_k^d + e_{jk}^{syn} + \delta e_{jk},$$

where we have emphasized the surface-consistent property of the shot and receiver delay times by dropping the shot and receiver subscripts:

$$t_{sj}^d = t_{rj}^d \equiv t_j^d;$$

$$t_{rk}^d = t_{sk}^d \equiv t_k^d.$$
In equation (3.8), \( e_{jk}^{\text{sys}} \) is the systematic error whose component parts are identified in Section 3.3.4. \( \delta e_{jk} \) is presumed to be a small random error that embodies any residual misfit due to the imperfect model specification and/or random first break measurement error. Note that even the traveltime decomposition with error terms, equation (3.8), is an approximation. It is quite possible in some instances that the systematic error actually diffuses into the delay time estimates themselves in the least squares fit. Nevertheless, we shall see that equation (3.8) can be successfully used as the basis for a systematic error analysis.

3.3.1 The surface stacking chart as a diagnostic tool in the analysis of systematic error

Milkereit (1989) recommends the use of surface stacking charts (SSC) as a means of quality control and general information display. The SSC is a contour plot in shot-receiver space of trace-specific information (e.g., trace edits, first break picks). The relationship between shot, receiver, midpoint and offset axes (assuming a straight line geometry) on the SSC is illustrated in Figure 3.2. One of his examples is a SSC display of reduced first break traveltime, \( T_{jk} \), defined by

\[
T_{jk} = \frac{X_{jk}^e}{V_{\text{red}}} - \frac{X_{jk}}{V_{\text{red}}},
\]

Here \( V_{\text{red}} \) is a “reducing velocity” which is a constant value that is representative of the regional bedrock refractor velocity. We expect the reduced traveltime to be a rapidly varying quantity in shot-receiver space, because from (3.8) and (3.10), \( \tau_{jk} \) is seen to contain short-wavelength information about the near-surface in the form of the delay times, as well as possible systematic error (neglecting the small random error).
3.3.2 The residual traveltime

The reduced first break plot described in the previous section is useful for highlighting changes in the near-surface and thereby assessing the severity of the statics problem everywhere along the line, but it tends to obscure any systematic errors, these being hard to distinguish from the high (spatial) frequency background. Following a suggestion by Weizhong Wang (personal communication, 1992), we have plotted the residual times in SSC format. We define the residual time, $T_{jk}^{\text{res}}$, by

$$T_{jk}^{\text{res}} = T_{jk} e - \frac{X_{jk}}{V_{\text{red}}} + t_j^d - t_k^d \quad .$$

Substituting equation (3.8) into this last expression gives

$$T_{jk}^{\text{res}} = X_{jk} \ast (p - V_{\text{red}}^{-1}) + e_{jk}^{\text{syn}} + \delta e_{jk} \quad .$$

Figure 3.2. Relationship between the shot, receiver, offset and midpoint axes on a surface stacking chart (SSC).
From (3.12), in the absence of any systematic errors, we expect the residual plot to display a smoothly varying character on a regional scale as it reflects (presumably) gradual lateral changes in the bedrock slowness parameter \( p \), and gradual changes in source-receiver offset. We expect symmetry about the midpoint axis because of the reciprocal property of the residual time \( (T_{jk}^{\text{res}}=T_{kj}^{\text{res}}) \). Superimposed atop the regional trend, we expect a small amount of random noise. For a judicious choice of reducing velocity, we anticipate numerically small values for \( T_{jk}^{\text{res}} \). By design, any systematic error will appear on the plot in the form of an easily recognizable pattern.

### 3.3.3 The reciprocal difference time

In addition to the residual plots, SSC plots of "reciprocal difference time" were created, again following a suggestion by Weizhong Wang (personal communication, 1992). The reciprocal difference time, \( \Delta_{jk} \), is defined by

\[
\Delta_{jk} \equiv T_{jk}^e - T_{kj}^e .
\]

By substituting equation (3.8) into this last expression, we may express the reciprocal difference time in terms of the error terms:

\[
\Delta_{jk} = \left( e_{j,k}^{\text{sys}} - e_{k,j}^{\text{sys}} \right) + \left( \delta e_{j,k} - \delta e_{k,j} \right) .
\]

The residual plots, which we used as the primary tool in the error analysis, were considered in conjunction with the reciprocal difference SSCs.

### 3.3.4 Sources of systematic error

In practice, several sources of systematic error may produce identifiable patterns on the residual plot. A "shot mispositioning" error arises when the vibrator trucks do not shoot exactly on the station. For instance, this may occur if the drivers are unable to locate the survey marker at a particular station because it is obscured by the roadside.
vegetation, and as a result, drive a few metres beyond the actual shot station. Then forward receivers record a first break that is systematically “early” by some fixed amount that corresponds to the portion of “missing” raypath between the actual and expected shot location, and similarly, reverse receivers detect a first break that is systematically “late”. The anomalous first break pattern will appear on the residual plot along a locus of constant shot, and is readily detected by the human eye.

When a receiver site suffers from a severe local statics problem, or when a receiver group is affected by a polarity reversal, the corresponding first break appears time-shifted on a shot gather compared to neighbouring first breaks. First break picks used in the refraction statics analysis were made on shot gathers using the automatic picker of the ITA interactive editor “VAQ2”. The cross-correlation based picker “tunes” the user’s initial crude estimate of the first breaks to the local dominant peaks along the first break event. Generally, the picker does a good job; however, it often cycle-skips over these time-shifted first breaks. Although VAQ2 provides a visual display of the tuned picks, the plotting scale is such that single cycle-skips often go undetected. This “tuning error” is manifest on the residual plot as an anomaly along a locus of constant receiver. Note that the fact that the statics are particularly problematic at a specific site underscores the importance of an accurate first break travelt ime measurement.

Another systematic error results when the side lobes of the Klauder wavelet render the detection of the dominant first break peak ambiguous. A side lobe may be erroneously identified as a first break peak. Typically, this “side lobe error” appears on the residual plot as an anomaly along a locus of constant shot, because the picking is performed on shot gathers.

A “recorder error” results when there is a zero-time discrepancy between the master and slave recorders. Because the slave recorder records the signal from the first 120
channels and the master records channels 121–240 (Figure 2.1), this error appears as a bulk time shift along the constant channel locus corresponding to channel 121. This type of error plagued the data from the Lithoprobe Kapuskasing Structural Zone (KSZ) transect (Weizhong Wang, personal communication, 1993), but as we shall see, was not a problem for SBC18.

Finally, "geometry errors" result when the geometry information is incorrectly entered into the trace headers. Because of the regular plotting patterns associated with all three modes of shooting (roll-on, roll-along, and roll-off) on the SSC, such errors are readily detected.

3.3.5 The residual and reciprocal difference plotting algorithms

The delay times were calculated by REFSTAT2 on a preliminary run using the error-contaminated picks. We wrote a FORTRAN program to calculate the residual and reciprocal difference values along the line using equations (3.11) and (3.13). A value of 5600 m/s, representative of the average bedrock velocity observed throughout the region, was chosen for the replacement velocity used in calculating the residual traveltimes. Because the shot interval for SBC18 is twice the receiver interval, it was necessary to incorporate a data interpolation scheme into our plotting algorithm in order to "fill in" the gaps in data coverage on the SSCs. In general, shooting took place on the odd station numbers, but there were many exceptions to this due to culverts, rough spots along the road, bridges, observer error and slide areas.

The first break travelt ime corresponding to a "missing" shot at a given source-receiver offset was estimated by forming a weighted average of two first break traveltimes, namely the one due to the nearest shot in the forward direction measured at the receiver of the same (nominal) offset and its counterpart in the reverse shot direction. Suppose that the station at position $x_i$ was not a shotpoint, and that the nearest shotpoints in the forward
and reverse directions were at sites $x_{i+ifor}$ and $x_{i-irev}$, respectively, where $ifor$ and $irev$ are positive integers. We estimated $T_{lm}^{est}$, the first break arrival time corresponding to the missing shot and the $m$th receiver, by

$$T_{lm}^{est} = \frac{irev \cdot T_{i+ifor \, m+ifor} + ifor \cdot T_{i-irev \, m-irev}}{ifor + irev}.$$  \hspace{1cm} (3.15)

Equation (3.15) gives the general method of interpolation; however geometrical complications associated with the roll-on and roll-off portions of the line necessitated the use of a slightly different scheme for estimating traveltimes near the ends of the spread. Also, we modified the interpolating process in the vicinity of the Fraser Canyon. Near the canyon edge the use of (3.15) would yield erroneous estimates.

Once calculated by the algorithm, the residual and reciprocal difference times were contoured and colour-plotted.

From equation (3.13), note that

$$\Delta_{jk} = -\Delta_{kj}.$$  \hspace{1cm} (3.16)

Because of this symmetry, all reciprocal difference information can be plotted on one side or other of the midpoint axis $i = j$ (Figure 3.3). To avoid redundancy in our information display, we plotted $\Delta_{ij}$ for $i \neq j$ and the arithmetic mean of $T_{ij}^{res}$ and $T_{ji}^{res}$ for $i < j$ (recall that residual times are reciprocal in the absence of any systematic error), hoping that this latter quantity would provide a useful reference or “tie-in” with the accompanying residual plots. Under this plotting convention, any type of systematic error that is specific to a single shot or receiver (such as those described in Section 3.3.4 below) is manifest on the reciprocal difference plot as an anomaly that follows an inverted “L” shape (Figure 3.3).

3.3.6 Result presentation and discussion

The very small master-slave recorder discrepancy revealed in Figure 3.4a is the only visible recorder error observed over the entire line. The insignificant time shift between
channels 120 and 121 is indicated on the figure. One of our initial concerns was that a recorder error might have occurred for the undershoot, when the master and slave recorder trucks were located on opposite sides of the canyon. Such an error would not be discernable on the residual plot because the geological discontinuity across the canyon coincides with the gap between channels 120 and 121 and would obscure any time shift. However, from the reciprocal difference plot (Figure 3.4b), we infer that there was negligible recorder error, since such an error would result in non-zero reciprocal difference times in the portion of shot-receiver space delineated by dashed box B in the figure. The zero-time discrepancy between the master and slave recorders could conceivably affect a significant proportion of the total number of traces along the line. The SSC plots, in revealing that such an error was negligible on SBC18, provided a reassurance of the excellence in data acquisition.
Figure 3.4a. Residual traveltime plot for VPs 27-199 (following page). A small recorder error is highlighted by the dashed box. To see, take a line of sight along the dashed arrow. The solid arrows indicate the transition between the master and slave recorders. The boldface dashed line indicates the transition between roll-on and roll-along shooting configurations. A polarity reversal at receiver site 248 shows up on this figure as a linear (vertical) anomaly. The inset displays a portion of the shot-receiver space near station 248 after correction of the polarity reversal. Note that the inset displays additional infilling of colour near the right edge of the plot compared with the uncorrected result, reflecting the fact that during the correction process, additional picks were made using the interactive editor (white areas within the shot-receiver space indicate an absence of first break picks).

Figure 3.4b. Reciprocal difference time plot for VPs near the Fraser Canyon (following page). The plotting convention is illustrated in Figure 3.3. The shot-receiver space is divided into three separate regions, A, B and C, each of which is associated with a different acquisition geometry. In particular region B, which corresponds to the undershoot, displays near-zero reciprocal difference times, suggesting that there was negligible recorder error when the master and slave recorders were located on opposite sides of the canyon.
A polarity reversal is also indicated in Figure 3.4a. Correction was performed using the ITA interactive editor “VAQ2” (all trace amplitudes were multiplied by a factor of \(-1\)). The corrected result is displayed in the figure inset.

A tuning error associated with receiver site 587 is indicated in Figures 3.5a (event A) and 3.5b (dashed boxes). From this latter figure, we note the anomaly is clearly characterized by the inverted “L” signature illustrated schematically in Figure 3.3. One limb of the “L” is systematically “high” whereas the other is systematically “low”. A careful inspection of the first break picks at this site revealed a severe local statics problem that had caused the automatic picker to cycle-skip over the correct first break traveltime. From the residual plots (Figures 3.5a and/or 3.5c), we infer that here, the statics are particularly severe. This is evident from the relatively rapid variation in residual traveltime observed throughout the shot-receiver space, even in areas that do not suffer from any systematic error. We recall from Section 3.3.2 that in the absence of systematic errors, the residual traveltime is generally a smoothly varying quantity. Rapid residual traveltime variation indicates (i) highly variable bedrock velocities or (ii) poorly-estimated delay-times. Either possibility points to an extremely heterogeneous near-surface. The error, which plagued several receiver sites along this portion of the line, was corrected by careful manual repicking. The corrected picks for station 587 are shown in Figures 3.5c and 3.5d.

A representative example of a shot mispositioning error is shown on the residual plot in Figure 3.5a (event B). Note that the residual traveltimes associated with the “reverse” receivers (receiver indices less than shot index) are systematically “high” compared to the background, while the forward residual traveltimes are systematically “low”. The error is also shown on the corresponding reciprocal difference plot in Figure 3.5b (solid boxes). As in the case of the recorder error, this type of error was not particularly problematic for SBC18. In fact the portion of the line displayed in Figure 3.5 (near the western edge
Figure 3.5. Residual traveltime and reciprocal difference time plots for VPs 400-641 located near the western edge of the Fraser Canyon (following page). The colour time scale for all four plots is shown in Figure 3.5b. Results are displayed both before and after correction of systematic errors. (a) Residual traveltime plot before correction. Event A is a tuning error. Event B is a shot mispositioning error. (b) Reciprocal difference time plot before correction. Here, the tuning error (event A in (a)) is indicated by the dashed boxes. The solid boxes highlight the shot mispositioning error (event B in (a)). (c) Residual traveltime plot after correction. Compare events A and B in (a) with the corresponding regions in this figure. The stippled pattern associated with the correction of the tuning error at receiver site 587 is an interpolating artifact. The solid box near the top of the figure indicates the zoom area shown in expanded view in Figure 3.6. (d) Reciprocal difference time plot after correction. Compare the boxed regions in (b) with the corresponding regions in this figure.
of the Fraser Canyon) is the only section which suffers from such errors. Because of the rapid near-surface variation, a departure from the surveyed shot site on the order of only a few metres could induce an appreciable change in first break arrival time.

Positive identification of this type of error was difficult because of the near-surface heterogeneity. In all, five systematic anomalies were attributed to shot mispositionings and were subsequently corrected. Only the first break traveltimes were corrected. Shot mispositioning corrections to reflection arrival times are negligibly small for all but the shallowest horizons. We adjusted the first breaks for VPs 513 and 511 by adding 10 ms to those traveltimes measured at forward receivers and subtracting 7 ms from those measured at reverse receivers (the forward and reverse corrections need not necessarily have the same magnitude in the case of an extremely heterogeneous subsurface). For VPs 547, 549 and 569, 9 ms was added to forward receiver traveltimes and 8 ms was subtracted from the reverse receiver traveltimes. The amount of correction was determined visually, by simply inspecting and comparing the residual and reciprocal difference plots before and after adjustment (the correct adjustment ideally restores the anomalous region to the background colours). The corrected picks are displayed in Figures 3.5c and 3.5d.

Note that the correction of both shot mispositioning errors and tuning errors due to severe local static anomalies led to interpolation artifacts on the final plots. Figure 3.6 is a zoom of the boxed portion (solid box) of the corrected residual plot of Figure 3.5c. The expanded view shows clearly that the points in shot-receiver space corresponding to real VPs are properly corrected. The stippled pattern presumably arises because equation (3.15) does not serve as a good interpolator in the case of severe local static anomalies.

A side lobe picking error is highlighted in the residual plot of Figure 3.7a (dashed boxes). Very strong first arrivals were recorded here, resulting in high amplitude Klauder wavelet side lobes. This type of error occurred at several locations along the line. In
Figure 3.6. Expanded view of solid box appearing in Fig. 3.5c. It is clear from this figure that the stippled pattern associated with receiver site 587 is an interpolating artifact.
each case, a careful manual repicking of the first break traveltimes rectified the situation (e.g., Figure 3.7b).

Figure 3.8a (residual plot) displays the sole geometry error that occurred on SBC18 (dashed box). The error was due to a data entry error we committed while transcribing the Sonix Exploration observer notes to ITA format. The corrected result is shown in Figure 3.8b.

3.4 Elevation correction

Once the systematic errors were identified and removed through the use of the SSCs, the corrected dataset was divided into two segments in the manner described in Section 3.2.1. The segments were input to the REFSTAT2 algorithm on two independent runs, and a final solution for the statics for all the stations on SBC18 was obtained. An examination of equation (3.7), the expression for the low frequency static component, reveals that REFSTAT2 does not refer the corrections to a specific datum plane. To see this consult Figure 3.9. In the figure $z(x)$ is defined as the distance between the surface and the datum plane; thus at station $k$

$$z(x_k) = e(x_k) - d,$$  \hspace{1cm} (3.17)

where $d$ is the datum plane elevation.

We seek to “replace” the near-surface structure by a medium of uniform velocity, $V_{repl}$, and constant surface topography, $d$. A straightforward analysis based on examination of Figure 3.9 reveals that the statics correction $S_k$ (elevation and low-frequency refraction statics) is calculated according to

$$S_k = \left( \frac{h(x_k) - z(x_k)}{V_{repl}} - \frac{h(x_k)}{V_w} \right).$$  \hspace{1cm} (3.18)

But the conventional elevation static at the $k$th station, $a_k$, is given by

$$a_k = - \frac{z(x_k)}{V_{repl}}.$$  \hspace{1cm} (3.19)
Figure 3.7. Residual plot for VPs 721-800 (following page; top). Ordinate is shot index and abscissa is receiver index. (a) Before correction. Side lobe picking errors are indicated by the dashed boxes. (b) After correction.

Figure 3.8. Residual plot for VPs 1000-1050 (following page; bottom). Ordinate is shot index and abscissa is receiver index. (a) Before correction. Note the geometry error (dashed box). (b) After correction.
Figure 3.9. Relationship between the datum plane and the refraction statics model.

From equations (3.18) and (3.19), together with (3.7) (which describes the low frequency statics computation), it is clear that the complete statics solution requires an explicit elevation correction. We performed the correction by means of the ITA module “DATM” (again, using a replacement velocity of 5600 m/s).

3.5 Data results after statics application—stacked sections

In order to assess the overall merit of the application of refraction statics, we designed a “brute stack” processing stream, which aside from its statics treatment was essentially identical to the one used by Western in the preliminary processing: geometric spreading correction, predictive deconvolution, refraction statics, CMP sorting, NMO correction using the same stacking velocity file, prestack and poststack trace balancing. From our discussion on deconvolution (Section 2.5), we recall one other processing difference, namely the discrepancy between prediction gap parameters, 24 ms (Western) versus 33 ms (our reprocessing). This accounts for the slightly higher frequency content of the Western sections here and in the data comparisons in subsequent chapters. We have displayed a portion of Western’s stacked section taken from near the western edge of the
profile in Figure 3.10a and present our corresponding brute stack in Figure 3.10b. With regard to this latter figure, we note that CMP station numbers have been multiplied by a factor of ten (ITA axis-labelling convention). This plotting convention is adopted in all subsequent figures which display our stacked sections. Enhanced continuity of reflectors is apparent in several regions of Figure 3.10b, particularly between 9.6–10.0 s.

The only truly disappointing result occurs in the vicinity of the western edge of the canyon, between midpoints 450 and 550. Here, the Western stacked section revealed virtually no signal. Visual inspection of the shot gathers showed that the S/N ratio was very low in this area (Section 2.4); thus, we expect that the residual statics algorithm used in the contract processing suffered from cycle-skipping problems. Although the first break picks were noisy, they were nonetheless discernible, so we had hoped that application of refraction statics would help to eliminate the severe statics problem in this region (we recall that the residual plots in Figure 3.5 suggest rapid near-surface variation). Unfortunately, the application of refraction statics was unable to draw out any signal. We attribute the absence of observed reflectivity to the severe noise problem in the area, and not necessarily to the subsurface geology.
Figure 3.10. Comparison of stacked sections after application of statics. Processing streams for the two sections are essentially identical except for statics (see text for details). (a) Western stacked section. (b) Our reprocessed brute stack.
Chapter 4
DIP MOVEOUT CORRECTION AND VELOCITY ANALYSIS

Compensation for the effect of source-receiver offset on observed reflection travel-times is conventionally achieved through application of the "normal moveout" (NMO) correction. However in the case of a dipping interface, NMO correction does not restore the associated reflection hyperbola to zero-offset time; rather it overcorrects, resulting in "too early" a reflection time. Dip moveout correction (DMO) removes the effect of reflector dip from non-zero-offset arrivals, thereby improving the stack quality for steeply dipping events.

In this chapter, we describe the theory which underlies the DMO algorithm and the practical considerations which arise when the method is applied to the crooked line problem. Because of its intimate relation to DMO, we also describe our velocity analysis.

4.1 DMO theory

The traveltime expression for a single, horizontal, uniform layer over a halfspace is the familiar normal moveout (NMO) equation

\[ t^2 = t_o^2 + \frac{4h^2}{V^2}, \quad (4.1) \]

where \(t_o\) is the two-way traveltime measured at zero offset, \(V\) is the layer velocity, and \(h\) is the half-offset. If we now allow this layer to dip, say at an angle \(\theta\) from the horizontal, (Fig. 4.1) then the corresponding traveltime equation is given by

\[ t^2 = t_o^2 + \frac{4h^2 \cos^2 \theta}{V^2} \quad (4.2) \]

(Dix, 1955).
From equation (4.2) we see the problem associated with performing a standard correction to normal moveout time, $t_n$, given by

$$t_n^2 = t^2 - \frac{4h^2}{V^2},$$

since $t_n < t_o$. The overcorrection due to application of NMO obviously misaligns events within a CMP gather, thereby impairing stack quality.

An optimum stack would be obtained if the NMO correction were performed using a "stacking" velocity $\frac{V}{\cos \theta}$. This observation suggests that we directly incorporate the moveout correction implied by equation (4.2) into the velocity analysis; in other words, we might expressly choose "too high" a velocity in the interest of optimizing the stack quality. There are several reasons why such a procedure is undesirable from a processing viewpoint. For instance, we require knowledge of the true subsurface velocity structure to perform a wave-equation migration (and the associated time-to-depth conversion). Moreover, one objective in seismic reflection processing, though ancillary to the delineation of geological features on a stacked section, is the acquisition of information about the P-wave velocity structure. Additionally, there is an ambiguity
concerning the choice of stacking velocity in the case of multiple dipping reflectors. In other words, which value of $\theta$ do we choose if there are several events, each associated with a different interface and therefore a different dip, that intersect (or at least are in proximity to one another) at a certain time $t$ and half-offset $h$ on the input CMP gather?

Operating in the common-offset domain (Fig. 4.2), Hale (1984) devised a scheme to remove the effect of reflector dip without prior knowledge of the dip angle by Fourier transforming the input NMO-corrected dataset. Although computationally expensive, the method now known simply as “DMO” is capable of performing a simultaneous correction for all dips. This correction, most significant for early times at large offsets and steep dips, restores the input common-offset sections to zero-offset sections. In an alternate interpretation of the action of DMO, Deregowski (1982) shows that the method effectively eliminates “reflector point dispersal” on a CMP gather by “migrating” energy associated with a particular reflection point on a common offset section updip to the midpoint that is associated with normal incidence to that reflection point (Fig. 4.3). For this reason, DMO is also referred to as “prestack partial migration”. Although the method assumes a constant velocity, empirical studies suggest that DMO works well on real data.

4.2 DMO—practical considerations

4.2.1 The time-variant DMO impulse response: synthetic tests

ITA has implemented the algorithm of Liner and Bleistein (1988), who use a logarithmic stretch of the time axis to perform a computationally efficient DMO correction, in the poststack module “DMO2F”. In our first synthetic test, we seek to recover the “DMO ellipse”, which is properly viewed as the DMO time-variant impulse response. Such a test serves to “calibrate” the DMO2F algorithm and verifies that we are correctly interpreting the input parameters.
Figure 4.2. The dataset occupies a volume in midpoint-offset-time space. There is no loss in generality in using the half-offset, instead of the offset, to describe the data space. For ease of display, only positive offsets are indicated in the figure. Under this data representation, both common midpoint and common offset gathers map to planes.

Figure 4.3. The elimination of reflector point dispersal by application of DMO. The energy associated with the reflection point \( P \) is initially "incorrectly" associated with the actual source-receiver midpoint \( y_i \). After DMO correction, the energy reappears in the common offset section at the trace associated with the "correct" midpoint \( y_f \), at time \( t=2D_o/V \).

Deregowski and Rocca (1981) show that a single impulse occurring on a common-offset section at a point \((t_p, y_p)\) may be associated with an elliptical reflector embedded
in a homogeneous earth whose shape is given by

\[
\frac{(y - y_p)^2}{\left(\frac{v^2 t_p^2}{4}\right)} + \frac{(2\xi)^2}{\left(t_p^2 - \frac{4h^2}{v^2}\right)} = 1 ,
\]  

(4.4)

where \( y \) is the midpoint position, \( h \) is the half offset, \( v \) is the velocity and \( z \) is the depth measured at \( y=y_p \).

Figure 4.4a is the input common offset section (half offset \( h=1200 \) m, midpoint spacing \( \Delta y=25 \) m). This section was produced by convolving a Klauder wavelet with five impulses on trace 70, these being equally spaced at 0.5 sec intervals and spanning a time window from 1.4–3.4 s. Figure 4.4b shows the same section after NMO correction using a velocity of 2000 m/s (i.e., the assumed medium velocity). This NMO-corrected section was input to the INSIGHT algorithm DMO2F. Figure 4.4c shows the DMO-corrected result. The coherent features on this figure are not the elliptical reflectors, but rather their corresponding zero-offset events, which still require migration. Figure 4.4d displays the migrated result. A simple conversion from two-way migrated time, \( t_m \) to depth (\( t_m=2z/v \)) allows a verification of the lengths of the semi-major and semi-minor axes, \( a \) and \( b \), respectively, for the first pulse at 1.4 s on Figure 4.4a. According to equation (4.4),

\[
a = \sqrt{\frac{v^2 t_p^2}{4}} = \sqrt{\frac{(2000^2)(1.4^2)}{4}} = 1400 \text{ m} = 56 \text{ traces} ,
\]  

(4.5)

and

\[
b = \sqrt{t_p^2 - \frac{4h^2}{v^2}} = \sqrt{1.4^2 - 4 \left(\frac{1200^2}{2000^2}\right)} = 0.721 \text{ s} .
\]  

(4.6)

The lengths are in agreement with Figure 4.4d.

4.2.2 Spatial aliasing considerations

The "energy smearing" action of DMO, which moves energy from a single midpoint to neighbouring ones on a common offset gather, is clearly illustrated by the impulse
Figure 4.4. DMO impulse response testing: (a) the input common offset section; (b) the NMO-corrected section (\(v=2000\) m/s); (c) the DMO-corrected section; (d) result of migrating (c). Note that the "smiles" appearing at later times on (d) are migration artifacts associated with edge effects.
responses in Figure 4.4c. We anticipate such "smearing" because from the migration-like action of DMO. Intuitively we might expect that DMO can act as an "interpolator" in the following sense: if the traces of an input common offset gather were interspersed with zero traces, then DMO would restore continuity of reflectors by shifting energy into the dead traces. This "interpolating property" is highly desirable from a practical viewpoint (Ronen and Liner, 1987; Bolondi et al., 1982; Deregowski, 1986). Since it permits, in effect, a finer sampling of midpoint space on the stacked section, beyond the obvious consideration of improving the final result from an interpretative viewpoint, DMO acts as a safeguard against the phenomenon of spatial aliasing during the application of subsequent processes (such as migration) which involve a 2-D Fourier transform from \( x-t \) (space-time) to \( f-k \) (frequency-wavenumber) coordinates.

The DMO process itself, being predicated on the well-known fact that all linear events in \( x-t \) space having a particular slope map to the same radial line in the \( f-k \) domain, may be susceptible to aliasing problems. Fortunately, we need not exploit the anti-aliasing feature of DMO in our present analysis. For a constant velocity medium, it is easy to show that the maximum slope of a reflection event on a constant-offset section is given by

\[
\left( \frac{dt}{dx} \right)_{\text{max}} = \frac{2}{V}.
\]  

(4.7)

This result, which corresponds to a \( 90^\circ \) dipping interface, is offset-independent. Taking \( V = 5000 \text{ m/s} \) as a lower bound on the velocity for SBC18, we may use equation (4.7) to calculate an upper bound on the time dip that we expect to see on the SBC18 constant offset sections:

\[
\left( \frac{dt}{dx} \right)_{\text{max}} = 4 \times 10^{-4} \text{ s/m}.
\]  

(4.8)

The frequency associated with the onset of aliasing, \( f_{\text{max}} \) is related to the spatial sampling interval \( \Delta x \) by

\[
f_{\text{max}} = \frac{1}{2\Delta x} \left[ \left( \frac{dt}{dx} \right)_{\text{max}} \right]^{-1}.
\]  

(4.9)
The spatial sampling interval on a common-offset section is simply the midpoint interval, which in our case is 25 m. Hence, from equations (4.8) and (4.9), we find that the threshold aliasing frequency for SBC18 is 50 Hz. Given the characteristics of the source sweep (frequency range 10–56 Hz), it is unlikely that DMO-induced aliasing effects are significant for SBC18.

Despite having been spared the problem of spatial aliasing for our present processing task, we decided to test firsthand the anti-aliasing characteristics of DMO. Ronen (1985) describes the theoretically rigorous procedure for the exploitation of the DMO anti-aliasing property. Here, we content ourselves with a less-sophisticated approach. We consider a synthetic model consisting of two steeply dipping reflectors. The corresponding zero-offset section is shown in Figure 4.5a. Next, we construct a synthetic common-offset section \( h = 1200 \) m. Note that the resulting time dip on this section (Fig. 4.5b) is well beyond \( \left( \frac{dt}{dx} \right)_{max} \), given the frequency content of the Klauder wavelet used in producing the synthetic. Figure 4.5c shows the result of applying DMO to the NMO-corrected version of the section (NMO was performed using the same velocity that was used in constructing the synthetic earth model). Spatial aliasing has caused a drastic mispositioning of the reflection event (e.g. white box in Figure 4.5c); this is evident from a comparison between the linear coherent energy within the white box and the corresponding event on the zero-offset section (Figure 4.5a), which represents the DMO ideal output. Next, we interspersed the traces in Figure 4.5a with zero traces, thereby doubling the size of the input gather. Figure 4.5d shows the result after DMO. We note that the “smily” artifacts present in Figures 4.5c and 4.5d are a manifestation of the superposition of the various DMO impulse responses (we performed similar synthetic tests for much shallower (unaliased) dips, and noted in such cases that the smiles interfered destructively everywhere, except along the reflection events).
It should be emphasized that our method is grounded only in intuition, and that we have chosen for simplicity the crudest of interpolation schemes. Infilling using the average of adjacent traces on a common offset gather would undoubtedly improve the results. We feel that the results of our simple test are worth mentioning. They bring to light a characteristic of DMO which is not universally recognized by the crustal seismic processing community but which, at least to a first-order approximation, may be readily exploited.

4.2.3 The partial stack

DMO correction is based on a two-dimensional theory. Although we hope to gain some improvement in the imaging of dipping reflectors through its application, we cannot directly port the method over to our three-dimensional situation. A complication arises because of the irregular offset distribution associated with a crooked line geometry. With reference to Figure 4.2, we do not expect more than one or two traces to map to a given common offset plane, and further, we expect that a very large number of such planes would be required to describe the entire collection of traces over the line. This is highly undesirable from a practical viewpoint, since the DMO algorithm operates separately on each common offset gather.

We need to group the traces according to some measure of nominal offset. The procedure we follow is analogous to the approach we adopted in performing the CMP binning. "Common offset binning" effectively decreases the number of common offset planes and increases the fold (i.e., the number of traces) within each common offset gather. Operating in the CMP domain, the procedure groups the traces into regularly-spaced offset bins (Figure 4.6). The bin assignment is based on the true source-receiver offset. In the case that several traces occupy the same bin, the arithmetic mean of the trace amplitudes is calculated and assigned to that bin. Since this averaging process involves
Figure 4.5. Testing DMO antialiasing properties: (a) zero-offset section for a synthetic earth model consisting of two steeply dipping reflectors; (b) constant offset section (half-offset=1200 m); (c) application of DMO without prior insertion of dead (zero) traces; (d) same as (c) except that dead traces were inserted prior to applying DMO. Note the mispositioning of the reflector in (c) (white box) due to spatial aliasing. By contrast, the same reflector in (d) (white box) has been restored to the correct zero-offset location in (x-t) space (compare (c) and (d) with the DMO ideal in (a)).
a summation or "stack" in the CMP domain, the common offset binning procedure is referred to as the "partial stack".

The partial stack has been implemented by ITA through the pre-stack module "PSTK". In the case that no traces occupy a particular bin, PSTK introduces a single zero (dead) trace. Although we do not expect the DMO process, given its energy smearing property, to be significantly affected by this artificial "seeding" of dead traces, we can improve the partial stack by infilling the empty bin with a trace formed by taking the mean of the two traces that occupy the surrounding (adjacent) bins, then discarding the seeded trace after applying DMO.

A second improvement to the ITA algorithm may be realized by recognizing that the DMO correction does not depend on the polarity of the offset, rather only on its absolute value (see equation (4.2)). Moreover, two traces in a single CMP gather having the same absolute value of offset but opposite polarity ought to be identical by reciprocity. While the PSTK module maintains offset polarity, the above considerations imply that we may bin the traces according to absolute value of offset. This has the effect of reducing the number of common-offset gathers input to the DMO algorithm, and therefore minimizing the run time. As well, the increase in the number of traces that contribute to the partial stack performed within each bin may help increase the signal-to-noise ratio along noisy portions of the line.

Weizhong Wang (personal communication, 1992) has modified the PSTK algorithm, incorporating these two improvements. We used his FORTRAN subroutine in designing our partial stacking algorithm, which we have illustrated schematically in Figure 4.6.

4.2.4 DMO and velocity analysis

We have seen that the complete compensation for reflection event moveout on a CMP gather may be achieved via two cascaded processes: NMO correction followed by
Figure 4.6. The partial stacking procedure: (a) the zoom box is defined on a common midpoint plane; (b) expanded view of zoom box together with definition of the offset bin; (c) result of applying partial stack to (b). Traces D, E (which lie in the same bin in (b)) and C (which, on the basis of absolute value of offset, also lies in this bin) are stacked together to form trace G. Although no traces in (b) lie within the bin corresponding to trace H, this trace is formed by taking the mean of its non-zero neighbour traces in (c).
DMO. The NMO correction requires *a priori* knowledge of the P-wave velocity field; however, such information is generally unavailable to the processor. The process of velocity estimation or "velocity analysis" entails performing the full moveout correction (NMO + DMO) for a number of different trial velocity profiles ("profile" here refers to a set of time-velocity pairs) and choosing an optimum one based on a certain coherency criterion. This velocity analysis procedure, summarized in flow chart format in Figure 4.7a, is prohibitively expensive from a computational viewpoint since DMO run-times are typically an order of magnitude larger than their NMO counterparts. We would prefer to interchange the order of the DMO and the NMO processes as they appear in Figure 4.7a, apply the DMO correction once, and iterate through only the NMO correction in performing the analysis. Hale (1984) shows that NMO and DMO do not commute, but that there exists a reasonable approximation to the theoretically accurate process of NMO followed by DMO, namely an *approximate* NMO correction, followed by DMO and a subsequent *residual* NMO correction. Conceptually, residual NMO is equivalent to removing the approximate NMO, then performing the true NMO correction. This suggests a practical scheme for velocity analysis (Figure 4.7b).

Of the many coherency measures proposed over the last twenty years for velocity analysis, the "semblance" criterion (Neidell and Taner, 1971) is the most commonly used by the exploration industry. Basically, the method involves fitting a set of normal moveout hyperbolae to an input CMP gather. Unfortunately, it is unsuitable for use on SBC18, and may have limited application in deep crustal reflection processing in general. There are several reasons for this. The signal-to-noise ratio is typically very low in the deep crustal regime, so the method encounters difficulties in event detection. Also, the velocities are sufficiently high that very little moveout is observed over the spreadlengths typical of most deep crustal experiments and there is an ambiguity in velocity resolution. In other words, all the deep events appear "flat" over the range of observation. Yet
another problem is the fact that deep reflectors typically lack substantial continuity, and as such, may not be detected by all the source-receiver pairs within a CMP gather.

Given the limitations of semblance, we have chosen to estimate velocities by means of the “constant velocity stack” (CVS). This method entails the simultaneous analysis of a series of test stack panels. Each panel is created by first performing the residual NMO correction using a constant trial velocity (Figure 4.7b), then stacking (see, for example, Yilmaz, 1987). The panels are displayed side-by-side on the graphics monitor, and the velocities chosen simply on the basis of the optimum stack. By performing the CVS analysis at regularly spaced intervals along the line, we can account for lateral velocity variations.

Having obtained a set of stacking velocity profiles along the line, the question remains as to their physical interpretation. When the spreadlength is small compared to the depth, the reflection event moveout for a horizontally-layered one-dimensional earth is governed
by the rms velocity, $V_{rms}$, given by

$$V_{rms}^2(t) = \frac{1}{t} \int_0^t v^2(u) \, du$$  \hspace{1cm} (4.10)

(e.g., Yilmaz, 1987). Of course the situation with real data is much more complicated than the simple model used in their analysis, but to a first order approximation, we can consider our set of stacking velocity profiles to be a measure of the RMS velocity field along the line.

### 4.3 The DMO/velocity analysis processing stream

The theoretical and practical considerations discussed in this chapter were incorporated into our DMO/velocity analysis processing stream (Figure 4.7b). The input dataset consisted of static-corrected CMP gathers. We constructed a single velocity-time profile representative of the regional velocity field, based on velocity information acquired from the preliminary processing. An approximate NMO-correction was then performed for the entire line using the ITA pre-stack module “NMO2”. This module was used to perform all NMO corrections in the subsequent processing, as well. It should be noted that the algorithm automatically mutes portions of the dataset that are NMO-stretched beyond a default threshold value (for a description of the phenomenon of NMO-stretching, see Yilmaz, 1987).

Next, we performed the partial stack. We selected an offset bin width (Figure 4.6b) of 200 m (the nominal offset increment observed in a CMP gather) and performed the partial stack over an offset range of $-8000$ m to $+8000$ m (this offset range was modified appropriately for the roll-on and roll-off sections of the line).

Next, we applied DMO using the ITA poststack module DMO2F. The algorithm sorts the partially-stacked CMP gathers to the common offset domain before proceeding with the DMO correction. Since it allows the specification of a minimum velocity, $V_{min}$,
the algorithm effectively acts as a dip filter, excluding all dips greater than \(2 / V_{\text{min}}\) (see equation (4.7)) from the analysis. Hence, we hope to benefit from the removal of any (unaliased) coherent noise whose common offset signature is of a steeply-dipping linear nature. Because of the large computational expense and the fact that the DMO correction is not significant at later times, only the first five seconds of data were input to the routine. The DMO correction was performed piecewise, since the algorithm could not process more than (approximately) 250 CMP gathers simultaneously. To avoid any edge effects that might be produced by the DMO algorithm, we allowed an overlap of 25 CMPs in constructing these input data segments. Even operating on a single truncated dataset (i.e., 250 CMP gathers, 0–5 sec), the algorithm required roughly 8 hours of run-time.

The approximate NMO correction was then removed by use of the prestack module "RNMO", and the CVS velocity analysis performed. Each test stack panel consisted of 50 midpoints. The analysis was repeated at 1.5 km intervals along the line to account for lateral velocity variation. The results obtained from noisy sections of the line, particularly between midpoints 4400 and 5800, were considered unreliable and were not used in constructing the final stacking velocity model. We noted that the DMO process actually degraded the stack quality along particularly crooked sections of the line. In these regions, velocities were estimated using only NMO correction. Below about 5 seconds, we noted that the quality of stack was insensitive to the choice of stacking velocity. A similar observation was made in the preliminary processing (Hostetler, 1989). As a result, we did not perform a detailed velocity analysis on the deeper portion of the dataset; rather, we used the stacking velocity information available from the preliminary processing.

4.4 Data results—stacked sections after DMO

Generally speaking, application of DMO yielded modest improvement. An example of a significantly enhanced dipping feature near the western edge of the line is displayed
in the stacked section of Figure 4.8b. For purposes of comparison, we show the corresponding portion of Western's final stack in Figure 4.8a. The improvement in the imaging of the east-dipping reflector is attributable to the DMO processing, and not to refraction statics, since the feature was barely visible on the brute stack (the brute stack processing flow is described in Section 3.5).

On the east side of the Fraser Canyon, DMO has apparently succeeded in providing an image of the Pasayten fault, a northeast-dipping thrust fault of regional scale. From a tectonic viewpoint, this represents one of the major accomplishments of the reprocessing (Chapter 8).
Figure 4.8. Comparison of Western stacked section and our section after application of DMO. (a) Western final stacked section near west edge of line. (b) Our result after application of DMO (otherwise same processing sequence as brute stack). Note the enhanced continuity of dipping events (white boxes, (a) and (b)) in our result.
Chapter 5
CROSSDIP CORRECTION

5.1 Introduction

Up to this point, our analysis has been based on the assumption of a straight line acquisition geometry. Central to this "pseudo-two-dimensional" treatment is the concept of the CMP bin. Obviously the many-to-one mapping, through which several scattered midpoints are assigned to the same effective midpoint for the purposes of subsequent processing, will impair the quality of the output stack compared to its true 2-D counterpart (i.e., the stack we would have obtained had the line been straight). Two questions require addressing: first, how might we compensate for the error introduced by the CMP binning approach; and second, having performed such a correction, is it possible to exploit the line crookedness to provide three-dimensional information about the subsurface? Before seeking the answers, we must understand the principal error source, the reflector "crossdip". Crossdip refers to the component of dip that is perpendicular to the slalom line.

In the first part of the chapter, we present the crossdip correction in the context of the general 3-D traveltime expression that we derived for a crooked-line survey conducted over a single constant-velocity dipping layer overlying a halfspace. Then, we present a method that we have developed for estimating the size of the correction. Finally, we describe the incorporation of the crossdip correction into our processing stream.

5.2 The 3-D CMP bin traveltime equation for a uniform dipping layer over a halfspace

In order to quantify the midpoint scatter in a direction perpendicular to the slalom line, we define the "transverse offset" $Y$ (Figure 5.1). In the case of a constant velocity
medium, Lamer et al (1979) calculate the correction for a single reflector exhibiting pure crossdip (i.e., no inline dip). They assume that processing to the stage of NMO-corrected CMP gathers reduces all traveltimes to those which would have been measured in a true zero-offset experiment (source and receiver coincidentally located at the midpoint). Then compared to its counterpart measured at the bin centre \( t_c \) (subscript "c" is a mnemonic for "bin centre"), the zero-offset reflection time \( t \) associated with a scattered midpoint is delayed by an amount \( \Delta T \) given by

\[
\Delta T = \frac{2 \sin C}{V} Y ,
\]

where \( C \) is the angle of crossdip and \( V \) is the velocity (Figure 5.1).

Generalizing their argument to a multi-layer model (uniform layers, arbitrary crossdips, no inline dip), they show that the appropriate delay may be obtained from equation (5.1) by replacing \( V \) with the velocity of the uppermost layer, and \( C \) with the angle of emergence of the normal incidence ray (measured from the vertical). In either case, the
implication is that the traveltime perturbation due to crossdip is linear in $Y$:

$$t = t_c + pY$$  \hfill (5.2)

for some "crossdip parameter" $p$.

Equation (5.2) suggests that the crossdip correction might be performed by stacking NMO/DMO-corrected CMP gathers along linear trajectories in ($Y$-$t$) space. Such a procedure is readily implemented using the existing ITA software (Wang, 1990). The ITA module "SLNR" performs the conventional "slant stack" filtering (e.g., Yilmaz (1987)) by summing input gathers along linear paths $\tau$ in ($X$-$t$) space given by

$$\tau = t_o + aX,$$\hfill (5.3)

and placing the output result at time $t_o$. Here, $X$ denotes source-receiver offset, $t_o$ is the two-way traveltime measured at zero source-receiver offset and $a$ is some specified slope. From equation (5.2), we wish to perform the analogous summation in ($Y$-$t$) space. This may be accomplished simply by first writing the transverse offset into the ITA trace header word that conventionally contains source-receiver offset information, then running SLNR.

Because of its ease of implementation, a crossdip correction based on (5.2) is attractive from a practical viewpoint. Unfortunately, the simple geometric argument used in deriving equations (5.1) and (5.2) does not account for the presence of an inline component of reflector dip; further, it ignores any variation in source-receiver azimuth for a set of midpoints lying within a particular bin. Along many portions of SBC18, we anticipate dip components in both the inline and crossline directions. Furthermore, the crookedness of the line results in a wide range of source-receiver azimuths.

In order to assess the validity of applying the proposed simple crossdip correction, we consider a three-dimensional model consisting of a single dipping reflector embedded in a medium of constant velocity $V$. Despite the obvious oversimplification of the model,
we hope that the insight gained from the present analysis will carry over to the real world situation. In Appendix A, we derive the reflection traveltime equation for a crooked-line CMP gather. Here we present the result with reference to Figure 5.2. For compactness of notation, we define

$$\gamma \equiv \sqrt{1 + \tan^2 I + \tan^2 C},$$

and

$$\delta \equiv \tan I \cos \rho + \tan C \sin \rho,$$

where $I$ and $C$ are the angles of inline dip and crossdip, respectively, and $\rho$ is the source-receiver azimuth. Next, we define the “crossdip slowness” $p$:

$$p \equiv -\frac{2 \tan C}{V \gamma}.$$

Then the reflection traveltime $t$ associated with an arbitrary midpoint is given by

$$t^2 = (t_c + pY)^2 + \frac{B^2}{V^2} \left(1 - \left(\frac{\delta}{\gamma}\right)^2\right),$$

where $B$ is the source-receiver offset and $t_c$ is the zero-offset two-way traveltime measured at the bin centre (Appendix A).

An identical expression (i.e., equation (5.7)) was derived independently by Weizhong Wang (personal communication, 1993), who is using it to develop a DMO procedure more suitable for crooked line processing than the conventional 2-D algorithm.

Note that in the limit of zero crossdip, for a source-receiver azimuth along profile, equation (5.7) reduces to the standard dip-moveout equation (cf. equation 4.2)

$$t^2 = t_c^2 + \frac{B^2}{V^2} \left(\cos^2 I\right),$$

a result which is independent of the transverse offset. In the limit of zero inline dip (source-receiver azimuth again along profile), we have

$$t^2 = (t_c + pY)^2 + \frac{B^2}{V^2}.$$
Fig. 5.2: The model assumed in deriving equation (5.7). (a) Plan view. The slalom line is coincident with the $x$-axis and the CMP bin centre is located at the origin. (b) Three-dimensional view of the earth model. The uniform dipping layer over a halfspace is defined in terms of its perpendicular distance from the origin, $d$, the layer velocity, $v$, and the angles of inline dip, $I$, and crossdip, $C$, defined in the figure.
where \( p \) simplifies to

\[
p_{\text{cross}} = \frac{-2 \sin C}{V} .
\]  

(5.10)

Considering equation (5.9), we see that the application of standard NMO has the effect of reducing the reflection traveltime \( t \) to the "incorrect" zero-offset two-way traveltime \( t'_o \) given by

\[
t'_o = (t_e + p_{\text{cross}}Y) .
\]  

(5.11)

The residual error in zero-offset traveltime is in agreement with the result of Lamer et al (1979) (equation (5.1)).

Returning to the general 3–D situation (equation (5.7)), we note that all dependence on the source-receiver azimuth is confined to the second term in the equation, the "moveout" term. For a typical CMP bin along SBC18, we might assume that \(-15^\circ \leq \rho \leq 15^\circ\) defines a "reasonable" domain for the source-receiver azimuth. If the crossdip and inline dip angles both vary between \(0^\circ\) and \(45^\circ\), the "moveout perturbation" term \((\frac{\delta}{\gamma})^2\) ranges between 0 and 0.54 (instead of extremizing (5.7) analytically, we arrived at this result iteratively by writing a simple FORTRAN program). Therefore 3–D effects on traveltime moveout may be described in terms of an increased "effective" moveout velocity \(V_e\) (for our region of interest then, \(V \leq V_e \leq 1.5V\)). In Section 4.3 we noted that the actual P-wave velocities of the mid to lower crust were sufficiently high that moveout effects were not significant for later events (beyond approximately 5 sec). We see here that the three-dimensionality serves to further reduce the amount of traveltime moveout; hence, we do not expect the crossdip effects to seriously hinder our efforts to reduce traveltimes to their zero-offset counterparts, except possibly at early times. The significant implication is that the crossdip effect is essentially manifest as a residual error in the zero-offset traveltime. A noteworthy corollary to this observation is that the crossdip correction, in contrast to
the NMO and DMO processes, may be significant for late times. On the basis of these observations, we feel that we are justified in correcting crossdip effects, at least to first order, by application of the slant stack.

Having ascertained that a first-order crossdip correction may be obtained through use of the slant stack, we now address the second of the two queries presented in the chapter introduction, namely the issue of whether or not knowledge of the crossdip parameter $p$ may be used to provide information about the true 3-D structure. As a general statement, quantitative estimates of true (3-D) reflector dip angle would be unreliable. It is only under very restrictive circumstances that the simple 3-D model (Figure 5.2) adequately represents the real earth situation, and even then, severe crossdip can adversely affect the success of the calculation. In Appendix B we show that for a range of modest crossdip angles, the inline dip component may be directly measured from the seismic section. In such a case an estimate of true dip, which requires knowledge of both orthogonal dip components (crossdip and inline dip), may be obtained. In this appendix, we perform a sample calculation for one of the shallow reflectors associated with the Pasayten fault, 10 km to the east of the Fraser Canyon.

5.2.1 Applying DMO before crossdip correction: synthetic tests

In order to test the validity of cascading DMO and the slant stack crossdip correction (equation (5.2)), we wrote some FORTRAN code to generate 100 synthetic crooked-line CMP gathers based on equation (5.7). We chose the midpoint spacing, the number of traces per gather (i.e., the CMP fold), and the source-receiver offset increment within a single gather based on the corresponding SBC18 nominal values. To simulate midpoint scatter, we synthesized transverse offset values using a random number generator (zero mean, standard deviation = 200 m). We generated a similar set of random numbers to simulate a spread in the source-receiver azimuth (zero mean, standard deviation = 15°).
We created a reflecting interface by specifying a normal distance to the first midpoint bin centre of 5000 m. We set both dip components equal to 30° and assumed a velocity of 5000 m/s.

In our first test, we performed a crossdip correction without application of DMO by slant stacking NMO-corrected gathers (NMO correction was performed using the true medium velocity). A satisfactory result was obtained using $p = 2 \times 10^{-4}$, which corresponds to a crossdip of 30° using the simple relation given in equation (5.1). Hoping that DMO would improve the result, we performed NMO followed by DMO before slant stacking. However, the application of DMO had a deleterious effect on the stack quality; presumably this was because of a violation of the 2-D assumptions which underlie the method. On the other hand, after running tests using different earth model parameters, we found that applying DMO actually improved results when the inline dip was much greater than the crossdip.

Therefore, the question of whether or not to include DMO in the crossdip correction stream appears to depend on the severity of the crossdip. Of additional interest is the fact that the dip measured on the synthetic stacked sections after crossdip correction corresponded approximately to the inline component of dip, an observation rooted in intuition, and moreover one which is in harmony with the investigation carried out in Appendix B.

5.3 Crossdip parameter estimation

The “constant slowness stack” (CSS) provides a straightforward approach for estimating the optimum value of crossdip slowness $p$. Operating in complete analogy to the constant velocity stack (CVS) described in the previous chapter, the procedure is simply implemented using the SLNR module, and has been used in previous crustal scale crooked line processing studies (Wang, 1992; Kim et al., 1992). As in the CVS case,
the optimum parameter choice is often ambiguous. In this section, we describe a method we have developed for determining $p$ that operates on NMO-corrected CMP gathers. It is based on the work of Key and Smithson (1990), who have devised a new method for velocity analysis which is based on the eigenstructure of the data covariance matrix. Compared to the traditional method of semblance, their procedure offers enhanced resolution of events closely spaced in time and velocity. Moreover, it offers robustness in the presence of both low S/N environments and static shifts.

We begin by considering a coincident source-receiver pair at the bin centre (zero transverse offset). We assume that a crossdipping reflector gives rise to a signal centred at time $t_c$ on the associated zero transverse offset trace (we assume here the existence of a wavelet of finite duration). We postulate that the reflection signal on the $i$th trace is delayed by an amount

$$r_i = pY_i$$

(5.12)

compared to the corresponding signal measured at the bin centre. Here, the transverse offset $Y_i$ has been subscripted to emphasize that it is a trace-specific quantity.

We consider a narrow time window comprising $N$ time samples, whose duration is on the order of the Klauder wavelet, centred on the linear trajectory or "search path" $t_i$ defined by

$$t_i = t_c + \tau_i.$$  

(5.13)

By following or "tracking" the signal along the trajectory (equation (5.13)), we may decompose each data trace $r_i(t)$ into its signal and noise components, $s(t)$ and $n_i(t)$ over the time window of interest:

$$r_i(t) = s(t) + n_i(t).$$

(5.14)
We have omitted the subscript in writing the signal term to emphasize that we are assuming that the signal component is common to all traces in the gather.

The procedure entails combing over a range of values for both $t_c$ and $p$, and finding the search path that best tracks the signal according to some coherency criterion. Hence, we are engaged simultaneously in a task of crossdipping event detection (finding $t_c$) as well as crossdip slowness resolution (finding $p$). Of course over the entire length of the data trace, we anticipate the presence of many crossdipping interfaces. The crucial assumption is that at any fixed value of zero transverse offset time $t_e$, there exists only one signal associated with a unique crossdip slowness. In other words, we assume that there are no intersecting crossdips for a given value of $t_c$. This assumption may be violated in complex geological environments.

We associate each trace with a different random variable and regard the trace amplitudes as realizations of these random variables which we group into a $1 \times M$ vector $r = (r_1, \ldots, r_M)$. Here the $r_i$'s denote the random variables for each trace and $M$ is the total number of traces in the CMP gather. Proceeding in an analogous manner for the signal and noise components, we write $s = (s, \ldots, s)$ and $n = (n_1, \ldots, n_M)$ where

$$r = s + n \quad (5.15)$$

In practice, Key and Smithson (1990) suggest stacking several adjacent traces along the trajectory given by (5.13) in order to reduce run-time and to increase the S/N, thereby reducing the row dimension of $r$, $s$ and $n$ from $M$ to $M'$.

The "covariance of $r_i$ and $r_j", R_{ij},$ is defined by

$$R_{ij} = \mathbf{E}[ (r_i - \mathbf{E}[r_i])(r_j - \mathbf{E}[r_j]) ] \quad (5.16)$$

where $\mathbf{E}[\bullet]$ denotes the expectation operator. The $M' \times M'$ data covariance matrix $R$ may be formed by taking expectations of the outer product of $r$ and $r^\mathsf{T}$. If we assume that
both signal and noise are zero mean with variances $\sigma_s^2$ and $\sigma_n^2$ respectively, that there is no correlation between signal and noise, and further that the noise is uncorrelated from trace to trace, then from equation (5.15) we have

$$R = E \left[r^T r \right] = \sigma_s^2 \zeta + \sigma_n^2 I \quad ,$$

(5.17)

where $I$ is the identity matrix and $\zeta$ is the matrix whose elements are identically equal to one. The largest (major) eigenvalue of $R$, $\lambda_1$, is equal to $M'\sigma_s^2 + \sigma_n^2$ (there appears to be an error in the paper of Key and Smithson (1990) with regard to the size of this eigenvalue), and the smallest (minor) eigenvalue is $\sigma_n^2$, of multiplicity $M'-I$ (i.e., $\lambda_2 \ldots \lambda_{M'}$ are equal to $\sigma_n^2$).

Of course in our statistical formulation there is no way of knowing the values of the true covariance elements, but we may estimate them nonetheless according to

$$\overline{R}_{ij} = \frac{1}{N} \sum_{n=1}^{N} r_i^n r_j^n \quad ,$$

(5.18)

where $[\ast]$ denotes an estimate, $N$ is the number of time samples per trace within the window of analysis, and $r_k^n$ is the $n$th sample within the window of the $k$th trace.

We may estimate the noise variance by taking the average of the minor eigenvalues of $\overline{R}$, our numerical estimate of the covariance matrix:

$$\overline{\sigma_n^2} = \frac{1}{M' - 1} \sum_{i=2}^{M'} \lambda_i \quad ,$$

(5.19)

We may then estimate the signal variance by

$$M' \overline{\sigma_s^2} = \lambda_1 - \overline{\sigma_n^2} \quad .$$

(5.20)

If there is a strong signal present along the trajectory defined by fixed values of $t_c$ and $p$, then we expect a relatively large value for the signal variance. The question
remains of how we might exploit the eigenstructure of the data covariance matrix to yield an optimal signal-noise discriminant appropriate for use as a coherency measure. A measure of the signal-to-noise ratio $s/n$ is provided by forming the ratio of the signal and noise variances:

$$s/n = \frac{\sigma_s^2}{\sigma_n^2}. \quad (5.21)$$

Key and Smithson (1990) define the "covariance measure" $C_c$ to be the product of the signal-to-noise estimate $s/n$ and a weighting function $W_c$ defined by

$$W_c = N \ln \left( \frac{\sum_{i=1}^{M'} \frac{A_i}{M_i}}{\prod_{i=1}^{M'} \lambda_i} \right)^{M'} . \quad (5.22)$$

Then from equations (5.20)–(5.22), we have

$$C_c = NM' \left( \ln \left( \frac{\sum_{i=1}^{M'} \frac{A_i}{M_i}}{\prod_{i=1}^{M'} \lambda_i} \right) \cdot \left( \frac{\sum_{i=2}^{M'} \lambda_i}{\lambda_1 - \sum_{i=2}^{M'} \lambda_i} \right) \right)^{M'}. \quad (5.23)$$

We have written a program which computes $C_c$ according to (5.23) for a range of trial values of $p$ and $t_c$. Because of the large dynamic range associated with the covariance measure, we have followed the suggestion of Key and Smithson (1990), and have applied an AGC gain to the input CMP gathers. We used the subroutines “TQLI” and “TRED2” of Press et al. (1986) to compute the eigenvalues of our estimated data covariance matrix. Best results were obtained using a time window comprising 35 time samples. In order that such a window encompass only the Klauder wavelet, we resampled the input CMP gathers using a sampling interval of 0.008 s. We used an increment of 0.03 s in searching $t_c$, and a slowness search increment of $1 \times 10^{-5}$ s/m. For 60-fold data, we found that a
partial stack obtained by summing 6 neighbouring traces provided acceptably short run
times (-8 minutes per gather for $0.8 \leq t_c \leq 4.5$ s and $-1.3 \times 10^{-4} \leq p \leq 1.3 \times 10^{-4}$ s/m).

Because we were using an estimate of the covariance matrix, the smallest eigenvalues
were often very close to zero. We found it necessary to add a small positive term
(0.01 was typically used) to the denominator of the argument of the natural logarithm
in equation (5.22) to ensure stability in the computation. We used the ITA interactive
velocity analysis module “VANMO” to produce colour contour plots of $C_c$ in $p-t_c$ space.

5.3.1 Algorithm testing

We tested the “covariance algorithm” extensively on synthetic CMP gathers contain-
ing crossdipping events which we created using the algorithm described in Section 5.2.1.
We varied both the signal-to-noise ratio and the amount of inline dip.

We present the result of one such test in Figure 5.3, where we have generated two
signals, each associated with a single crossdipping interface, and superposed them on
a single CMP gather. The first corresponds to $p = -4 \times 10^{-4}$ s/m, $t_c = 1.6$ s, and the
second to $p = +4 \times 10^{-4}$ s/m, $t_c = 3.2$ s (no inline dip). We added random noise to the
CMP gather and sorted to transverse offset before proceeding with the covariance analysis.
Unfortunately, the VANMO software does not display the colour scale alongside the plot,
although we know from the documentation that it is a logarithmic one. It is evident from
the figure that the algorithm has detected the two signals and correctly resolved the
crossdip slownesses. Nevertheless, the peak associated with the later event, whose signal
strength is substantially weaker than its earlier counterpart (even after AGC), is barely
visible. Aside from illustrating the success with which the algorithm is able to detect
signal in a noisy environment, this example serves to illustrate a problem associated with
the display of the covariance measure because of its large dynamic range. Of course it is
possible to apply a gain function, such as AGC, to the display itself, although there may be
great subjectivity associated with such a process. The display problem notwithstanding, the algorithm generally yielded encouraging results in the synthetic environment.

Unfortunately, the covariance algorithm produced disappointing results when we applied it to real data. The undershoot processing sequence described in Chapter 6 entailed binning along a series of slalom lines that cut through the midpoint scatter in the vicinity of the Fraser Canyon. From the several short profiles we created in the E-W direction, we were aware of the presence of a dipping reflector at approximately 3.6 s. Results of the constant slowness stack tests that we performed on the E-W profiles suggested that there was no significant crossdip component. However, as part of the analysis, a slalom line was constructed along the direction of regional strike, essentially the N-S direction (the CMP bin map is presented in Figure 6.7), and we expect that the inline dip observed on the E-W sections at 3.6 s gives rise to a crossdip on the N-S line. The associated crossdip slowness is equal to the (readily measurable) time dip on the E-W stacked sections, namely $-1.3 \times 10^{-4}$ s/m (examine Figures 7.6b-d near 3.6 s). We performed a constant slowness stack analysis on the N-S line, and determined that slant stacking using $p = -1.0 \times 10^{-4}$ s/m provided optimum reflector continuity at 3.6 s (compare Figures 5.4a and 5.4b). Hence, there is general agreement between the optimum crossdip correction and the observed time dip on the E-W profiles.

Figures 5.5 and 5.6 show the result of applying the covariance algorithm to two CMP gathers along the N-S line, where we tested slowness values on the interval $-1.6 \times 10^{-4} \leq p \leq 1.6 \times 10^{-4}$ s/m. The gather displayed in the first of these figures corresponds to trace 101 in the stacked sections of Figure 5.4, while the gather shown in the second figure corresponds to trace 45. Note that the algorithm appears to have "correctly" resolved the crossdipping feature in the first of these figures, but not in the second. Yet from Figure 5.4 we know that the crossdip correction is significant in the vicinity of both the midpoints in question. We feel the poor results are due to an error
Fig. 5.3: Covariance analysis for a synthetic CMP gather. Two synthetic crossdipping events are embedded in the noise, one at t_c = 1.6 s, p = -0.0004 s/m (event A), the other at t_c = 3.2 s, p = +0.0004 s/m (event B). The algorithm has succeeded in correctly resolving both events in p-t_c space, although the peak corresponding to event B is barely visible on the plot. AGC scaling of the righthand panel would help better highlight this later peak.
Figure 5.4. (a) Stacked section from N-S slalom line. No crossdip correction has been applied. Arrows (B), (C) and (D) at top of figure indicate intersection points of E-W lines (corresponding to Figures 7.6 b-d, respectively). Note the defocussed energy near 3.8 s. (b) Same section after crossdip correction using a crossdip slowness of 0.0001 s/m. Reflector continuity near 3.6 s has been significantly enhanced.
in model specification. The left panels in Figures 5.5 and 5.6 show the CMP gathers after sorting to transverse offset. Note that there is no obvious linear dipping signature associated with the energy near 3.6 sec. In fact, over the entire line, very few crossdipping events displayed continuity on a length scale representative of the range of transverse offsets typically observed within a single CMP gather. Nevertheless, the crossdip correction obviously enhances the image of the reflector in a “bulk” sense, as can be seen from a comparison of Figures 5.4a and 5.4b.

We performed the covariance analysis on many CMP gathers along the main SBC18 profile and found, without exception, that the constant slowness stack gave a more reliable estimate of crossdip slowness. In spite of the promise it displayed in the synthetic environment, the covariance algorithm was deemed unsuitable for inclusion in the SBC18 processing stream.

5.4 The crossdip correction processing stream

The first step in the crossdip correction process is the calculation of the transverse offset. We wrote a program to compute the transverse offset for a series of input CMP gathers. In designing the algorithm we considered the SBC18 line to be composed of two discrete segments separated by the Fraser Canyon, a western half essentially oriented in the E-W direction, and a (nominally) eastern half principally oriented in the N-S direction (Figure 1.1). We applied the program separately to these two portions, choosing south and east as the direction of positive transverse offset in the analysis of the (nominally) western and eastern segments, respectively.

After computing the transverse offset, we performed the CSS analysis in piecewise fashion, operating on 180 NMO-corrected input CMP gathers at a time. We used a crossdip slowness window defined by \(-1.3 \times 10^{-4} \leq p \leq +1.3 \times 10^{-4}\) s/m and a slowness increment of \(\Delta p = 1.0 \times 10^{-5}\) s/m in running the ITA prestack module “SLNR”.

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Fig. 5.5. Covariance analysis applied to a CMP gather extracted from the N–S slalom line near the Fraser Canyon. The algorithm appears to have correctly identified the crossdipping event at 3.6 sec.
Fig. 5.6. Covariance analysis applied to a CMP gather extracted from the N–S slalom line near the Fraser Canyon. The algorithm does not detect any crosstipping event near 3.6 sec. Compare with Fig. 5.5.
Assuming a homogeneous medium of constant velocity 6000 m/s and no inline dip, this
slowness window maps to a range of crossdips $-23^\circ \leq C \leq 23^\circ$ (see equation (5.1)).
Each output CSS file consisted of 27 stacked panels (one panel per slowness), with
each panel containing 180 traces. Unfortunately, we were unable to pick the slownesses
interactively using the graphics monitor due to limitations in the ITA plotting software,
so we produced hardcopies of the panel tests using the 36" plotter. From these paper
sections, we determined the set of slowness-time profiles that gave rise to the optimum
stack.

5.5 Data results—stacked sections after crossdip correction

The crossdip correction significantly enhanced continuity of reflections in many areas
of SBC18. In general, the correction was not very important beyond approximately 8
s, a statement attesting to the essential flatness of the lower crustal fabric. Figure 5.7a
displays a portion of the brute stacked section near the western edge of the line. Figure
5.7b shows the same section after application of crossdip correction (otherwise same
processing sequence). For the event at 2.0 s, the improvement in both reflector strength
and continuity is quite impressive (here, a crossdip slowness of $8 \times 10^{-5}$ s/m was used
in performing the correction). As well, subtle improvements may be noted elsewhere
in the figure.
Figure 5.7. Example of crossdip correction. (a) Brute stack taken from near western edge of SBC18. No crossdip correction has been applied. (b) Same as (a) except that crossdip correction has been applied using a crossdip slowness of 0.00008 s/m. Note the improved image of the reflector at 2.0 s (compare boxed regions in (a) and (b)).
Chapter 6
PROCESSING NEAR THE FRASER CANYON

6.1 Introduction

In Chapter 1, we stated that a more detailed subsurface characterization in the vicinity of the Fraser fault figured among our primary research objectives. Results of the contract processing revealed apparent mid-crustal continuity of reflectors across the fault zone, implying that the fault does not penetrate the middle crust. In an effort to clarify questions concerning both fault geometry and depth extent, we performed additional processing of that portion of the dataset acquired near the Fraser Canyon, exploiting the large degree of midpoint scatter to yield three-dimensional subsurface information. Predictably, DMO tests yielded disappointing results and no dip moveout correction was applied. Processing efforts were complicated by the incomplete removal of static shifts in the main processing stream (Chapter 3), crossdip effects, and rapid spatial variation of subsurface structure which rendered the character of the output stack very sensitive to the choice of binning geometry.

6.2 CMP binning

In order to perform a "pseudo 3-D" subsurface characterization near the Fraser fault, we considered several different binning geometries which cut through the midpoint scatter. During the course of the preliminary processing, Western experimented with several different bin lengths along this part of the line; it was concluded that exceedingly large bins (-5 km) did not produce satisfactory results (John Varsek, personal communication, 1992). We tested bin lengths ranging from 700–2000 m before choosing the binning schemes or "swaths" presented in Figures 6.1–6.7. Unfortunately, ITA does not provide an algorithm for plotting the CMP bin maps; consequently, "screendumps" were used
in the following result presentation (midpoints, indicated in red, have been multiplied compared with the displayed station numbers (black)). Details of the plotting convention are provided in Figure 6.1. Although we have provided approximate distance scales for each of these figures, it should be noted that the aspect ratios are not precisely 1:1.

6.3 Static and crossdip effects

In Chapter 3 we noted that the weathering layer heterogeneity appeared to be particularly severe near the western edge of the Fraser Canyon. On our brute stack, reflection horizons in this area lacked the coherency apparent on the section produced by the preliminary (contract) processing, where the statics calculation was performed simultaneously on the entire line, rather than on the two separate segments corresponding to either side of the canyon. Incidentally, our initial concern that the Western statics analysis might have induced false continuity of reflectors across the fault zone (where we anticipate the possibility of abrupt geological discontinuity) was allayed when we learned that the parameters chosen for the computation caused very little structural smoothing (Peter Cary and Roger Hawthorne, personal communication, 1993).

In Section 3.2.1 we stated that we split the dataset into two segments—one on either side of the Fraser River—before proceeding with the refraction statics analysis. Therefore, we anticipate a deterioration in the quality of the statics solution near the canyon due to edge effects. We noted that the near-surface model broke down over the final kilometer leading up to the western edge of the canyon, where the REFSSTAT2 algorithm assumed a constant velocity bedrock model \( V = 5643 \) m/s. In addition, it is quite possible that the spatial sampling in this area was too coarse to permit an accurate analysis of rapidly varying near-surface structure (recall that roll-on and roll-off acquisition schemes were used near the canyon (Figure 2.2)). Reliable estimation of the high-frequency static
Fig. 6.1. Main slalom line across Fraser Canyon. Some bin locations in the figure are obscured by the source-receiver midpoints (dark blue).
Fig. 6.2. Slalom line (west side of Fraser Canyon).
Fig. 6.3. Slalom line (east side of Fraser Canyon).
Fig. 6.4. Southern slalom line (undershoot).
Fig. 6.5: Central slalom line (undershoot).
Fig. 6.6: Northern slalom line (undershoot).
Fig. 6.7: Slalom line in NW–SE direction (undershoot).
component is contingent on an adequate distribution of sources and receivers whose refractor raypaths contribute to the analysis.

The fact that the statics computation was based on a two-dimensional earth model represents an additional shortcoming. In September, 1992, shortly after we completed our refraction statics analysis, Hampson-Russell Software Services Ltd. donated their three-dimensional refraction statics software “GLI3D” to the UBC Department of Geophysics and Astronomy. Their iterative statics computation comprises two basic steps: first forward modelling by ray-tracing head waves through the “current” 3-D earth model estimate (an initial model is specified by the user), then refining the solution by considering the discrepancy between the modelled first break times and the observed picks. The latter objective is achieved through the method of Generalized Linear Inversion (see, for example, Hampson and Russell, 1984). Although the algorithm inverts for a discrete set of model parameters defined on a 2-D grid, for the purposes of ray-tracing, the near-surface model is represented by a series of continuous functions $E(x,y)$, $D_i(x,y)$ and $V_i(x,y)$, representing the surface elevation, the elevation at the base of the $i$th layer, and the velocity within the $i$th layer, respectively. After each iteration of the inversion, these functions are updated according to the following procedure: First the $x$-$y$ space is divided into a series of rectangular “patches”. Then within each patch, a bi-cubic polynomial is calculated which best fits the subset of model parameters lying inside the patch in a least-squares sense. Across the patch boundaries, these functions are constrained to be continuous with continuous first derivatives.

Because of the smoothing inherent to the bi-cubic spline, the earth model is forced to exhibit relatively low-frequency spatial variation. This gives rise to a short wavelength discrepancy between the predicted first break traveltimes computed from the final (smooth) model and the actual picks. The remaining high frequency static component
is estimated by a surface-consistent decomposition of this discrepancy (it is simply apportioned into a shot and receiver component), so that the final static solution embodies all orders of spatial variation. We specified a near surface model consisting of a single weathering layer overlying a bedrock refractor and tested the algorithm on the profile associated with the slalom line of Figure 6.2. The result is presented in Figure 6.9, while the corresponding result based on our 2-D refraction statics analysis (brute stack) is shown in Figure 6.8. Application of 3-D refraction statics appears to have slightly enhanced reflector continuity; however we do not feel that the degree of improvement was sufficient to warrant a complete 3-D analysis in the vicinity of the canyon. On the other hand, inclusion of the 3-D algorithm into the general processing stream would likely have improved results along particularly crooked sections of the profile, although time did not permit the undertaking of such a task.

6.3.1 Cascading crossdip and residual statics corrections

If indeed the refraction-based analysis failed to completely correct for the short wavelength static component, then we would hope that a subsequent application of surface-consistent residual statics would remove any remaining time shifts associated with near-surface variation. In Section 3.1, we stated that crossdip effects may detract from the reliability of the residual statics solution. Conversely, the presence of static traveltime anomalies may hinder estimation of the optimum crossdip slowness. Nevertheless, we follow the procedure proposed by Kim et al. (1993) and cascade the two processes, correcting first for crossdip, then applying residual statics.

We estimated the surface-consistent residual statics using the method of Ronen and Claerbout (1985). Rather than performing a time-term decomposition of reflection traveltimes in the manner of the Western proprietary algorithm MISER, their method entails choosing the set of shot and receiver statics that maximizes the stack power. Like
Figure 6.8. Brute stack corresponding to CMP bin map in Fig. 6.2. Brute stack processing stream is described in Section 3.5.
Figure 6.9. Stacked section corresponding to CMP bin map in Figure 6.2 after application of 3-D refraction statics (otherwise same processing stream as Figure 6.8).
the MISER algorithm, it is unable to reliably estimate the long-wavelength component of the static anomaly but is successfully used in the oil industry for the computation of high-frequency statics. Theoretically speaking, the fact that we are applying both refraction and residual statics represents an improvement to the processing flow of Kim et al. (1993), because our analysis accounts for the presence of both long and short wavelength statics.

We performed the residual statics analysis on the crossdip-corrected profile corresponding to the swath displayed in Figure 6.2. After considerable testing, we chose to maximize the stack power for all midpoints on the west side of the canyon over a very large time window (2-11 s). The resulting residual statics solution was applied to all profiles described in this chapter with the exception of the one located on the eastern side of the canyon (Fig. 6.3), the latter profile not being associated with any shots or receivers located on the western side of the canyon.

6.4 Summary of processing stream and data results

The undershoot processing flow is summarized in Figure 6.10. The creation of multiple swaths through the midpoint scatter associated with the undershoot succeeded in providing a more detailed characterization of the fault zone (the associated result interpretation is presented in Chapter 8). Cascaded application of crossdip correction and residual statics resulted in a modest improvement in the appearance of the final stacked sections compared with the brute stacks. Figure 6.11 is the final stacked section for the swath illustrated in Figure 6.2. Comparison of this figure with the corresponding brute stack (Figure 6.8) reveals, in particular, an enhancement of the continuity of the east-dipping event at approximately 3.4 s.

As a means of calibrating our work against the preliminary processing, in Figure 6.12 we present a composite stacked section which we created by merging the final stacks
Figure 6.10. Processing sequence used in the vicinity of the Fraser Canyon.

for the swaths presented in Figures 6.2 and 6.4. The resulting binning geometry closely approximates that used in the preliminary processing. Comparison of our composite stack (Figure 6.12) and the corresponding Western result (Figure 6.13) indicates that essential features of the reflectivity are present in both sections. Note in particular the strong reflectivity at 5.0 s which appears to be continuous across the fault zone (the reflective band also extends onto the section corresponding to the eastern edge of the canyon, which is not displayed in these two figures), providing a basis for the interpretation of a shallow depth extent for the Fraser fault.

Although the plot scaling has arguably biased the displays in the favour of our result (Figure 6.12), we feel that these two sections are of comparable quality from the viewpoint of reflector continuity, this in spite of our more elaborate processing sequence. Ronen and Claerbout (1985) state that their residual statics method (i.e., the statics algorithm
Figure 6.11. Same stacked section as Fig. 6.8, except that crossdip correction and residual statics have been applied. Note the improvement in reflector continuity at approximately 3.4 s.
Figure 6.12. Composite stacked section created by merging the final stacked sections for the swaths displayed in Figures 6.2 and 6.4. This binning geometry is essentially identical to the one used in the preliminary processing. Compare with Fig. 6.13.
Figure 6.13. Western stacked section near the Fraser Canyon. Compare with Fig. 6.12.
we used in reprocessing the data near the Fraser Canyon) is more robust in the presence of noise than the standard "time-term decomposition" algorithm based on the work of Wiggins et al. (1976) (i.e., the algorithm used in the contract processing); however, experience has shown that the latter method often produces "better" results on noisy data (Peter Cary, personal communication, 1993). We were unable to perform comparative testing of the two methods because the ITA software does not support a version of the time-term algorithm. Nevertheless, its inclusion in the Western processing stream has apparently yielded a satisfactory result in this region, in spite of the fact that the method does not account for reflector crossdip.
Chapter 7 FINAL PROCESSING AND RESULT PRESENTATION

In this short chapter, we describe the three final stages of processing—data merging, coherency filtering and migration—and present the output sections.

7.1 Data merging

In Chapter 4, we noted that the application of DMO degraded the stack quality in many areas along the line, while improving it in others. To create the final SBC18 stacked section, we included the DMO version of the stack only for those regions along the line where it appeared superior to its “non-DMO” counterpart (stations 60–120 and 950–1300). We merged the DMO and non-DMO stacks using the ITA poststack modules “MERGE_TRACE” and “MERGE_TIME”. We first performed the time merge, joining the upper 5.0 s of the DMO stack with the lower 11 s of the corresponding non-DMO stack. We specified an overlap window between 4.5 and 5.0 sec, in which the two datasets were blended using a linear weighted taper. Next, we performed the spatial or “trace” merge to join adjacent portions of the DMO and non-DMO stacks. Here, we used the full 16 s of input data and specified a trace overlap zone of 20 traces over which the linear weighted taper was applied. All final sections displayed in this chapter have been merged in this manner, with the exception of those associated with the special processing near the Fraser Canyon described in the previous chapter.

7.2 Coherency filtering

The low S/N associated with deep crustal profiles presents a problem from the viewpoint of result presentation. The compressed spatial scale at which the sections are displayed often exceeds the resolution capability of the human eye, rendering event detection difficult, if not impossible. Automated coherency filtering provides an objective
means of isolating coherent, linear signal from a background noise field, thereby improving the display.

We used the coherency filtering algorithm of Lane (1991). For a given input point \((x_0, t_0)\), the algorithm computes semblances over a lateral trace window for a range of apparent slownesses (i.e., time dips). A coherency measure equal to the semblance value raised to a user-specified exponent is then computed for each time dip, and the trace amplitude at the point is smoothed by averaging the amplitudes over the trace window along the direction of maximum coherency. Finally, multiplication of this smoothed result by the maximum value of the coherency measure yields the output trace amplitude at \((x_0, t_0)\). Additional noise rejection is accomplished by the introduction of a plotting bias. A constant negative value, specified as a percentage of the average absolute amplitude computed over the entire dataset, is added to all trace amplitudes. When the coherency-filtered dataset is plotted using the variable area display only (i.e., no wiggle trace), only very large positive values contribute to the plot.

Because linear events significantly shorter than the window length are attenuated, we specified a lateral window equal to the shortest length of coherent, linear energy observable on our final zero-offset stacked section (11 traces). We computed semblances for 11 time dips between ± 0.34 s/m, these bounds corresponding to the maximum time dips observable on the final zero-offset section. Based on a visual inspection of the output of several algorithm test runs, we chose a semblance exponent of 1.0 (i.e., the coherency measure was set equal to the semblance) and a plotting bias of −150%. Larger (magnitudes of) plotting biases tended to obscure weaker coherent events, and smaller values introduced undesirable levels of background noise.

### 7.3 Migration

Migration seeks to restore dipping reflection events to their true subsurface loca-
tions—by shortening and steepening them, as well as by moving them in the updip direction—and to collapse diffraction hyperbolae. In spite of its largely successful and widespread use in the petroleum exploration industry, wave-equation migration often yields "disappointingly poor results" (Warner, 1987) when applied to crustal datasets. Deep reflection horizons often appear to migrate "best", in a subjective sense, at velocities which are up to 50% lower than the interval velocities derived from an analysis of the stacking velocity profiles. Warner (1987) has postulated that the problem is related to spurious gaps in reflector continuity on the input stacked section caused by amplitude distortion and attenuation in the near surface. Under his hypothesis, the apparent truncation of a reflection horizon is not accompanied by a diffraction hyperbola, and consequently the migration algorithm must invent a synclinal reflection to cancel out the "missing" hyperbola, resulting in a "smiley" migrated section. The lateral extent of the smiles increases with increasing migration velocity and two-way traveltime, so that the problem is essentially manifest in the deeper portion of the section. In addition, several more "conventional" problems known to adversely affect the migration process can particularly beset deep crustal profiles, namely a low S/N, the incomplete removal of crossdip effects, and a poorly-determined velocity structure.

In short, our final stacked section is not ideally suited for migration. Nevertheless, we have proceeded with the application of a phase-shift migration in the hope of gaining some information concerning the approximate positioning of reflectors. We used the ITA poststack routine "PSMIG_FAST", specifying full (i.e., maximum) aperture width. The amount of aperture refers to the effective width, in traces, of the diffraction hyperbola along which summation occurs (the migration process entails summing energy along a hyperbolic trajectory and placing the output result at the hyperbola apex). After considerable testing, we chose to perform the migration on the coherency-filtered version of the zero-offset section (no plotting bias was applied). The S/N enhancement provided
by the filtering appeared to improve the result. The zero-offset section was broken into
two segments, one on either side of the Fraser Canyon, before migration. In order to
allow reflectors near the edges of the section to migrate freely, we appended one thousand
zero traces to either end of each of the two input segments.

We used a single regional velocity function which we derived by considering the
spatial average of our stacking velocity profiles. Although ITA supports a phase-shift
migration algorithm that honours lateral velocity variations, we chose not to use it because
of the limitation imposed on the number of input traces (such a limitation effectively
restricts the aperture width). Because migration algorithms are extremely sensitive to
errors in the velocity structure, several portions of the migrated result display “smiles”
characteristic of overmigration (because many of the smiles appear at early times, they
may be positively related to overmigration, and not to the problem associated with
apparent reflection truncation which we have described above).

Again we emphasize that our present objective is an approximate restoration of
subsurface reflector geometry. Chun and Jacewitz (1981) present formulae that describe
the repositioning of a dipping reflector on a migrated section for a constant velocity
medium. In order to assess the validity of applying the phase-shift migration algorithm,
we selected several reflectors that were positively identifiable on both the final stacked
(zero-offset) section and its migrated counterpart. We compared the analytical mapping
based on their formulae (assuming a “reasonable” constant velocity) to the output of
the wave-equation algorithm. In each case, the amount of reflector repositioning was
comparable, suggesting that we are justified in assuming that application of the algorithm
provides a crude sense of reflector repositioning.

7.4 Result presentation

The set of stacking velocity profiles (both RMS and interval velocities) and the station
elevations are shown in Figure 7.1 together with the upper 3.0 s of the final SBC18 zero-offset stacked section. The horizontal-to-vertical aspect ratio is 2:1 assuming a constant velocity medium of 6000 m/s. In other words, events appear "stretched" in the horizontal direction (this aspect ratio was chosen in order to reduce clutter in the display of velocity information). CMP station numbers have been multiplied by a factor of ten in this and in all subsequent figures in the chapter. Note that the "main" slalom line created in the undershoot processing (Figure 6.1) was used to join the eastern and western halves of SBC18 across the canyon in producing the final zero-offset stacked section. This section is displayed in its entirety in Figure 7.2. In order to facilitate visual inspection of this result, we have again chosen a 2:1 aspect ratio. The optimal display properties associated with the coherency-filtered version of this section (Figure 7.3) permitted the use of a 1:1 plotting ratio. Traces in this figure are plotted using the "variable area only" format (i.e., without wiggle traces). Note the considerable signal enhancement compared with the "standard" display in Figure 7.2.

In Section 7.3, we stated that the zero-offset section (Figure 7.3) was split into two segments, one on either side of the Fraser Canyon, before migration. The two migrated results are displayed side-by-side in Figure 7.4 (1:1 aspect ratio). In Section 7.3, we also noted that these sections appeared overmigrated based on the presence of "smily" artifacts at all traveltimes. For purposes of comparison, we migrated that segment lying on the west side of the canyon using a velocity profile equal to 85% of the "true" velocity profile that was used in the migration corresponding to Figure 7.4 (migration using a "reduced" velocity field is commonly performed in the oil industry). We present the result in Figure 7.5 (1:1 aspect ratio). This section reveals a somewhat less smily character at early times than its counterpart displayed in Figure 7.4. On the other hand, both sections appear to be heavily contaminated by smiles at later times, presumably because of the problem related to the spurious gaps in reflector continuity (Section 7.3).
The final stacked sections associated with the various slalom lines created near the Fraser Canyon are displayed in Figures 7.6a–g (2:1 aspect ratio). The coherency-filtered versions are plotted at the same aspect ratio in Figures 7.7a–g.
Figure 7.1. Station elevation and stacking velocity information for SBC18 together with upper 3.0 s of final zero-offset stacked section (following page). Both RMS and Dix interval (Dix, 1955) velocities are tabulated. CMP bin numbers have been multiplied by ten in this display and in all subsequent figures in this chapter. Gap in coverage near the Fraser Canyon is represented schematically by the insertion of zero traces near CMP station 6400, and again near station 6850. Top panel shows the western half of the line, including the undershoot. Note that the "main" slalom line created in the undershoot processing (bin map in Figure 6.1) was used to join the eastern and western halves of SBC18 across the canyon. Bottom panel shows the portion of the line east of the Fraser Canyon. Horizontal-to-vertical aspect ratio is 2:1 assuming a constant velocity of 6000 m/s.
Figure 7.2. SBC18 final zero-offset stacked section (following page). CMP bin fold is plotted as a function of bin number at the top of the figure. Aspect ratio is 2:1 assuming a constant velocity of 6000 m/s.
Figure 7.3. Coherency-filtered version of SBC18 final zero-offset stacked section (following page). CMP bin fold is displayed at top of figure. Aspect ratio is 1:1 (no horizontal stretching) assuming a constant velocity of 6000 m/s.
Figure 7.4. SBC18 migrated section (following page). Coherency filtering has been applied both before and after migration. The final zero-offset stacked section (Figure 7.3) was broken into two segments before migration—one on either side of the Fraser Canyon—and one thousand zero traces were appended to either end of each segment. In an effort to display the two segments side-by-side in this figure, we have omitted presentation of those portions of the dataset corresponding to the zero trace padding on both the east side of the western segment and on the west side of the eastern segment. Both of the omitted portions displayed little signal, with the exception of a strong west-dipping event located east of the canyon at 11.0 s. We have schematically indicated the presence of this reflector in the figure (boldface dashed box). No undershoot data were used in performing the migration; the resulting gap in the migrated dataset is indicated schematically in the figure by an appropriately scaled blank region in the vicinity of the canyon. Aspect ratio is 1:1 assuming a constant velocity of 6000 m/s.
Figure 7.5. Result of migration using a "reduced" velocity profile (following page). Coherency filtering was applied both before and after migration. The velocity function was obtained by scaling the velocity profile used to produce the final migrated section (Fig. 7.4) by a factor of 0.85. Migration was performed on that portion of the dataset lying to the west of the Fraser Canyon. One thousand zero traces were appended to either edge of the dataset. Compared with Fig. 7.4, this result reveals a somewhat less "smily" character at early times, although many events are likely undermigrated. Aspect ratio is 1:1 assuming a constant velocity of 6000 m/s.
Figure 7.6. Final stacked sections produced by Fraser Canyon processing stream described in Chapter 6 (following page). Each of the seven sections corresponds to a different slalom line. Inset at the bottom of each section provides a schematic map view of the associated slalom line. CMP bin numbers are referenced to bin maps displayed in Chapter 6. CMP bin fold is displayed at top of each section. Aspect ratio is 2:1 assuming a constant velocity of 6000 m/s. Section (a) corresponds to bin map in Figure 6.2, (b) to Fig. 6.4, (c) to Fig. 6.5, (d) to Fig. 6.1 (i.e., the "main" canyon traverse which is also displayed in the final zero-offset stacked section of Figs. 7.1-7.3), (e) to Fig. 6.6, (f) to Fig. 6.7, and (g) to Fig. 6.3.
Figure 7.7. Final stacked sections near the Fraser Canyon after coherency filtering (following page). Details of the correspondence between sections and bin maps are provided in caption of Fig. 7.6. CMP bin fold is displayed at the top of each section. Aspect ratio is 2:1 assuming a constant velocity of 6000 m/s.
Chapter 8  INTERPRETATION

8.1 Introduction

Varsek et al. (1993) have provided a geological interpretation of the results of the SBC18 preliminary processing (Figure 8.1). Key features of this interpretation are addressed in the body of this chapter. They include: (i) possible mid-crustal termination of the Fraser fault based on the apparent continuity of reflectors across the fault zone (trajectory “FFA” on figure); (ii) mid-crustal flattening of the Pasayten fault (symbol “PF”); (iii) correlation of the east-dipping homocline located at the western edge of the profile (“Insular ramp”) with exposed east-dipping thrust faults of the Coast belt; and (iv) the existence of a west-side-down Moho step east of the canyon (“east flank crustal root”). An alternate interpretation of the western portion of the contract-processed result is given by Monger and Journeay (1992) (Figure 8.2a). In this chapter, we present a revised interpretation which we have based on the results of our reprocessing.

8.2 Surface geology

In Figure 8.3, we have reproduced the local geology map of Figure 1.4. From west to east the profile crosses: (i) the Late Cretaceous Scuzzy Pluton, a post-accretionary magmatic arc; (ii) the polymetamorphic and plutonic rocks of the Settler Schist, bounded on the east by the steeply northeast-dipping Kwoiek fault which delineates the boundary between the Central and Eastern Coast belts; (iii) Middle and Early Jurassic fine-grained clastics of the Ladner group (of Methow terrane affinity); (iv) Permian to Jurassic oceanic rocks of the Bridge River terrane; (v) the Fraser fault; (vi) a thin slice of wrench-faulted Methow terrane consisting of Early and Middle Cretaceous fine to coarse clastics of the Jackass Mountain Group and Early and Middle Jurassic volcanic-rich fine clastics of the
Figure 8.1. SBC18 initial interpretation superimposed atop migrated and coherency-filtered section produced by the preliminary processing (adapted from Varsek et al., 1993). Interpretation is based partially on relationship between SBC18 and other Lithoprobe reflection lines in the southeastern Cordillera. Key features of this interpretation are addressed in the text. Aspect ratio is 1:1 assuming a constant velocity of 6000 m/s.
Figure 8.2. (a) Geological cross-section and interpretation of portion of SBC18 profile west of Fraser Canyon (from Monger and Journeay, 1992). Migrated and coherency-filtered stacked section was produced by the preliminary processing. Geological cross-section is based on surface mapping east of the profile. (b) Revised cartoon drawn by Murray Journeay (personal communication, 1993), identifying reflectors CB1 and CB2 (Figs. 8.4 and 8.5) with deep roots of the Coast Belt Thrust System ("CBTS" in figure).
Dewndey Creek Formation; (vi) the Pasayten fault; and (vii) Triassic and/or Jurassic granodiorite of the Mount Lytton Complex (of Quesnel terrane affinity).

Varsek et al. (1993) have indicated that the line crosses the intersection of the northwest-trending Yalakom fault and the north-trending Fraser fault (Figure 8.1, symbols “YF” and “FF”). However, Coleman and Parrish (1991) claim that the present-day southern extent of the Yalakom fault lies approximately 65 km to the north of our line (beyond the northern edge of the map in Figure 8.3), where it is truncated by the Fraser fault; thus, we infer that SBC18 crosses only the Fraser fault at the canyon.

8.3 Result interpretation

The shortcomings associated with crustal-scale migration (Chapter 7) preclude a reliable interpretation based solely on the migrated results, in spite of the fact that these sections have restored, albeit crudely, the “correct” reflector geometry. Thus, we performed our interpretation, of which a schematic representation is displayed in Figure 8.4, by concurrent analysis of both the zero-offset and migrated datasets, positively identifying coherent events on the former type of section, and gaining a sense of the associated subsurface geometrical relationships from the latter. For ease of reference, we have reproduced the coherency-filtered zero-offset stacked section (Figure 7.3) in Figure 8.5. Events are indicated on both the migrated and unmigrated sections (Figures 8.4 and 8.5). The following discussion of our interpretation proceeds, in a general sense, from west to east.

The upper crust near the west end of SBC18 is characterized by fairly strong west-dipping reflectivity between 1.0–4.0 s (Figures 8.4 and 8.5, CMP stations 27–300—note that CMP stations are multiplied by ten in these figures). This observation implies that the overlying (presumably homogeneous) Scuzzy Pluton is a shallow feature (< 3 km). Such a suggestion is corroborated by inspection of the regional Bouguer gravity map
Figure 8.3. SBC18 profile superimposed atop local geology map of Monger (1989), reproduced from Fig. 1.4 (following page). Action of migration of events A and B defined in Fig. 8.4 is indicated on the figure (symbols A and B, respectively). Study area of Grieg (1992) and the immediate area surrounding the SBC18 traverse of the Pasayten fault are also displayed (boxes 1 and 2, respectively).
Figure 8.4. SBC18 revised interpretation based on the results of our reprocessing (following page). Interpretation is superimposed atop the coherency-filtered, migrated section displayed in Fig. 7.4. Geological units, fault surface traces, and terrane linkages are indicated on the figure. Solid hand-drawn lines indicate aspects of our interpretation which are strongly supported by the data. Weaker inferences are indicated by dashed hand-drawn lines. Events are identified on both the migrated section appearing in this figure and the zero-offset section displayed in Fig. 8.5. Aspect ratio is 1:1 assuming a constant velocity of 6000 m/s. CMP station numbers are multiplied by ten in this display.
Figure 8.5. Final zero-offset stacked section after coherency filtering, reproduced from Figure 7.3 (following page). Geological units, fault surface traces, and terrane linkages are indicated on the figure (legend is provided in Fig. 8.4). Events are identified on both the zero-offset section appearing in this figure and the migrated result shown in Fig. 8.4. Aspect ratio is 1:1 assuming a constant velocity of 6000 m/s.
(Riddihough and Seeman, 1982), which does not indicate a substantial anomaly in the vicinity of the pluton. On the other hand, these maps reveal a pronounced gravity low over the Guichon Batholith (≈30 mgal) 25 km to the west, a pluton which Ager et al. (1973) estimate to be of considerable depth extent (> 12 km) based on the results of a detailed gravity survey which they conducted across its surface projection.

At later times, the zero-offset section reveals a thick sequence of east-dipping reflectors between 6.5–10.5 s (Figure 8.5, events CB1 and CB2). A similar pattern of reflectivity is observed on SBC13, corresponding roughly to the same position along strike (Varsek et al., 1993, see Figure 1.1 for line location). These events migrate well to the west of the western edge of the profile (Figure 8.4, events CB1 and CB2). Varsek et al. (1993) suggested that these ramping homoclines represent deep levels of the Coast Belt Thrust System (CBTS). In a general sense we concur with this interpretation; however, we disagree with their speculation that this ramp contains Insular superterrane material. Refraction studies by O'Leary (1992) and Zelt et al., (1992) suggest that few pockets of Insular crust have penetrated eastward beyond the Harrison Lake fault, which lies approximately 50 km west of the western edge of the profile. In collaboration with M. Journeay (personal communication, 1993), we have modified the cartoon of the CBTS (Figure 8.2a) based on the reprocessed sections. A revised drawing is shown in Figure 8.2b.

A prominent subhorizontal band of reflectors at 6.0–7.0 s may be traced from the western edge of the line eastward to station 400 on the zero-offset section (Figure 8.5, event B1). The lack of signal between stations 450 and 550 on this section is attributable to a low S/N and does not necessarily reflect a lack of structure at depth; consequently, it is possible that these reflectors extend into this noisy region. A similar pattern of reflectivity is arguably present on the eastern half of the section (Figure 8.5, event
B2). Based on our current knowledge of the regional geology and structural setting, it is difficult to place these features within the local tectonic framework, although we cannot rule out the possibility of a structural origin for this event(s). Possibly, the banded reflectivity may be related to zones of layered porosity located between the 450 °C–730 °C isotherms, which in our study area occur at approximate depths of 13 km and 20 km, respectively (corresponding to respective two-way traveltimes of 4.0 s and 7.0 s) (Lewis et al., 1992). It is interesting to note that rather than being confined to a zone between the two isotherms, SBC18 reflectivity is prominent throughout the section. Several Lithoprobe reflection profiles located in the southeastern Cordillera reveal strong reflectivity only within the band bounded by the two isotherms (Lewis et al., 1992).

Beneath stations 250–370, a sequence of west-dipping reflectors is visible on the zero-offset section at very late times (Figure 8.5, event A at 12.0–14.0 s). Migration moves this package approximately 30 km to the east-northeast, to the opposite side of the Fraser fault beneath the eastern portion of the Mount Lytton granodiorite, just above the (inferred) trace of the Moho. The considerable lateral displacement of the event(s) under the action of migration presents difficulties in presentation, although in Figure 8.4 we have schematically sketched it below station 1000 at -11 s (event A) (see Figure 7.4 caption for details). Because in reality these deep crustal reflectors are located well to the southeast of the line, we show the restored reflector geometry in map view on the regional geology map (Figure 8.3, event A). East of the Fraser fault, another west-dipping fabric is visible at 12.5–11.5 s beneath stations 1100–1240 (Figure 8.5, event B). This package migrates 15 km to the northeast, to a position beneath the Spences Bridge volcanic group, where it also appears as a deep crustal feature flattening into the Moho (event B, Figures 8.3 and 8.4). A similar pattern of reflectivity is observed on SBC12, SBC13, and SBC17 (profile locations shown in Figure 1.1).
Samarium–neodymium (Sm-Nd) (Lambert and Chamberlain, 1990) and initial strontium ratio ($^{87}\text{Sr}/^{86}\text{Sr}$) (Armstrong and Ghosh, 1990) isotopic studies indicate that the western edge of the North American craton is presently located 75 km west of the Okanagan Valley, a finding which led Varsek et al. (1993) to postulate that west-dipping fabric present at lower levels of SBC18 represents faults which accommodated synorogenic tectonic overlap of the Intermontane superterrane onto the craton. We acknowledge the possibility that the west-dipping ramps A and B represent deep levels of this tectonic overlap, although in contrast to the Varsek et al. (1993) interpretation (Figure 8.1, notation "RMBD"), we do not feel that the character of the lower crust directly beneath SBC18 justifies the westward extension of the overlap zone beyond the Fraser fault.

Approximately 70 km to the northwest of the survey, reconnaissance surface mapping of the Kwoiek fault by Journeay (1990) revealed steeply northeast-dipping ($50-80^\circ$) fault strands. Unfortunately, this feature intersects SBC18 along a noisy portion of the line (Figure 8.5, station 450), which probably accounts for the fact that we were unable to image related structure at depth. A (nominally) east-dipping reflector located at 3.5 s beneath stations 530–630 (Figures 8.4 and 8.5, event C) apparently truncates a (nominally) west-dipping event and, very speculatively, might represent deep levels of the fault.

Varsek et al. (1993) provided three possible interpretations for the subsurface trajectory of the Fraser fault (Figure 8.1). In particular, Trajectory FFA is based on (i) apparent mid-crustal continuity of reflectors across the fault zone (undershoot region) at 5.0 s and (ii) the premise that such reflections predate motion along the fault. This interpretation implies that the major geological discontinuity has a relatively shallow depth extent. However, quantitative modelling and inversion of magnetotelluric data acquired along a profile nearly coincident with SBC18 suggest that the fault penetrates the entire crust and has a sub-vertical geometry (Jones et al., 1991).
Our detailed analysis of the undershoot region (Figure 7.7) reveals that the strong sub-horizontal reflectivity at 5.0 s is essentially confined to the southern portion of the midpoint scatter, which, significantly, is approximately coincident with the location of the CMP bin swath that was used in the preliminary processing. Prominent, continuous reflectors at 5.0 s are apparent in the section corresponding to our southern swath (Figure 7.7b). However, they appear to die out or “diffuse” to the north, as evidenced by inspection of the sections corresponding to the central, main and northern swaths (Figures 7.7c-e, respectively) as well as the northwest-southeast swath (Figure 7.7f). Whereas both the “main” (Figure 8.5, event D—bin map in Figure 6.1) and “western approach” (Figure 7.7a) profiles reveal strong reflectivity at 5.0 s near the western edge of the canyon, implying a relatively large spatial extent for the associated subhorizontal fabric, the undershoot region is apparently characterized by a very localized pocket of enhanced reflectivity. This suggests that there is no single, well-defined package of reflectors extending across the fault zone, and that apparent mid-crustal continuity across the fault derives from the fortuitous alignment of structurally unrelated reflectors.

Moreover, Figures 8.5, 7.7c, 7.7d and 7.7e reveal the presence of stronger reflectivity on the east side of the canyon. This juxtaposition of different structural styles appears to extend to great depth (beyond 10.0 s), and implies that (i) the fault roots in, or possibly penetrates, the deep crust and (ii) it has a sub-vertical geometry. Such an interpretation is compatible with the results of the magnetotelluric study of Jones et al. (1991). Finally, it is worth noting that our reprocessed results, which suggest that the Moho is flat across the entire profile (an observation we discuss later in this section), do not favour Trajectory FFC (Figure 8.1) which is based on west-side-down Moho displacement.

East of the Fraser canyon, the line cuts obliquely across the surface trace of the polyphase Pasayten fault (Figure 8.6c), which is properly viewed as a fault system rather
than a single feature. Based on mapping and geochronometric studies conducted in the nearby Eagle Plutonic Complex (Figure 8.3, box 1), Grieg (1992) has identified at least two discrete episodes of transpressional faulting along this complex structure. A mid-Eocene stage of brittle east-side-up thrusting apparently succeeded a mid-Cretaceous period of ductile sinistral, east-side-up reverse displacement. The presence of northeast-dipping foliation planes near the surface expression of the fault (Figure 8.3, box 1) led him to suggest a similar dip orientation for the fault itself. Several foliation planes of comparable orientation are exposed in the vicinity of the SBC18 traverse of the fault (Figure 8.3, box 2) which, on the basis of his argument, imply that the fault dips to the northeast in our study area.

Our reprocessing has significantly enhanced the subsurface image near the Pasayten fault. However, the obliqueness of the profile to regional strike has complicated the interpretation. Dipping events P1 and P2 (Figures 8.4 and 8.5, expanded view in Figures 8.6a and 8.6b), which were not visible on the section produced by the contract processing, are of particular interest. They appear to cut off a southwest dipping feature at 2.0 s (Figures 8.6a and b, event R) whose strong reflectivity may be related to intercalation of fine-grained clastics and volcanic rocks of the Methow basin. Event P2 was not visible prior to the application of a significant crossdip correction. By considering the size of this correction and by assuming that the zero-offset section has effectively imaged the inline component of the dip, we estimated a true dip $\phi$ of 41° and a dip azimuth of [E30°N] (true dip and dip azimuth are defined in Figure A.1). We describe the calculation in Appendix B. Uncertainties have not been formally calculated, although we feel that the assignment of a 20% fractional uncertainty to our estimates of both dip and dip azimuth would adequately account for measurement errors and error in model specification (the calculation is based on a simple constant velocity model). We estimated the true dip for event P1 to be 45°, [N40°E].
Figure 8.6. Expanded view of sections near Pasayten fault. Nominal surface trace of fault (as inferred from local geology map) is indicated by "PF" on all three figures. (a) Zero-offset section. Northeast dipping reflectors P1 and P2 are indicated by blue arrows. (b) Migrated section. We have extrapolated P1 and P2 (blue dashes) to the (local) surface. (c) CMP bin map. Slalom line crosses the surface projections of P1 and P2 at locations shown in the figure. Because of severe line crookedness, midpoint stations (red) do not increment uniformly.
Both dips are comparable with the orientation of nearby foliation planes, suggesting that events P1 and P2 represent imbricate thrust faults associated with a northeast-dipping Pasayten fault system. Further to the south in Washington state, Potter and Prussen (1987) interpreted the fault to be east-dipping, based on the presence of east-dipping reflectors on COCORP profile W7 (see Figure 1.1 for line location). Based on the observation that reflectors P1 and P2 are listric into the upper crust at 2.0 s (Figures 8.6a and 8.6b), we infer a shallower depth extent for the fault system than that which was originally proposed by Varsek et al. (1993) (cf. Figures 8.1 and 8.4).

Restoration of 130 km of dextral motion along the Fraser fault places the eastern half of SBC18 opposite the northwest trending Yalakom fault (Figure 8.4, symbol “YF”), which itself has accommodated more than 100 km of right lateral offset (Kleinspehn, 1985 in Coleman and Parrish, 1992). Approximately 80 km to the northeast of our line, Coleman and Parrish (1992) identified this feature as a northeast-dipping dextral reverse-slip fault; thirty kilometres northwest of their study area, the fault is nearly vertical (Schiarizza et al., 1990, in Coleman and Parrish, 1992). Varsek et al. (1993) recognized the possibility of the existence of deep levels of this fault beneath the eastern portion of SBC18. It is our opinion that such structure is not visible on either version of the SBC18 section (i.e., our reprocessed section and/or the contract-processed result on which Varsek et al. (1993) based their interpretation). As another point of variance, a reconstruction based on 130 km of Fraser fault displacement places the northeast–dipping Mission Ridge fault well to the south of the eastern half of SBC18, thereby invalidating the proposal of Varsek et al. (1993), which was based on restoration using an older estimate of dextral offset (85 km), that the line crosses the deep root of this extensional feature (Figure 8.1, “MRF”).

By definition, the reflection Moho is the deepest laterally extensive, coherent package
of high-amplitude reflectivity appearing on a reflection section (Klemperer et al., 1986). The precise nature of its relationship to the crust-mantle boundary as well as to the refraction Moho, which is defined in terms of a P-wave velocity increase (Steinhart, 1967), is not fully understood (Jarchow and Thompson, 1989). On SBC18 we identify the reflection Moho as a broad band of reflectivity with a traveltime thickness of -0.4 s (-2.5 km), centred at approximately 11.2 s (Figure 8.4, event M). It is essentially flat across the entire line. At the extreme eastern end of the profile, a sequence of strong reflectors centred at 11.0 s (Figure 8.4, event MM) may provide evidence for a Moho time step of -0.2 s (-0.6 km, west-side-down ramp), although such a small jump is likely insignificant given the tenuous nature of reflection Moho identification.

On the other hand, Varsek et al. (1993) indicate a significant west-side-down Moho step (approx. 0.7 s) at this location (Figure 8.1, notation “MOHO”).

Their interpretation is apparently based on the observation that the zone of strong reflectivity corresponding to event MM (Figure 8.4) is time-shifted by roughly -0.3 s (i.e., it appears earlier) on the Western final stacked section. Interestingly, near the eastern end of the line, all events positively identifiable on both sections (i.e., our reprocessed result and the contract-processed section) demonstrate the same time shift independent of arrival time, whereas elsewhere along the profile, there is no such discrepancy. This points necessarily to a variance between the two statics computations (in both cases, a datum plane of 1000 m was used for the elevation correction). We are unable to provide a definitive answer as to which result is “correct”, although it is worth recalling the fact that the residual statics algorithm employed in the preliminary processing is unable to resolve long wavelength static components, whereas a refraction-based treatment can theoretically remove all spatial orders of static variation.

Zelt et al. (1993) performed traveltime inversion and amplitude forward modelling of
refraction data acquired along Southern Cordillera Refraction Experiment Line 3, 100 km to the south of SBC18 (Figure 1.1). Their velocity model reveals a very slight shallowing of the (refraction) Moho proceeding eastward from the Coast Belt into the Intermontane belt (from 37 to 34 km depth). However, because these depth estimates are associated with an uncertainty of ±1.5–2.0 km, they do not necessarily imply the existence of a west-side-up Moho ramp (Barry Zelt, personal communication, 1993), and as such, are unable to resolve the discrepancy observed between the two reflection sections.

Nevertheless, the results of the preliminary processing of reflection lines SBC13 and SBC17 indicate the presence of significant east-side-down Moho displacement beneath the western flank of the ECB (Varsek et al., 1993). This may provide an argument for the existence of west-side-down Moho displacement near the eastern flank of the ECB, since the combination of these two displacements amounts to the existence of a crustal root beneath the mountains of the ECB, which are higher than those of the flanking regions. Varsek et al. (1993) identify the west-side-down Moho ramp on SBC18 with the eastern flank of such a crustal root (Figure 8.1), and make a similar interpretation on SBC12 although in the case of the latter profile, we do not feel that the data necessarily suggest a Moho step.

Moreover, our interpretation of an essentially flat Moho across the entire profile is possibly at variance with the proposal of Varsek et al. (1993) that the crust thins eastward of the Coast-Intermontane belt boundary. They postulated that the shallowing of the Moho below SBC18 (i.e., the east flank of the crustal root) is spatially correlated with the mid-crustal flattening of two normal faults of opposite dip, the east-dipping Mission Ridge fault (Figure 8.1, “MRF”) and the west-dipping Coldwater fault whose surface trace lies approximately 90 km to the east of the profile, which essentially flank the intervening region of (presumed) Eocene crustal extension. Under this scenario, these two
extensional faults accommodated down-dropping of the upper crust, and consequently, crustal thinning.

The fact that variance between the two Moho interpretations is attributable to a discrepancy in the statics computations underscores the tenuous nature of deep crustal reflection interpretation. Although our model does not explicitly support the proposal of Varsek et al. (1993) of a crustal root beneath the Eastern Coast Belt, nor does it reinforce their argument for crustal thinning to the east of SBC18, our results do not preclude either possibility. Reprocessing of reflection lines SBC13, SBC14, and SBC17, all located on the west side of the Fraser fault (see Figure 1.1 for locations), is currently in progress at the UBC Department of Geophysics and Astronomy; additionally, reprocessing of lines SBC11 and SBC12 lying to the east of the fault has been proposed. These efforts will likely assist in providing definitive answers with regard to Moho structure, as well as further elucidation of regional tectonic processes.
Chapter 9 SUMMARY AND FINAL DISCUSSION

9.1 Crooked line processing

The quality of the preliminary processing is excellent, and as a general statement the application of advanced processing techniques yielded a modest improvement over the contract-processed result.

Application of predictive deconvolution apparently succeeded in removing short-period multiple reflections. Although strictly speaking, this type of deconvolution is founded on the assumption of a minimum phase wavelet, we showed that for a prediction gap sufficiently large so as to encompass the effective width of the Klauder wavelet, the method is relatively insensitive to the zero-phase properties of the Vibroseis source signature.

Statics were successfully treated through employment of a refraction-based procedure which honoured both long and short wavelength static variations. Success was due largely to the excellent algorithm available in REFSTAT2, but the use of the surface stacking charts in diagnosing and correcting systematic error also contributed significantly towards honing the final statics solution. Had the REFSTAT2 algorithm been used as a “black box”, with the systematic errors going unchecked, it is likely that the results would have suffered some appreciable degradation. By far the most significant correction from the viewpoint of the number of affected first break picks was that for “side lobe” error. Still, the number of picks that suffered from this phase misidentification accounted for less than 1% of the total. The inclusion of a 3–D statics algorithm (e.g., the Hampson and Russell algorithm) in the processing flow would likely improve results, especially along very crooked portions of the line. Thus we envision an improved statics stream consisting of (i) the application of a preliminary run of REFSTAT2 in order to extract delay time
information; (ii) construction of surface stacking charts (SSCs) for first break systematic error identification and correction; and (iii) a final pass with the 3-D refraction statics algorithm.

Dip moveout correction (DMO) has succeeded in enhancing the subsurface image near the Pasayten fault. On the other hand, application of DMO actually degraded the quality of the output stack along several other portions of the line, probably because the 2-D assumptions underlying the algorithm were violated.

We derived a 3-D common midpoint (CMP) bin reflection traveltime expression which explicitly separates the inline dip and crossdip effects. Examination of this equation revealed that the crossdip effect may be considered as a perturbation in the zero-offset traveltime, and consequently, that a first order correction may be implemented by use of slant stack techniques. The method we developed for crossdip slowness estimation showed promise in the synthetic environment but performed poorly when we applied it to our real data. We attribute the disappointing results to an error in model specification: if the geology were less complex (i.e., the interfaces were adequately approximated by our simple 3-D model), this might represent a viable means of efficient crossdip parameter estimation.

The creation of several CMP bin swaths through the midpoint scatter in the vicinity of the undershoot enabled a "pseudo 3-D" characterization of the subsurface near the Fraser fault. In this region where the statics were deemed to be complex, cascaded application of crossdip correction and residual statics improved the image relative to our brute stack, although the time-term residual statics algorithm used in the contract processing also appeared to yield an acceptable result.

Migration yielded disappointing results. Degradation of migrated section quality in the shallower portion of the line (above - 5 s) was probably owing to the low S/N, the
crookedness of the profile and the fact that our migration algorithm did not honour lateral velocity variations. Contamination of the deeper sections by smiles, a phenomenon which is expected in the deep crustal environment, was observed. In spite of such shortcomings, migration represents a necessary component of the processing stream; thus our approach was to first positively identify events on the zero-offset section, then to examine the migrated section in order to obtain a crude sense of true subsurface reflector geometries.

9.2 Interpretation

Particularly significant aspects of our revised interpretation include:

i) inference of a deep crustal extent (or possible crustal penetration) for the Fraser fault;

ii) correlation of two northeast-dipping reflectors, not visible on the sections produced by preliminary processing, with southwest-directed thrusting along the Pasayten fault;

iii) correlation of two east-dipping homoclines near the western edge of the profile with roots of the Coast Belt Thrust System (CBTS), leading to a revision of (deeper levels of) the CBTS structural section;

iv) inference of an essentially flat Moho (crust-mantle boundary) across the entire line.

It is important to note that our interpretation of a deep crustal extent (crustal penetration (?)) for the Fraser fault represents a suggestion based on a lack of evidence for mid-crustal continuity of reflectors across the fault zone and on the juxtaposition of different structural styles at depth on either side of the Fraser Canyon, rather than a definitive statement. Additional reprocessing preserving relative amplitude variations will likely be performed so as to better define the precise nature of the discrepancy between the two crustal fabrics. Our work and that of Varsek et al. (1993) are pioneering efforts in the sense that they represent the first-ever attempts at characterizing any of the several major right-lateral strike-slip fault systems of the Cordillera using seismic reflection techniques. As part of the proposed Lithoprobe Slave-Northern Cordillera Lithospheric Evolution
(SNORCLE) transect, two reflection profiles across the Tintina fault in the south-central Yukon are planned. These lines will cross the fault at high angle on continuous roads, and thus will not suffer from the same complications in acquisition (i.e., undershoot) and line crookedness which plague SBC18. The anticipated higher data quality will hopefully provide additional insight into the character of these impressive features.
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Appendix A: Derivation of the 3-D CMP bin reflection traveltime expression (equation (5.7))

We begin the analysis by considering the general reflection traveltime expression of Levin (1971) for a 3-D dipping planar interface embedded in a constant velocity halfspace. With reference to Figure A.1, he showed that the traveltime \( t \) for a source at the origin and a receiver at position \((x,0,0)\) is given by

\[
t^2 = \left(\frac{2D}{V}\right)^2 + \frac{x^2 (1 - \sin^2 \phi \cos^2 \theta)}{V^2},
\]

where \( D \) is the perpendicular distance between the source-receiver midpoint and the interface, \( V \) is the (constant) medium velocity, \( \phi \) is the dip angle, and \( \theta \) is the dip line azimuth, measured clockwise from the \( x \)-axis to the dip line.

This expression may be massaged to obtain the general traveltime expression for an arbitrary source and receiver location in terms of the perpendicular distance \( d \) between the origin and the interface, and the associated direction cosines \( \alpha, \beta, \) and \( \phi \). These three angles are measured between the line \( d \) and the \( x, y \) and \( z \) coordinate axes, respectively (Figure A.1a). Angles \( \alpha \) and \( \beta \) are defined on the open interval \([0^\circ, 180^\circ]\), and \( \phi \) is defined on the open interval \([0^\circ, 90^\circ]\). We consider a source at some arbitrary position on the surface, say \((x,y,z)=(x,0,0)\) (Figure A.2). We express the (arbitrary) receiver location in terms of the absolute value of the source-receiver offset, \( B \), and the source-receiver azimuth, \( \rho \). This azimuth is measured clockwise from a line parallel to the \( x \)-axis \((0 \leq \rho \leq 360^\circ)\). It is convenient to introduce a new set of coordinate axes \((x',y',z')\) as shown in the figure. In terms of the primed coordinate system, the traveltime for the new source-receiver pair \( t' \) may be written as

\[
t'^2 = \left(\frac{2D'}{V}\right)^2 + \frac{B^2 (1 - \sin^2 \phi' \cos^2 \theta')}{V^2},
\]

where all primed quantities are defined analogously to their unprimed counterparts.

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Figure A.1. (a) three-dimensional model consisting of a single dipping layer over a halfspace; (b) definition of the dip line azimuth, $\theta$. 
Figure A.2. A new source-receiver pair has been introduced to the model shown in Fig. A.1. The source is located at \((X, \Psi, 0)\); the receiver position is defined by the source-receiver azimuth, \(\rho\), and the absolute value of the source-receiver offset, \(B\). In terms of the rotated coordinate system shown in the figure, the receiver is located at the point \((x', y', z') = (B/2, 0, 0)\).
The first step in the analysis involves expressing the primed quantities $D', \phi'$, and $\theta'$ in terms of unprimed parameters. Because lines $d$ and $d'$ are parallel, as are the vertical axes of the two coordinate systems under consideration, it is easy to see from Figure A.2 that

$$\phi' = \phi,$$  \hfill (A.3)

and moreover that

$$\theta' = \theta - \rho.$$  \hfill (A.4)

The calculation for $D'$ is somewhat more involved. $D'$ is related to $d'$, the perpendicular distance measured from the primed origin down to the interface, by

$$d' = D' + \frac{B}{2} \cos \alpha',$$ \hfill (A.5)

where $\alpha'$ is the direction cosine measured between the $x'$ axis and the line $d'$. Because $d'$ runs between the primed origin, $(x,y,z) = (x,0,0)$, and a point on the interface, say $(x,y,z) = (x_0, y_0, z_0)$ we may write

$$d' = \sqrt{(X - x_0)^2 + (\Psi - y_0)^2 + z_0^2},$$ \hfill (A.6)

and

$$x_0 \cos \alpha + y_0 \cos \beta + z_0 \cos \phi = d,$$ \hfill (A.7)

the latter expression arising from the fact that the equation for the reflecting plane is

$$x \cos \alpha + y \cos \beta + z \cos \phi = d.$$ \hfill (A.8)

Because the line $d'$ is perpendicular to the plane, and therefore parallel to the plane normal $n = (\cos \alpha, \cos \beta, \cos \phi)$, we have

$$X - x_0 = \lambda \cos \alpha,$$  \hfill (A.9)

$$\Psi - y_0 = \lambda \cos \beta,$$  

$$-z_0 = \lambda \cos \phi,$$
where $\lambda$ is some arbitrary scalar. Recalling that

$$\cos^2 \alpha + \cos^2 \beta + \cos^2 \phi = 1 \quad (A.10)$$

manipulation of (A.7) and (A.9) gives

$$\lambda = X \cos \alpha + \Psi \cos \beta - d \quad (A.11)$$

Substituting (A.9) into (A.6) yields

$$d' = |\lambda| \quad (A.12)$$

We arrive at the expression for $d'$ in terms of unprimed parameters by considering (A.11) and (A.12):

$$d' = |d - X \cos \alpha - \Psi \cos \beta| \quad (A.13)$$

To resolve the ambiguity of sign in (A.13), note that the surface projection of the reflector is a line satisfying the relation

$$x \cos \alpha + y \cos \beta = d \quad (A.14)$$

Hence, in order that the new source and receiver not lie "beyond" the surface trace, we require that

$$X \cos \alpha + \Psi \cos \beta \leq d \quad (A.15)$$

and consequently that

$$d' = d - X \cos \alpha - \Psi \cos \beta \quad (A.16)$$

The next step in the analysis is to express $\cos \alpha'$ in terms of the unprimed quantities. Note that the direction cosine angle $\alpha'$ depends only on the rotation of the primed
coordinate system relative to the original system, and not on its translation. Hence, we may use the standard transformation equations for a rotated coordinate system:

\[
x = x' \cos \rho - y' \sin \rho , \\
y = x' \sin \rho + y' \cos \rho , \\
z = z' .
\]  

(A.17)

From Figure A.3, it is apparent that

\[
\cos \alpha' = \frac{x' P}{L} = \frac{x P \cos \rho + y P \sin \rho}{L} ,
\]  

(A.18)

and since

\[
\cos \alpha = \frac{x P}{L} , \\
\cos \beta = \frac{y P}{L} ,
\]  

(A.19)

that

\[
\cos \alpha' = \cos \alpha \cos \rho + \cos \beta \sin \rho .
\]  

(A.20)

We obtain the desired expression for \(D'\) from (A.5), (A.16) and (A.20):

\[
D' = d - X \cos \alpha - \Psi \cos \beta - \\
\frac{B}{2} \left( \cos \alpha \cos \rho + \cos \beta \sin \rho \right) .
\]  

(A.21)

Finally, we may rewrite (A.2) in terms of unprimed parameters, using (A.3), (A.4) and (A.21):

\[
\eta^2 = \left\{ \frac{2}{V} \left( d - X \cos \alpha - \Psi \cos \beta - \frac{B}{2} \left( \cos \alpha \cos \rho + \cos \beta \sin \rho \right) \right) \right\}^2 + \\
\frac{B^2}{V^2} \left( 1 - \sin^2 \phi \cos^2 (\theta - \rho) \right) .
\]  

(A.22)

Although (A.22) gives the general result for the reflection traveltime for an arbitrarily located source-receiver couplet in terms of the unprimed parameters, it is not yet in a
Figure A.3. A line of length L extending from the origin to some termination point P. For display purposes, we have translated the primed coordinate system defined in Figure A.2 to the unprimed origin.
form amenable to an analysis of the crossdip problem. We now seek to express $t'$ in terms of the inline dip angle, $I$, and the angle of crossdip, $C$ (Figure A.4). In the following analysis, we assume that $I$ and $C$ may take on both positive and negative values on the closed interval $(0^\circ, 90^\circ)$ (the corresponding restrictions imposed on the direction cosines $\alpha, \beta$, and $\phi$ ensure that no singularities arise in manipulating the system of equations \{(A.10), (A.23) and (A.24)\} below).

From the figure it is easy to see that

$$\tan I = \frac{\cos \alpha}{\cos \phi}, \quad (A.23)$$

and that

$$\tan C = \frac{\cos \beta}{\cos \phi}. \quad (A.24)$$

From these last two expressions and (A.10), we obtain

$$\cos \alpha = \frac{\tan I}{\sqrt{1 + \tan^2 I + \tan^2 C}}, \quad (A.25)$$

Note that we have used the fact that

$$\text{sgn} (\cos \alpha) = \text{sgn} (\tan I), \quad (A.26)$$

where

$$\text{sgn}(x) = \begin{cases} +1, & x > 0 \\ -1, & x < 0 \end{cases} \quad (A.27)$$

In writing equation (A.26), we have considered (A.23) together with the fact that $\cos \phi \geq 0$ for the domain of definition of $\phi$.

Similarly, the corresponding expression for the $y$ direction cosine is found to be

$$\cos \beta = \frac{\tan C}{\sqrt{1 + \tan^2 I + \tan^2 C}}, \quad (A.28)$$
Figure A.4: Definition of the crossdip angle, $C$, and the inline angle, $I$. 
while the dip angle $\phi$ is related to the inline dip and crossdip by

$$
\sin \phi = \sqrt{\frac{\tan^2 I + \tan^2 C}{1 + \tan^2 I + \tan^2 C}}.
$$

Having expressed the direction cosines in terms of the angles of inline dip and crossdip, we perform a similar analysis for the dip line azimuth, $\theta$. From Figure A.1, we see that

$$
\cos \theta = \frac{\cos \alpha}{\sin \phi},
$$

and

$$
\sin \theta = \frac{\cos \beta}{\sin \phi}.
$$

We may use equations (A.25), (A.28) and (A.29) to rewrite these last two expressions:

$$
\cos \theta = \tan I
$$

and

$$
\sin \theta = \tan C.
$$

Using the double angle formula

$$
\cos (\theta - \rho) = \cos \theta \cos \rho + \sin \theta \sin \rho,
$$

together with (A.32) and (A.33), we find that

$$
\cos^2 (\theta - \rho) = \frac{(\tan I \cos \rho + \tan C \sin \rho)^2}{\tan^2 I + \tan^2 C}.
$$

By substituting these expressions for $\cos \alpha$, $\cos \beta$, $\sin \phi$, and $\cos^2 (\theta - \rho)$ (equations (A.25), (A.28), (A.29) and (A.35), respectively) into (A.22), we obtain, after some
simplification,
\[ t'^2 = \left\{ t_c - \left( \frac{2}{V\gamma} \right) \left( X \tan I + \Psi \tan C + \frac{B\delta}{2} \right) \right\}^2 + \frac{B^2}{V^2} \left( 1 - \left( \frac{\delta}{\gamma} \right)^2 \right) , \]  
(A.36)

where we have defined
\[ \gamma \equiv \sqrt{1 + \tan^2 I + \tan^2 C} , \]  
(A.37)
\[ \delta \equiv \tan I \cos \rho + \tan C \sin \rho , \]  
(A.38)
and
\[ t_c \equiv \frac{2d}{V} . \]  
(A.39)

Note that \( t_c \) is the two-way zero-offset traveltime measured at the unprimed origin.

A hypothetical common midpoint scatter plot associated with a single CMP bin is presented in Figure A.5. We have sketched in two midpoints, both displaying considerable transverse offset \( Y \). Note that for any midpoint which lies inside the bin, we have the relation
\[ X = -\frac{B}{2} \cos \rho , \]  
(A.40)
\[ \Psi = -\frac{B}{2} \sin \rho + Y , \]
so that (A.36) reduces to
\[ t'^2 = (t_c + pY)^2 + \frac{B^2}{V^2} \left( 1 - \left( \frac{\delta}{\gamma} \right)^2 \right) , \]  
(A.41)

where we have defined
\[ p \equiv \frac{2\tan C}{V\gamma} . \]  
(A.42)

Equation (A.41) is the result presented in Chapter 5, equation (5.7).
Figure A.5. A common midpoint scatter plot. The CMP bin is centre is located at the coordinate origin. Two midpoints associated with different shot-receiver pairs are shown in the figure. Both midpoints are substantially offset from the bin centre.
Appendix B: Demonstration that the crossdip-corrected zero-offset section effectively images the inline component of dip

We recall from Chapter 5 that the crossdip correction effectively restores reflection traveltimes within an NMO-corrected CMP gather to the normal incidence two-way traveltime measured at the bin centre $t_c$. Hence an expression for the time slope (apparent slowness) $\frac{\delta t}{\delta x}$ of a 3-D dipping event on a crossdip-corrected zero-offset section may be derived by considering the traveltime difference between two normally incident rays which intersect the slalom line at different points. With reference to Figure A.1, a simple geometric argument reveals that $d$, the perpendicular distance down to the reflecting plane measured at the origin, differs in length from its counterpart measured at a distance $\delta x$ along the slalom line (i.e., the $x$-axis) by an amount

$$\delta x \cos \alpha \quad . \tag{B.1}$$

The resulting discrepancy in normal incidence two-way traveltime $\delta t$ is given by

$$\delta t = \frac{2\delta x \cos \alpha}{V} \quad , \tag{B.2}$$

from whence we can determine the time slope:

$$\frac{\delta t}{\delta x} = \frac{2 \cos \alpha}{V} \quad . \tag{B.3}$$

From Figure A.4,

$$\sin I = \frac{\cos \alpha}{\sqrt{\cos^2 \alpha + \cos^2 \phi}} \quad , \tag{B.4}$$

a result we can rewrite using (A.28) and (A.10):

$$\sin I = \frac{\cos \alpha}{\sqrt{1 + \tan^2 I + \tan^2 \phi}} \quad . \tag{B.5}$$
From this last expression we see that in the limit of zero crossdip, \( \sin I = \cos \alpha \). Thus from (B.3) we recover the well-known result for the time slope associated with a reflector exhibiting pure inline dip:

\[
\left( \frac{\delta t}{\delta x} \right)_{\text{inline}} = \frac{2 \sin I}{V} \tag{B.6}
\]

(e.g. Hale, 1984). Moreover for a significant range of modest crossdips (say, \(-20^\circ < C < 20^\circ\), although the range increases with increasing magnitude of inline dip), we see from equation (B.5) that

\[
\sin I \approx \cos \alpha \tag{B.7}
\]

and thus from equation (B.3) that

\[
\frac{\delta t}{\delta x} \approx \frac{2 \sin I}{V} \tag{B.8}
\]

In other words, a first order estimate of the inline component of dip angle \( I \) may be obtained by simply measuring the time slope on the crossdip-corrected zero-offset section. This result is in harmony with our intuition.

As an example, we consider one of the dipping reflectors associated with the oblique traverse of the Pasayten fault. We orient our coordinate axes such that the \( x-z \) plane contains the zero-offset section, with the positive \( x \)-axis pointing in the direction of increasing station number. By direct inspection of the zero-offset section (Figure 8.6a) and using (B.8), we estimated an inline dip angle \( I \) of 40\(^\circ\) (we assumed a constant velocity of 5600 m/s in this calculation). The optimum crossdip parameter was determined from the CSS analysis to be 0.00013 s/m. From (A.42), we calculated a crossdip angle \( C \) of 15\(^\circ\). Substitution of these results for \( C \) and \( I \) into (B.5) revealed that \( \sin I = 0.98 \cos \alpha \), suggesting that we were justified in measuring the inline dip angle from the zero-offset section. From (A.32) and (A.29) we calculated a dip line azimuth \( \theta \) of 18\(^\circ\) and a true dip angle \( \phi \) of 41\(^\circ\). Taking into account the fact that our \( x \)-axis was oriented at [N40\(^\circ\)E], we inferred that the feature dips 41\(^\circ\) [E30\(^\circ\)N].