2D and 3D Reflection Seismic Studies over Scandinavian Deformation Zones

EMIL LUNDBERG

The study of deformation zones is of great geological interest since these zones can separate rocks with different characteristics. The geometry of these structures with depth is important for interpreting the geological history of an area. Paper I to III present 2D reflection seismic data over deformation zones targeting structures in the upper 3-4 km of the crust. These seismic profiles were acquired with a crooked-line recording geometry. 2D seismic processing assumes a straight recording geometry. Most seismic processing tools were developed for sub-horizontally layered structures. However, in the crystalline rocks in Scandinavia more complex structures with contrasting dip directions and folding are common. The crooked-line recording geometries have the benefit of sampling a 3D volume. This broader sampling can be used to gain knowledge about the true geometry of subsurface structures. Correlation with geological maps and other geophysical data along with seismic data modeling can be used to differentiate reflections from faults or fracture zones from other reflectivity, e.g. mafic bodies. Fault and fracture zones may have a large impedance contrast to surrounding rocks, while ductile shear zones usually do not. The ductile shear zones can instead be interpreted based on differing reflectivity patterns between domains and correlations with geology or magnetic maps. Paper IV presents 3D reflection seismic data from a quick-clay landslide site in southern Sweden. The area is located in a deformation zone and structures in unconsolidated sediments may have been influenced by faults in the bedrock. The main target layer is located at only 20 m depth, but good surface conditions during acquisition and careful processing enabled a clear seismic image of this shallow layer to be obtained. The research presented in this thesis provides increased knowledge about subsurface structures in four geologically important areas. The unconventional processing methods used are recommended to future researchers working with data from crooked-line recording geometries in crystalline environments. The imaging of shallow structures at the quick-clay landslide site shows that the 3D reflection seismic method can be used as a complement to other geophysical measurements for shallow landslide site investigations.

Keywords: Azimuthal binning, Crooked-line geometry, Cross-dip, Fault zone, Hard rock seismics, MTFC, Quick clay, Shear zone, UDZ

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This thesis is based on the following papers, which are referred to in the text by their Roman numerals.


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Additional publications during my PhD studies, which are not included in this thesis:

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<th>Abbreviation</th>
<th>Description</th>
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<tbody>
<tr>
<td>2D</td>
<td>Two-dimensional</td>
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<tr>
<td>3D</td>
<td>Three-dimensional</td>
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<tr>
<td>CDP</td>
<td>Common Depth Point</td>
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<tr>
<td>CMP</td>
<td>Common Mid Point</td>
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<tr>
<td>NMO</td>
<td>Normal Moveout</td>
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<td>DMO</td>
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<td>s</td>
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<td>ms</td>
<td>millisecond</td>
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<tr>
<td>TWT</td>
<td>Two Way Time</td>
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<tr>
<td>S/N</td>
<td>Signal-to-Noise ratio</td>
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<tr>
<td>ka</td>
<td>thousand years</td>
</tr>
<tr>
<td>Ma</td>
<td>million years</td>
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<td>Ga</td>
<td>billion years</td>
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1. Introduction

Geophysics is the study of the Earth structure, its physical condition and evolution based on the methods and principals of physics. The first seismographs were developed in the late 1800s and early 1900s. These instruments could measure the arrival time of elastic waves generated by earthquakes at large distances. Pressure waves travel faster than shear waves and these waves were, therefore, also called P-waves (primary wave) and S-waves (secondary waves), respectively. By observing P- and S-wave arrival times for different earthquakes around the globe, the inner structure of the Earth was revealed. The presence of the Earth's core was detected by Richard Oldham in 1906 and in 1909 Andrija Mohorovic showed the existence of a velocity discontinuity separating the crust and the mantle (the Moho). Controlled source seismic exploration began in the 1920s and 1930s for oil prospecting (Shearer, 1999). Controlled source seismic methods can be used for detecting velocity variations in Earth materials e.g. with refraction seismic and traveltime tomography. The reflection seismic method detects interfaces between Earth materials. Changes in the subsurface seismic velocity and density (the material impedance) cause seismic waves to reflect back at an interface if a sharp enough impedance contrast is encountered. Multi-source and -receiver common midpoint stacking was patented in 1956 and since then the quality of reflection seismic profiles has improved due to the increased signal-to-noise ratio acquired (Shearer, 1999). The reflection seismic method can be used in both near-surface studies as well as deep crustal studies with penetration depth ranging between 10s of meters to Moho depth's of 50 km or even more. What depth that can be targeted depends on the power of the source generating the seismic waves at the surface, as well as the properties of the subsurface materials.

The principal objective of this thesis was to apply 2D and 3D reflection seismic imaging methods over geologically interesting areas in Scandinavia. Seismic data acquired over the crystalline hard rock environment in Scandinavia requires special attention to processing due to the significantly different subsurface conditions encountered compared with the sedimentary environments, often subhorizontal structures, for which most standard seismic processing techniques were developed. Acquisition along crooked 2D lines furthermore challenges the seismic data processing, but it also enables 3D information to be extracted from the data, which is especially useful in the crystalline environment, where structure orientation can vary rapidly both
laterally and with depth (Rodriguez Tablante, 2006). Deformation zones are zones where rocks have been altered by tectonic forces including processes such as faulting, folding, shearing, extension and compression (MacDonald and Burton, 2003). These are of high geological interest since these zones can separate rock units with different characteristics. Deformation zones are also where movement has occurred and, hence, the geometry of a deformation zone can help us unravel the geological history of an area. Most of the permeability of crystalline bedrock is also located along fault zones and fracture zones and this is where gas and fluids can be transported, which is important for constructions, infrastructure and near surface environmental aspects (Twiss and Moores, 2001). Interpretation of fault zones from reflection seismic data in the crystalline environment can be challenging. In sedimentary environments a fault can be detected as laterally continuous reflectors are offset along a steep fault. In the crystalline environment laterally continuous reflectors are rare. The surface reflection seismic method, however, generally reveals subhorizontal structures more easily than steeply dipping structures (Harjes et al., 1997). Different geophysical techniques are, therefore, usually required for correlating seismic reflections with faults. Therefore, one objective of the thesis was to apply different geophysical techniques, as well as seismic modeling, to improve the seismic data interpretation.

In this thesis I present new reflection seismic data from four different areas in southern Scandinavia. The first three papers (Papers I-III) are focused on the Ullared Deformation Zone, the More-Trøndelag Fault Complex and the Dannemora area, and target the upper 3 to 4 km of the crust. The main objective of these studies was to increase the understanding of deformation zone geometries at depth in order to better understand the geology of the areas. In the Dannemora area one of the seismic profiles crossed the Dannemora mine, however, my contribution to Paper III consisted mainly of processing and interpreting a seismic profile connecting the Dannemora area with the Forsmark area, and I will, therefore, focus on that part of Paper III.

In Paper IV the main target was the near surface normally unconsolidated sediments above the crystalline bedrock, located at about 30 to 80 m depth. The study area of Paper IV is located in a major deformation zone. In this study the deformation zone has been important for shaping the area and possibly also for affecting some of the features observed in the sediments. In this study 3D reflection seismic data were collected for mapping lateral discontinuities in layers, within the normally consolidated sediments, beneath a landslide scar. The biggest challenge was that the main target horizon is located at a very shallow depth, about 20 m or less, below the surface.

The main part of the research is presented in chapter 5 as paper summaries. In the preceding chapters I provide a background to the geology of the study areas in chapter 2 and also give a background on deformation zones in chapter 3. In chapter 4 I introduce the reflection seismic method and discuss the non conventional processing methods that were applied in my research as
well as the modeling that was conducted in Paper II. These chapters will hopefully help the readers understand why we study deformation zones and how the geology in Scandinavia, and especially in deformation zone areas, affects the seismic data processing and interpretation. After the paper summaries in chapter 5 I draw some conclusions from my work. I will also discuss some recent data from my study areas.
In this chapter I will give a brief overview of the geology in Scandinavia with emphasis on the southern region where the study areas of my research are located. Figure 2.1 shows the locations of my study areas. These have many similarities. The Precambrian basement in Scandinavia was formed and later deformed and metamorphosed during several orogenic episodes (Lindström et al., 2000). The rocks may, therefore, display complex structures such as multiple folding and faulting. The study areas are also located in, or across, fault zones or deformation zones and they also share the more recent history with glaciations covering Scandinavia, the latest disappearing around 10 ka ago. The glaciations covering Scandinavia have shaped the landscape and formed unconsolidated sedimentary layers on top of the bedrock (Lindström et al., 2000). These unconsolidated materials are of importance for the seismic processing that is required, and in paper IV it is the main target of the investigation.

Figure 2.1. Lithotectonic units of the Sveconorwegian orogenic belt with major shear zones marked. Modified after Bingen et al. (2008a). The study areas of my research are marked with white stars. The Bergslagen region in the Fennoscandia Foreland is separated by two tectonic domains characterized by high ductile strain shear zones. The study area of Paper III is partly located within this high ductile strain zone.
2.1. Geology of Scandinavia

The rocks in Sweden and Norway constitute a part of the Fennoscandian Shield (or Baltic Shield). The Fennoscandian rocks are present between the Kola Peninsula, in Russia, and northern Norway through Finland and Sweden and to southern Norway. The Swedish geology (not always confined to the Swedish borders) is described in e.g. Lindström et al. (2000) and is summarized here. The oldest rocks are found in the northeast, with ages of about 3 Ga. The rocks then generally become younger towards the southwest. The youngest rocks (of about 1 Ga) are found in southwestern Sweden and southern Norway. The rocks with ages between 4.6 Ga (the formation of the Earth) and about 545 Ma belong to the Precambrian. Some younger Phanerozoic rocks (0 to 545 Ma) also occur in Scandinavia e.g. in southern Sweden (Skåne) and on the islands Öland and Gotland and in the Oslo fjord. Also some Phanerozoic rocks, preserved from erosion in downfaulted blocks due to impact of meteorites, can be found in Sweden, such as those around the Lockne, Siljan and Dellen lakes. The most prominent Phanerozoic rocks are, however, the Caledonides which form a 100 to 300 km wide belt from northern Norway to Stavanger in the south. Most of the Caledonides are located in Norway, but the eastern part stretches between Treriksröset and the northern part of Dalarna in Sweden. The Caledonides formed roughly 400 Ma ago by thrusting of sheets from the west towards the east (see e.g. Gee, 1975).

2.2. Sveconorwegian orogen

The rocks in the southwestern part of Sweden and Norway were deformed and metamorphosed during the Sveconorwegian orogen between 1140 Ma and 900 Ma (Bingen et al., 2005; Bingen et al., 2008a and Bingen et al., 2008b). The Sveconorwegian belt includes the Western Gneiss Region (WGR) the Jotun and Lindås nappes and large basement windows in western Norway (Tucker et al., 1990; Bingen et al., 2001; Corfu and Andersen 2002; Skår and Pedersen 2003; Lundmark et al., 2007; Lundmark and Corfu, 2008). Although, the pre-Caledonian linkage to Fennoscandia is speculative, these units have clearly been affected by Sveconorwegian reworking. The orogen resulted from a multiphase collision between Fennoscandia and an unknown continent, possibly Amazonia. The main part of the Sveconorwegian belt consists of five lithotectonic units (Figure 2.1). The Eastern Segment can be directly linked to the Fennoscandian foreland (Wahlgren et al., 1994; Persson et al., 1995 and Söderlund et al., 1999), while the other units (the Telemarkia, Bamble, Kongsberg and Idefjorden Terranes) have been transported significantly relative to Fennoscandia. The five units are separated by crustal scale shear zones.
2.3. The Ullared Deformation Zone (UDZ)

The UDZ is located in the southern part of the Eastern Segment of the Sveconorwegian province close to the Mylonite Zone. The Eastern segment consists of Transcandinavian Igneous Belt (TIB) rocks, with protolith ages of c. 1730-1660 Ma, that were underthrust, reworked and exhumed during the Sveconorwegian orogeny (Bingen et al., 2008b; Möller et al., 2007; Wahlgren et al., 1994). The TIB rocks are mainly granitoid and porphyritic rocks located along a north-south belt characterized by high magnetic anomalies (Lindström et al., 2000). The UDZ is an approximately 10 km wide shear zone system. Both mylonitic shear zones and gneissic foliation are aligned along the extension of the shear zone system (Möller et al., 1997). The UDZ is one of a few structures worldwide known to contain decompressed eclogite facies rocks of Precambrian age. Metamorphism leading to eclogitization has been dated to a maximum of 972 ± 14 Ma, (Johansson et al., 2001) and a minimum of 956 ± 7 Ma (Andersson et al., 1999; Möller and Söderlund, 1997), i.e. during the last stage of the Sveconorwegian orogeny (the Dalane phase) that was dominated by relaxation and extension as the Sveconorwegian belt collapsed (Bingen et al., 2008b).

2.4. Møre-Trøndelag Fault Complex (MTFC)

Most of the onshore part of the MTFC is located in the WGR. The WGR is composed of Fennoscandian gneisses, mainly 1650-950 Ma orthogneisses (Austrheim et al., 2003; Skår and Pedersen, 2003), that were reworked and exhumed in the final stage of the Caledonian orogeny. The WGR is structurally overlain by continental and oceanic allochtons. These rocks were downfolded into the basement gneisses and can now be observed as long lobes and tongues extending E-W to ENE-WSW (Braathen et al., 2000; Terry and Robinson, 2003; Hacker et al., 2010). The MTFC probably formed during the final stage of the Caledonian Orogeny in the Scandian phase (Grønlie and Roberts, 1989). Main phases of activity include Devonian sinistral strike-slip, early Permian sinistral transtension and late Jurassic normal to dextral strike-slip faulting (Grønlie and Roberts, 1989; Séranne, 1992; Sherlock et al., 2004). Redfield et al. (2005) suggested possible Cenozoic normal dip-slip, reflecting uplift of the Norwegian mountains while offshore basins were subsiding. The MTFC consists of several fault segments. The Tjellefonna fault is located to the south and was defined based on several fault localities mapped along a strong topographic lineament (Redfield and Osmundsen, 2009).
2.5. The Dannemora area

The Bergslagen region is located in the south-western part of the Svecokarelian orogen in the Fennoscandian Shield and is one of the most important mining districts of Sweden. A thorough description of the geology of the Bergslagen district can be found in Stephens et al. (2009), and a short summary of the geology in the Dannemora area is provided here. Exposed rocks are mainly intrusive, meta-intrusive or metavolcanic rocks formed 1.9 and 1.8 Ga. These rocks were affected by Svecokarelian (1.9-1.8 Ga) deformation and metamorphism, and in the western part to some extent also by Sveconorwegian (1.0-0.9 Ga) deformation and metamorphism. The Bergslagen region is bounded to the north and south by ductile high strain tectonic belts that anastomose around tectonic lenses with lower strain. The bedrock is commonly folded and affected by lower ductile strain within these lenses. These high strain belts strike WNW-ESE with subvertical dip. The Forsmark proposed site for storage of spent nuclear fuel is located inside such a low strain tectonic lens (Svensk Kärnbränslehantering AB, 2008). Further south is the Dannemora mine that is situated to the west of the Skyttorp-Vattholma Fault Zone (SVFZ) and the Österbybruk Deformation Zone (ÖDZ). The northern part of the Dannemora study area (Paper III) belongs to the tectonic domain consisting of high strain ductile shear zones around the Forsmark area (Stephens et al., 2009).

2.6. Surface geology in Scandinavia

There is not much evidence of how the landscape in Scandinavia looked like in the past. The area has been above sea level for much of the last 400 Ma and older mountain chains have been eroded for a long time. Only remnants of the latest (Caledonian) orogeny can still be seen. The surface geology of Sweden is largely shaped by the repeated glaciations in the Pleistocene (Lindström et al., 2000). Erosion occurs mainly due to chemical and biological weathering of the rocks (Nichols, 1999). In some regions with hot and humid climate the weathered layer can reach thicknesses of up to hundred meters (e.g. Koita et al., 2013). When the rocks are uplifted due to tectonic processes the weathered material can be washed away by wind and water. The eroded material is then deposited along streams and rivers or lakes and in the sea. Water is essential for the weathering process (Nichols, 1999) and, therefore, weathering can potentially be more effective in fractured rock where water is more abundant. In Sweden valleys have formed along eroded fracture zones, but in some cases these valleys have been filled with glacial sediments and are not clearly visible on the surface. Rivers can locally be correlated with fracture/fault zones and lakes can be found between faults, e.g Lake Vättern in southern Sweden (Fredén, 2002). Many large rivers and
fjords in Sweden and Norway follow fault zones such as the Göta River along the Götaälv Zone (Paper IV) and several fjords follow the different segments of the MTFC (Paper II). The Pleistocene ice sheets removed weathered materials from the rocks and, therefore, the weathering layer on the crystalline rocks in Scandinavia is thin (Lindström et al., 2000). The weathered materials were deposited elsewhere by different glacial and post glacial processes. Material transported by the ice was not sorted and often a thick till, consisting of a variation of different grain sizes from large boulders to fine grained clay particles, covers the bedrock. Eroded materials that were transported by rivers were sorted so that finer materials reached furthest to the lakes or seas and the coarser materials were deposited in streams and rivers (Nichols, 1999). Most of the southern parts of Sweden were covered by sea water or lakes because the land was suppressed by the isostatic load of the ice during the ice sheet regression. Therefore, we can find thick layers of clay in areas that are now well above the current sea level. Due to seasonal variations of melting of the ice, during the ice sheet regression, layering between finer and coarser materials occurs (Lindström et al., 2000).

2.7. Götaälv Zone (GÄZ)

The GÄZ is an important structural boundary in the southwestern gneiss region of Sweden (Bingen et al., 2008b, Park et al., 1991). The deformation zone is gently west dipping and has been reactivated repeatedly through geological time. The latest reactivation probably occurred during the last stage of the Sveconorwegian orogeny (at c. 1 Ga) when the western block moved towards the west (Lindström et al., 2000). The Göta River follows the GÄZ between Lake Vänern and Gothenburg. Thick layers of clays and sand/silt have been deposited along the shorelines of the Göta River during and since the last glaciation c. 10 ka. Some of these clays were deposited in a marine environment due to the isostatic effects of the heavy ice sheet. When the ice sheet melted these clays were uplifted, due to isostatic rebound, and were, therefore, exposed to fresh water. Fresh water may leach salt ions from the marine clay making the clay structure more sensitive. The highly sensitive clays are referred to as quick clays. Further details on quick clay formation can be found in Rankka et al. (2004) and in Paper IV. In the study area the bedrock is a gneiss-granite to granodiorite and it is exposed just south of the 3D area. Boreholes from drillings made by the Swedish Geotechnical Institute (SGI) indicate clays (including quick clays) in the upper 20 m (Löfroth et al., 2011) above a sandy-silty layer. Permeable layers may be an important geological feature for quick clay formation, since they can lead fresh water into the less permeable clay formations (Rankka et al., 2004).
3. Deformation zones

Deformation zones are zones where rocks have been deformed by processes such as faulting, folding, shearing, extension and compression (MacDonald and Burton, 2003). These zones are of great importance for several reasons. Deformation zones can separate different rock units from one another and the geometry of a deformation zone can, hence, give clues to the geological history of an area. In active tectonic regions, movements occurring along faults can cause earthquakes. In areas with less tectonic activity a fault or fracture zone can be an important channel for gas and fluids and therefore influence underground constructions such as tunnels and sites for storage of spent nuclear fuel. Deformation zones can also shape the landscape by creating steep cliffs or forming a depression where rivers flow.

3.1. Definition of faults

A thorough description of faults can be found in e.g. Twiss and Moores (2001) or Davis and Reynolds (1996) and a summary of these is provided here. A fault separates two sides that have been displaced, relative to each other, parallel to the fault plane or fault surface. The scale of a fault can range over a distance of meters to thousands of kilometers. Fault like features in the scale of centimeters are often instead referred to as shear fractures. Movement often occurs along several closely spaced fault surfaces and these fault surfaces then form a fault zone. The largest fault zones include tectonic plate boundaries. Faults can be divided into three main groups depending on the orientation of the relative displacement or slip. In Figure 3.1 the characteristics of different types of faults are shown. Dip-slip faults have the main slip component approximately parallel to the dip of the fault plane and strike-slip faults have the main slip component approximately parallel to the strike of the fault plane. Oblique-slip faults have both a dip-slip component and a strike-slip component. Faults can then be further divided into groups depending on the relative movement along them. Dip-slip faults, where the hanging wall block has moved down relative to the footwall block are called normal faults and when the hanging wall block has moved up relative to the footwall block we call it a reverse fault or a thrust fault depending on its dip angle. Strike-slip faults are defined as right-lateral (dextral) or as left-lateral (sinistral) depending on in which direction the block on the oppo-
site side of the fault trace has moved. Oblique faults can then be described using the sum of the dip-slip and strike-slip component. Rotational faults have the slip component changing rapidly along the distance of the fault. What type of fault that develops in a certain setting depends largely on the stresses that the rock unit is subjected to. In general, normal faults develop during extension and reverse faults develop during compression. Strike-slip faults can for example develop as a transfer fault separating different domains of normal faulting in extensional provinces. The stresses by which rocks are subjected to can, however, change both laterally and horizontally. Ramsay (1967), for example, described the process of tangential longitudinal folding, where extension occurs above and compression below a neutral surface. Carminati et al. (2001) modeled the stress accumulation in central Italy and explained the coexistence of compressional and extensional seismic events by introducing crustal scale discontinuities (i.e. faults).

![Faulted blocks showing the displacement for different types of faults.](image)

**Figure 3.1.** Faulted blocks showing the displacement for different types of faults. Modified from Twiss and Moores (2001).

In which setting the faulting takes place will also affect what type of fault that is formed (Passchier and Trouw, 2005; Twiss and Moores, 2001). In the near surface (approximately 1-10 km) a fault zone consists of cataclastic rocks. At deeper levels, approximately when the temperature reaches 250-350° C, the fault zone will be mylonitic. Figure 3.2 shows the how the characteristics of a fault change with depth. Cataclastic fault rocks are crushed and shattered by brittle deformation. The grain size can vary and individual fragments are usually angular and sharp. Cataclastic fault rocks usually lack
internal linear or planar structures. Shear fractures may also develop in some cases. The dominating set of shear fractures are aligned parallel to the fault, but a conjugate set at an angle of approximately 60° may also form (Twiss and Moores, 2001). These shear fractures form a fracture zone. Fracture zones may accompany a fault, but fracture zones can also be associated with folding or igneous intrusions. Mylonitic fault rocks have a strong planar or linear texture subparallel to the fault plane and are formed by recrystallization of mineral grains during ductile deformation. The original grain size is reduced to a very fine-grained matrix although relict mineral grains can occur within the fine-grained matrix (Passchier and Trouw, 2005). The width of a fault zone in general increases with depth, from brittle faults in the near surface to narrow ductile shear zones and then to wide ductile shear zones with striped gneiss at depth (see Passchier and Trouw, 2005 for further details of different fault rocks).

Figure 3.2. Schematic block diagram, showing a fault zone and its change with depth. The width of the fault zone is increasing with depth and the rock type changes from cataclastic in the upper 10-15 km to mylonitic in the deeper part (see text for explanation). Modified after Twiss and Moores (2001).
3.2. Deformation zones in Scandinavia

In old geological rocks, such as those in Scandinavia, deformation zones formed at a deep crustal level are exposed on the surface due to unroofing processes and millions of years of erosion. The deformation zones may, however, have been reactivated several times in the geologic history and, therefore, the deformation zones on the surface may include cataclastic rocks and mylonitic rocks as well as strongly folded rocks. The deformation zones may have been reactivated during different stress regimes and several different movements along the deformation zones may have occurred. The different stages of deformation can sometimes be seen by detailed microscopic analysis of microstructures (Passchier and Trouw, 2005). Deformation zones can sometimes show up as a strong linear feature in the surface topography. For example, the post-glacial Pärvie fault in the north of Sweden that was formed or reactivated during the regression of the ice-sheet at 10 ka ago, can in some places show a 5-10 m step in the topography (Lundqvist and Lagerbäck, 1976; Lagerbäck, 1979). Other deformation zones can show up as low topographic lineaments eroded by water flowing in rivers, such as the Göta River following the Götaälv zone (see Paper IV) and also several fault segments of the More-Trondelag Fault Complex appear to coincide with fjords (see Paper II). However, in some cases the deformation zones cannot be detected without the aid of geophysical methods, especially when the rocks are covered by glacial and post-glacial sediments and the exposure is limited.

![Figure 3.3](image)

Figure 3.3. Example of deformation zones in Scandinavia. A ductile shear zone that formed at large depth is now exposed near the surface. The deformation zone may have been reactivated at shallower depth, by brittle faulting and fracturing. Pleistocene glaciations may have enhanced the surface expression of the deformation zone (a) or covered the deformation zone with sediments (b).

Figure 3.3 shows a general view of deformation zones in Scandinavia. A ductile deformation zone formed at a depth of at least 10-15 km (Figure 3.2) is presently located in the near surface. The deformation zone may have been reactivated with brittle faults and fractures overprinting the ductile deformation. Strong fracturing may lead to increased weathering and erosion of the fault zone and a topographic low may form. In other cases the fault zones
may have been completely covered by glacial and post-glacial sediments. At greater depth the deformation zone can be wider and if the strain rate is low the grain size can be coarse and the rock may appear weakly deformed (Passchier and Trouw, 2005). Ductile shear zones can, therefore, be difficult to recognize in high grade gneisses, since all rocks are deformed to some extent. The rocks in a high grade ductile shear zone may exhibit a compositional layering forming a ribbon mylonite or striped gneiss (Passchier and Trouw, 2005). In Scandinavia, gneisses are common. Most of the Precambrian bedrock in Sweden has been metamorphosed during successive orogenic events. Deformation has led to faulting and folding of the rocks and at larger depths minerals may be aligned forming a linear or planar fabric typical for the gneissic rocks, common in the Western Gneiss Region (Paper II) and in the Eastern Segment of the Sveconorwegian orogenic belt (Paper I).

3.3. Geophysical techniques for deformation zone investigations

The geometry of a fault plane or shear zone and the kinematics (slip components) can be revealed by detailed analysis of fault plane features and microtectonic structures of exposed deformation zones (Passchier and Trouw, 2005). To reveal the continued geometry at depth, however, geophysical measurements are needed. In this section some geophysical methods that can be used for fault detection and deformation zone investigation will be briefly introduced. It is always an advantage if more than one geophysical method is used in order to reduce the ambiguities in each method.

The magnetic susceptibility of a rock describes to what extent the rock can be magnetized by a magnetic field. The magnetization can be measured using a magnetometer. Magnetometers can be used both on the ground and also placed on a helicopter or aircraft. Airborne geomagnetic measurements can cover large areas in short time and, hence, give valuable regional data for mapping geological rock boundaries and structural features (Telford et al., 1986). Gravity measurements are commonly carried out to complement magnetic surveys. Variations in the Earth's gravitational field are caused by differences in the density of rocks in the subsurface. It is the acceleration due to gravity that is measured by the use of a gravimeter or accelerometer. Gravity surveying can be performed on land but also as an airborne method with similar advantages as the geomagnetic airborne measurements (Telford et al., 1986). Faults usually show up as linear features on magnetic or gravity anomaly maps. If the fault also separates two rocks with different magnetic properties or density, the linear feature is accompanied by an abrupt change in the magnetic or gravity signature. A fault may, however, show up as either a magnetic high or as a magnetic low anomaly depending on the fault
properties as well as on the properties of the surrounding rock. Figure 3.4 shows an example where the fault does not show a clear gravity signature, except for in the southern part where the fault appears to truncate and offset a large positive gravity anomaly. A detailed discussion about anomalies in magnetic and gravity data over various fault structures can be found in Prieto (1996).

*Figure 3.4. Bouguer gravity anomaly map of the Dannemora area (courtesy of Geological Survey of Sweden). A sharp change in the gravity anomaly marked by the purple ellipse, is observed on the eastern and western sides of the Skyttorp-Vattholma Fault Zone.*
The resistivity method measures variations in the subsurface electrical resistivity. An electrical current is transmitted to the ground between two electrodes and the electrical potential can then be measured between other electrodes. Most rock minerals are insulators and the most important factor for resistivity measurements in crystalline rocks is, therefore, usually the water content. Fractured rock containing water or clay-weathered rocks can show much lower resistivity than fresh crystalline rock, making the resistivity method a great tool for fault and fracture zone detection (Loke, 2004). A deformation zone composed of ductile shear zones can be detected on magnetic maps as a zone with strong linear pattern (Paper I and Paper III).

3.4. How to interpret deformation zones in crystalline rocks from reflection seismic data

Numerous examples of seismic data acquired over deformation zones can be found in the literature. Due to the differences in characteristics and geometry, as well as the availability of other geological and geophysical data, each study provides a different approach in terms of interpretation. Observations from borehole logging have shown that fluid filled fracture zones (Green and Mair, 1983; Lüschen et al., 1996; Juhlin and Stephens, 2006), mylonite zones and ductile shear zones (Christensen and Szymansky, 1988; and Xu et al., 2009) can contribute to increased crustal reflectivity. However, also other sources for increased crustal reflectivity occur, such as, dolerite sills (Juhlin, 1990) and contrasts within metasedimentary rocks (Miao et al., 1994). The main task for an interpreter is to distinguish the different types of reflectivity from each other.

Brittle fault zones or fracture zones (that may be related to a fault zone) can occur in the upper 10-15 km of the crust. These zones can have a large impedance contrast to the intact rock. Ayarza et al. (2000) showed that fluid filled fractures may also generate significant P-S converted energy. Fluid filled fracture zones can be prevalent as deep as 12 km (Ganchin et al., 1998).

Mylonitic rocks have a small impedance contrast to surrounding rocks and could potentially be more difficult to image with reflection seismics (Fountain et al., 1984). Deformation zones formed in high grade metamorphic rocks may consist of several mylonitic shear zones and the surrounding rock can also be deformed and metamorphosed. Gneissic fabric and elongated mafic bodies are often parallel to the deformation zones. Hynes and Eaton (1999) noticed that domain boundaries, which could often be correlated to shear zones and faults, are rarely associated with reflections stronger than the internal reflectivity of the domains. Instead the fault zones and shear zones can be interpreted based on the location of boundaries between do-
mains of contrasting reflectivity. Laminated structures and anisotropic effects, caused by preferred mineral orientation, can increase the amplitude of reflections significantly (Fountain et al., 1984; Jones and Nur, 1984; Christensen and Szymanski, 1988). Wide shear zones may consist of thin mylonitic zones in a highly deformed crystalline rock forming such a laminated structure with a strong preferred mineral orientation.

The ideal situation is when borehole data, intersecting the reflectivity of interest, is available. In other situations, surface geophysical/geological data or geophysical depth sections, must be used for comparison with where reflections project to the surface or near surface. This requires that reflections are dipping and extend to the surface. Although flat lying shear zones and fracture zones may be the simplest seismic target to image, they can be the most difficult ones to interpret.
4. Reflection seismic method

4.1. Seismic data acquisition

The reflection seismic method makes use of near vertical reflected elastic energy from subsurface interfaces. Elastic waves are generated using an explosion, weight-drop, hammer, air gun or a vibrator. A geophone (or receiver) on the ground records the ground motion and this record is referred to as a trace. In 2D data acquisition, source points and receivers are located along a straight line. If dipping structures are expected, the line should be acquired perpendicular to the strike of the geological structures (Yilmaz, 2001). For 3D data acquisition, source and receiver positions are located in a grid and several dip directions will be covered for each source point. The inline and crossline directions and offsets are, however, still important to consider for ensuring that an adequate offset and azimuth distribution is acquired.

4.1.1. 2D Crooked line acquisition

Many of the sources used in reflection seismic data acquisition cannot be used in forested or mountainous areas and can only be moved along roads or trails (vibrators, weight-drops, vehicle mounted hammers, etc.). In these cases the acquisition may follow a crooked line. With a crooked-line acquisition the midpoints are spread laterally within each common midpoint gather. Figure 4.1 illustrates crooked-line acquisition geometry where trace midpoints are spread laterally within a CMP bin. Reflections that originate from interfaces that have a dip direction other than parallel to the stacking line will not stack coherently and the quality of the stack can be significantly reduced. The lateral spread can on the other hand also allow some 3D information to be extracted from the reflection seismic data since the strike and dip of reflections can be estimated through data analysis. This is especially useful in hard rock environments where rapidly changing dip directions, both laterally and vertically, can occur (Larner et al., 1979; Nedimović and West, 2003a; Rodriguez Tablante, 2006; Wu, 1996; Wu et al., 1995).
Figure 4.1. Schematic block diagram, illustrating a midpoint spread of a CMP bin and parameters used in cross-dip analysis; $y_i$ - midpoint offset, $\beta$ - inline dip, $\alpha$ - crossline dip, $V$ - bedrock velocity and $\Delta T$ – cross-dip correction.

4.2. Seismic data processing

Reflected signal and noise, such as wind, rain and vibrations from cars, are recorded in the seismic traces. Special techniques are then used to process the seismic traces in order to enhance the reflected seismic signal and to suppress noise. One way to enhance the seismic signal is to use several geophones for each source record and to use multiple source points. All recorded traces are sorted and traces containing seismic signal from the same subsurface points, common depth point (CDP) are gathered and stacked in order to enhance the reflected seismic signal. The normal moveout (NMO) correction accounts for the delayed arrival times at receivers that are offset from the source. If interfaces are dipping it is better to use the term common midpoint (CMP), since the depth of the reflecting points from a dipping interface will vary within the CMP-gather depending on the source and receiver offset and azimuth. Dip moveout (DMO) migrates the reflected seismic energy to the correct CMP before stacking. Both NMO and DMO, however, assume the reflectors have a dip direction parallel to the stacking line. If the stacking line is oblique to the subsurface structures the dip in the migrated stack will
be underestimated. Detailed descriptions of conventional seismic data processing can be found in Yilmaz (2001).

For crooked-line seismic data the midpoints are spread out away from the stacking line. This midpoint spread challenges the standard straight line 2D processing techniques. However, crooked line data processing has some pitfalls. If the data are stacked along a slalom line, transparent zones may occur where the stacking line is bent, due to low fold and limited offsets (Wu, 1996). This can easily be avoided by stacking along a straight line. Stacking along a straight line also simplifies the interpretation since, reflections will not be curved due to a curved stacking line. It is also possible to treat the 2D crooked-line seismic data as a sparse 3D data set (Nedimović and West, 2003b; Malehmir et al., 2009), although the sparse sampling in the cross-line direction may cause migration artifacts. Using extra source points in the cross-line direction can increase the cross-line sampling and make 3D processing useful, without adding significant extra time and costs during acquisition (Paper III). Another problem may occur in sections where the acquisition line is near perpendicular to the stacking line. These segments will contain several traces with almost the same offset and these will also stack in the same CMP-gather. Ground roll noise or multiples will be stacked coherently regardless of the velocity used for NMO corrections. This problem is most easily avoided by performing a partial stacking of traces with equal offsets or by omitting traces from these perpendicular segments (Wu, 1996).

In the next two sections I will describe two processing methods that I have used in order to extract 3D information from crooked-line 2D data. The cross-dip correction method was introduced by Larner et al. (1979) and has been used successfully in crystalline environments (Wu et al., 1995; Rodríguez Tablante, 2006). Approaches similar to the azimuthal binning analysis have previously been used by Tsumura et al. (2009) and Kashubin and Juhlin (2010) for analyzing the strike of structures in the crystalline environment.

### 4.2.1. Cross-dip correction

Cross-dip corrections of reflection seismic data acquired on a crooked-line as proposed by Wu et al. (1995), is a means of improving the quality of the stacked section. The dip component of a reflector can be divided into an inline-dip component, \( \beta \) and a crossline-dip component, \( \alpha \) (Figure 4.1). For a trace with a midpoint located at a perpendicular distance \( y_i \) from the stacking line a travel-time shift (\( \Delta T \)) will be described by Equation 4.1.

\[
\Delta T = \frac{2 \cdot y_i \cdot \sin(\alpha)}{V} \tag{4.1}
\]

Since Equation 4.1 uses a constant velocity (\( V \)) for the conversion between depth to the reflector and the travel time it is important to consider whether
any significant lateral velocity variations are present. Fortunately the crystal-
line rocks in Scandinavia commonly display negligible lateral velocity varia-
tions unless significant lithological boundaries occur. The cross-dip analysis
is more sensitive to the chosen cross-dip angle than to the velocity (V), espe-
cially for smaller cross-dip angles.

\[ \text{Cross-dip angle} \]

\[ \text{Time shift (ms)} \]

\[ \text{Velocity (m/s)} \]

\[ \text{0} \quad 100 \quad 200 \quad 300 \quad 400 \]

\[ \text{0} \quad 10 \quad 20 \quad 30 \quad 40 \quad 50 \quad 60 \quad 70 \quad 80 \quad 90 \]

**Figure 4.2.** Diagram illustrating the variation in cross-dip angle for different bedrock velocities at certain time shifts. The calculations are based on a midpoint offset of 1000 m. For small cross-dip angles the bedrock velocity may vary significantly from the true velocity and an estimation within 5° of the cross-dip angle is still achieved. For higher cross-dip angles 55°-60° the velocity can vary about 250 m/s from the true velocity for achieving an estimate within 5° of the cross-dip angle.

Figure 4.2 shows the variability between the cross-dip angle and the required time shift for different velocities (V). These were calculated for a perpendicular midpoint offset (y) of 1000 m. For a large cross-dip angle (55-60°) and a low velocity difference of 250 m/s between the true and the estimated bedrock velocity, a 5° cross-dip angle accuracy is obtained. For lower cross-dip angles or smaller midpoint offsets the accuracy of the chosen velocity (V) is even less important. By performing a series of cross-dip corrections for different cross-dip angles (α) and comparing the continuity and amplitude of each reflection the optimum cross-dip angle can be estimated. By using the inline dip and the cross-line dip the true strike and dip can be estimated. Note that for some geometries the estimated strike and dip will differ considerably from the true strike and dip (Nedimović and West, 2003b). Especial-
ly the true strike direction can vary significantly for gently dipping reflections (García Juanatey et al., 2013).

4.2.2. Azimuthal binning analysis

If a stacking line is oblique to the structures we want to image then the dip of these structures will be less than the true dip on stacked and migrated sections. With the azimuthal binning analysis we rotate the stacking bins in order to stack midpoints located in the direction of the strike of the subsurface structures to be imaged. By rotating the bins the seismic stack will more accurately image the true dip and the reflections will stack more coherently. Figure 4.3 shows how CMP bins are rotated optimally for imaging a specific subsurface structure with strike oblique to the stacking line. However, this analysis still requires that midpoints are spread out about the stacking line. For a straight 2D line rotating the bins will not move midpoints to new CMP:s, since all midpoints are located on the stacking line. Rotation of the stacking bins is equivalent to rotating the stacking line. Setting up several different geometries and running all the processing from start is, however, time consuming and unnecessary. Instead each midpoint can be shifted to a new CMP along the initial stacking line using equation 4.2.

New CMP number = original CMP number + [ m.o. / tan(γ) ] / binsize (4.2)

Figure 4.3. Schematic block diagram showing the azimuthal binning analysis. In (a) the bins are perpendicular to the stacking line. In (b) the bins have been rotated perpendicular to the dip-direction of the subsurface structure.

This shift is calculated based on the perpendicular midpoint offset (m.o.) from the stacking line and the angle (γ) between the stacking line and the new bin direction illustrated in Figure 4.4. The bin size/bin increment along
the stacking line is used to translate the distance of the shift along the stacking line to a CMP number shift.

Azimuthal binning will work better for structures that are fairly close to perpendicular to the stacking line. If large rotations of the CMP bins are required then smearing of the data will make interpretations difficult.

![Figure 4.4. Parameters used for calculating the new CMP for a trace with midpoint offset (m.o.) from the stacking line and for an angle \( \gamma \) between the stacking line and the rotated CMP-bin.]

4.2.3. Near surface effects on seismic data

In hard rock environments in Scandinavia the weathering layer in the bedrock is usually insignificant. Instead there is often an unconsolidated layer of glacial or post-glacial sediments above the bedrock (Figure 3.3). The thickness and composition of this layer can change rapidly with velocities ranging between 250 m/s and 2500 m/s (Stümpel et al. 1984; Juhlin et al., 2002; Büker et al., 1998). This rapidly changing layer can cause time delays between traces over short distances and may also cause strong source generated noise, contaminating and interfering with the seismic signal from shallow reflections. The source generated noise consists of direct, refracted, guided and surface waves (Robertsson et al., 1996a,b; Roth et al., 1998). First arrivals can be removed with a top mute function and surface waves can often be removed by frequency filtering (Wu et al., 1995), but direct Sv-waves may have similar dominant frequencies as the reflections and can, therefore, instead be attenuated in, for example, the \( f-k \) domain (Juhlin, 1995). In order to preserve high frequencies when processing reflection seismic data from the crystalline environments, the application of refraction static corrections is necessary (Juhlin, 1995; Wu et al., 1995; Adam et al., 2000; Schmelzbach et al., 2007a). Refraction static corrections based on traveltimes from critically refracted seismic energy (first arrivals) are used to calculate the delays caused by the near surface low velocity layer (Lawton, 1989).
4.3. Interpretation of reflection seismic data

Seismic data interpretation involves correlating the seismic data with other geological and geophysical data sets. This correlation usually involves surface 2D geological or geophysical maps, 2D geophysical data sections or 1D borehole data. When correlating seismic data with other geophysical data an important aspect to consider is that the methods measure different subsurface physical parameters. These differences are the strengths of the combined interpretation, since the interpretation aims at translating the geophysical data into geological units. When comparing seismic data with geological data an important aspect is the difference in scale and how accurate the geological data are. Some surface geological data can be very detailed and have to be generalized to fit the seismic data, but in other cases the geological data are based on sparse field observations and perhaps low resolution airborne geophysical data. Important aspects of the interpretation of seismic reflection data are the coherency of events and continuity of reflections and also seismic attributes such as the amplitude, phase and the frequency. More than 50 distinct seismic attributes (Chopra and Marfurt, 2005) can be defined, including for example amplitude, phase, frequency and amplitude versus offset (AVO). Attributes can be used to visually enhance a seismic section by superimposing a seismic attribute such as the reflector strength or interval velocities, but they can also be used to make estimates on physical properties, which are useful for comparison with geophysical borehole data, that may provide a detailed estimate of rock properties.

The coherency, amplitude and continuity of reflections can be compared within a stacked section. Areas with low reflectivity can be related to granitoid rocks, anorthosites, and gneisses (Gibbs, 2013) or other geological units lacking internal structures. Coherent reflections may represent geological units with pronounced internal structure, such as sedimentary basins or deformed units with mylonitic zones or elongated mafic bodies (Hynes and Eaton, 1999). A localized strong continuous reflection or set of reflections may indicate some kind of structural boundary or lithological boundary (Ganchin et al., 1998) or a fracture zone (Juhlin and Stephens, 2006). Very high amplitudes are often referred to as bright spots. Bright spots could indicate fluid saturated zones or magma chambers (Brown et al., 1995) or a high density mineralization (Malehmir and Bellefleur, 2009). A seismic impedance contrast (i.e. a density and/or seismic velocity contrast) across a boundary is necessary to produce reflections. A geological boundary does not necessarily coincide with a significant impedance contrast. These boundaries can instead be interpreted based on changes in reflectivity patterns or by means of using other geophysical data.

An important aspect of seismic data interpretation is the resolution of the data. A gneissic rock having a pronounced internal structure, consisting of elongated minerals and mineral aggregations, may appear transparent be-
cause the internal structures are too small to produce reflections of the seismic waves. If, however, elongated mafic bodies or discrete mylonitic zones occur in the gneissic fabric, then these structures may be large enough to produce reflections of the seismic waves. The vertical resolution of seismic data only depends on the wavelength of the seismic waves (i.e. the ratio between seismic velocity and frequency). It determines how well we can resolve steps or gaps in continuous reflections and also if we can resolve multiple boundaries, such as the upper and lower boundary of a layer. The horizontal resolution (the Fresnel zone), however, depends both on the wavelength and of the depth to the reflector (Yilmaz, 2001). Deep targets have to be larger laterally. The resolution of deeper targets is also less (both vertically and horizontally) because the higher frequencies of the seismic data are being attenuated faster than the lower frequencies and, therefore, deeper targets are sampled with a lower frequency (i.e. longer wavelength). Seismic targets that are smaller than the Fresnel zone will act as a diffractor and diffractions are suppressed by conventional seismic data processing. Diffractions can, however, contain information about geological details and can, therefore, instead also be enhanced in specialized processing schemes (Schmelzbach et al., 2007b; Malehmir et al., 2009).

The geometry of structures is also important to consider. Steep structures can be more difficult to image. Specialized processing can resolve reflections also from steep boundaries such as the San Andreas Fault in California (Louie et al., 1988), but with conventional processing a unit with vertical structures may appear transparent on the seismic section. The geometry of the seismic acquisition line must also be considered. Line orientation compared to the strike of structures can affect the strength of reflectivity (Eaton and Hynes, 2000) and bends in the acquisition line can cause false transparent zones on the seismic section (Wu, 1996). Before interpreting a seismic data set, such non-geological features in the data must be considered and accounted for.

4.3.1. Seismic data modeling

It is often useful to analyze how reflections in source gathers correspond to the reflections in stacked data. This may help to better project dipping reflections to the surface and to see if the stacked data could be improved, especially in the near surface, where reflections may be visible on source records, but not stacked properly on a section. On crooked profiles this can be difficult since the offset distribution is irregular. Traveltime modeling can then be useful. I have used a constant velocity ray tracing code (see Ayarza et al., 2000 for details) to calculate traveltimes for different reflector geometries. If stacked sections are in 2D and the modeling uses the 3D configuration of sources and receivers, there will be some room for error in fitting calculated traveltimes with the stacked data. The data can, however, also be compared
with source-gathers, and can, therefore, be more accurately fitted to the real data. The modeling assumes a constant bedrock velocity, which could be more or less consistent with a crystalline environment unless any lithological boundaries occur across the ray path, and the reflector is assumed to be a thin layer or simple interface in a homogeneous medium. The modeling also assumes a planar reflector surface, this is not always expected in the crystalline environment, but it still allows for a reasonable fit of modeled data with real data in many cases. Densities and seismic velocities are assigned to the reflector and to the surrounding rock. Appropriate reflection coefficients based on published equations (Aki and Richards, 1980) are calculated for both P- and S-wave reflections. This allows the comparison of reflections from a fracture zone model, with lower density and seismic velocity than the surrounding rock, and a mafic reflection model, with higher density and seismic velocity than the surrounding rock. Figure 4.5 shows an example where the absolute value of the reflection coefficients from fracture zone models and mafic models based on specific reflector geometry are compared (from Paper II). The reflection coefficients depend on the reflector model that is used, but also, largely, on the angle of incidence. Especially the P-S phase is highly dependent on the angle of incidence. This modeling was used to correlate a P-S reflection (visible in source-gathers) and a P-P reflection in a stacked section to a fracture zone model (in Paper II).

![Figure 4.5](image)

**Figure 4.5.** Comparison of the magnitude of reflection coefficients (|Rc|) for different reflecting boundaries. Geometry of boundary is almost perpendicular to receiver line and dipping 55°. (A) - Mafic boundary in gneiss. (B) – Fault zone boundary in gneiss (Vp/Vs ratio 1.6) and (C) – Fault zone boundary in gneiss (Vp/Vs ratio 1.8).
5. Paper summary

5.1. Paper I: High resolution reflection seismic imaging of the Ullared Deformation Zone, southern Sweden

5.1.1. Summary

The Ullared Deformation Zone (UDZ) is one of a few structures worldwide known to contain decompressed eclogite facies rocks of Precambrian age. Möller et al. (1997) outlined the UDZ as a shear zone or a shear zone system based on aeromagnetic data interpretation. The principal objective of the reflection seismic survey was to provide geometrical information of the deformation zone at depth. Due to the logistics of the Vibsit 1000 source (hydraulic hammer mounted on an excavator) the seismic data were acquired along existing roads. The recording line was, therefore, crooked and slightly oblique to the UDZ. Figure 5.1 shows the UDZ on an aeromagnetic anomaly map and the crooked acquisition line slightly oblique to the magnetic pattern of the UDZ. This oblique and crooked-line recording geometry reduced the stacking quality of the seismic data. Two methods were tested in order to increase the stacking quality and estimate the 3D geometry of some observed reflections. Cross-dip analysis described by Wu et al. (1995) and here in section 4.2.1 and an azimuthal binning procedure described in section 4.2.2 were applied to the seismic data. The final stacked and migrated section of the UDZ is shown in Figure 5.2. Most reflections in the stack are dipping towards the north on the N-S directed seismic profile.
Figure 5.1. Aeromagnetic anomaly map over the UDZ. White lines indicate the outline of the UDZ; approximate intersection of reflections D, E and F with CMP – line are marked with white stars. Projections to the surface, in dip direction, of reflections D-K (except reflection J) are marked with light grey bars oriented in the calculated strike direction. Boundaries between interpreted units are marked with white lines. Enlarged inset in upper right corner shows how northwest striking structures to the east appear to be cross-cut by north-northwest striking structures to the west. Crooked seismic acquisition line and the straight stacking line are shown with CMP numbers labeled for reference. Magnetic data courtesy of the Geological Survey of Sweden.
Figure 5.2. Final migrated stacked section with variable cross-dip corrections along the profile. 20° cross-dip towards east between CMP 100-820 and 1026-1150. 45° cross-dip towards east between CMP 821-1025 and 0° (no cross-dip) between CMP 1151-1380. Key reflections D-K are labeled with the calculated strike and dip (RHR) and the interpreted units are colored. Stacked fold along the profile on top. V ≈ H.

The cross-dip analysis showed that most reflections also had a cross-dip component dipping towards the east. Figure 5.3 shows an example of a cross-dip analysis of a reflective package between reflections H and I in Figure 5.2. The reflectivity is more coherent and continuous for cross-dips between 15° and 25° towards the east. Using the inline dip and the cross-dip, the true dip and strike of reflections could be estimated.

Figure 5.3. Parts of unmigrated stacks comparing the effect of different cross-dip corrections for reflections in the B unit (see Figure 5.2).
The azimuthal binning procedure also allowed the strike of reflections to be estimated. By rotating the CMP bins before stacking and comparing the stack visually, the optimum stack direction was found. This stack direction coincides with the strike of the reflectors imaged in the stacked section. Figure 5.4 shows an example of azimuthal binning analysis for reflection F. The results from the two methods were compared and showed good agreement. The cross-dip analysis provided the most visually clear results and was applied to the final stacked section.

![Figure 5.4. Parts of unmigrated stacks comparing the effect of different binning azimuths for reflection F.](image)

The geometry of several key reflections or groups of reflections, estimated using the results from the cross-dip analysis and the azimuthal binning procedure, are marked on the final stacked section (Figure 5.2). These key reflections represent the transition between one set of reflections and another and are, therefore, not necessarily the strongest reflections in the stack. Hynes and Eaton (1999) noted that domain boundaries are rarely associated with reflections stronger than the internal reflections of the domains in high pressure metamorphosed rocks. Characterization of different units must, therefore, be based on reflectivity patterns and geometrical observations. The strength of reflections, within the shear zone, can depend on, for example, the size and geometry of minor mylonite zones and mafic lenses or on the composition of the mafic lenses, for example the presence of eclogites. Reflections from the UDZ could be grouped into four units based on their differences in reflectivity and geometry. Some of the key reflections could be projected to surface and marked on the aeromagnetic map. Four units correlated directly with magnetic domains seen on the aeromagnetic map. Units A1 and B show a cross-cutting relationship with oppositely dipping reflectivity in the stack (Figure 5.2) and kinks on the magnetic lineations (Figure 5.1). This indicates that the units A1 and B have a different deformational history. Eclogites have only been found in the northernmost unit. The unit containing eclogites is displayed in a 3D view with the magnetic map in Figure 5.5. The seismic data and the detailed aeromagnetic analysis support the interpretation of the UDZ as a shear zone system. At least two or possi-
bly three stages of deformations in the UDZ were suggested. At an early stage A1 and A2 were formed. At a later stage deformation in unit B resulted in reorientation and/or cross-cutting of the earlier structures. A possible intermediate stage would be a separate evolution of units A1 and A2.

Figure 5.5. Perspective view of the final migrated stacked section and the magnetic anomaly map with interpreted boundaries of unit A1. G – Gällared.

5.1.2. Conclusions

The seismic data over the UDZ show how 3D information can be extracted from a relatively small 2D crooked-line data set using easily implemented processing tools. The seismic data support the interpretation of the UDZ as a shear zone system and the UDZ could be separated into four different units. Structures of the northernmost eclogite bearing unit dip approximately 20-30° towards the northeast. At deeper levels, these structures appear to become more sub-horizontal. This change in apparent dip, however, appears to be mostly correlated with a change in strike of the structures with depth as confirmed by the cross-dip analysis. Seismic surveying over complex structures, such as high grade gneiss terranes, may benefit from a crooked-line recording geometry if a 3D survey is not possible. Including some off-line source points close to straight segments and/or gaps in the profile could additionally increase the mid-point scatter and add further possibilities for interpretation.
5.2. Paper II: High resolution reflection seismic profiling over the Tjellefonna fault in the Møre-Trøndelag Fault Complex, Norway

5.2.1. Summary

The objective of this study was to reveal the deeper structure of one of the fault segments of the Møre-Trøndelag Fault Complex (MTFC) in Norway, namely the Tjellefonna fault. The MTFC is one of the most prominent fault zones of Norway both onshore and offshore, but in spite of this little is known about the deeper structure of the individual fault segments. Two profiles, approximately 7 and 9 km long, were recorded on each side of the Tingvollfjord (Figure 5.6).

The profiles followed crooked lines and rugged terrain and, therefore, careful processing was required. Soft ground conditions and the proximity to buildings and farms, housing animals, caused some gaps in the source point locations along the profiles. These gaps mostly affected the refraction static calculations and, hence, lowered the stacking quality, especially in the upper 0.2 s, in some sections of the profiles. Since the profiles were acquired with a crooked-line recording geometry, an azimuthal binning procedure and a cross-dip analysis, described in Paper I, were applied to the data in order to...
investigate possible out-of-the-plane reflections. The tests, however, did not improve the stacking quality compared with the standard stacking procedure, indicating that the imaged structures strike nearly perpendicular to the stacking line. Linking reflections from shot records to reflections in the stacked sections was difficult due to the crooked line recording geometry causing irregular offsets and also due to the weak coherency of reflections in the upper 0.2 s in the stacked sections. Therefore, reflection traveltime modeling was performed using a constant velocity ray tracing code, see Ayarza et al. (2000) for details. Traveltimes for different reflector geometries were calculated and compared with reflections seen in source gathers and in the stacked section. Since the stacked sections are in 2D and the modeling uses the 3D configuration of sources and receivers, there is some room for error. The data are, however, also compared with source-gathers and hence the fitting of modeled traveltimes will be more sensitive. On the northeastern seismic profile, S.P. 1, shown in Figure 5.7, a disrupted reflective package forming a gentle antiform was imaged at about 0.5 km depth. A similar antiform is imaged in the southwestern seismic profile, S.P. 2, shown in Figure 5.8, but with the strongest reflectivity at about 1 km depth.

Figure 5.7. Final migrated stack of S.P. 1 with marked position of reflection S2. Arrows mark sections affected by sharp bends in the recording line (compare with Figure 5.6). Elevation displayed on top. V ≈ H.
Figure 5.8. Final migrated stack of S.P. 2 merged with migrated stack using high stacking velocities. Green zone – LVA; Blue zone – LVB refers to zones with low near surface bedrock velocity observed on first refracted arrivals in source gathers. Reflections S1 and R1-R5 marked. Location of axial plane of antiform from geology map (Figure 5.6a). Elevation displayed on top. $V \approx H$.

This antiform is only seen on the northwestern part of S.P. 2, while the southeastern part shows weak reflectivity. Some reflections with steeper dips are imaged on S.P. 2. Some of these reflections could be modeled and correlated with reflections seen in source-gathers. One of these reflections can be correlated to a P-S converted reflection in source gathers and is coming to the surface in the topographic low, suggesting that this reflection is related to a fault. Modeling based on P- and S-wave velocities and densities from samples collected along S.P. 2 and using a reflector geometry that fits the P-P reflections observed in the stacked section was performed. Modeling indicated that the P-S converted wave seen in source-gathers is correlated with a reflection off a fault zone boundary. The boundary location also correlates well with a low resistivity anomaly from a resistivity profile acquired along the center of S.P. 2. Figure 5.9 shows the resistivity profiles and the reflection seismic data in 3D view with geology and the interpreted fault.
Figure 5.9. Seismic profiles 1 and 2 and resistivity profiles 4, 5 and 7 plotted with geology in 3D perspective view. Blue dashed lines indicate the topographic lineament and red solid lines mark reflections S2 and S1/R1-R5. See Figure 5.6 for locations. The blue plane indicates the modeled fault plane with strike 55° and dipping 55° towards southeast. The plane extends to 400 m in this figure. Antiform structures are enhanced with grey squares. The fold hinge line seems to be subparallel to the fault plane.

5.2.2. Conclusions

The Tjellefonna fault was imaged using two reflection seismic profiles. The fault dips 50°-60° towards the southeast and extends to a depth of at least 400 m and most likely at least to a depth of 1.3 km. The fault is imaged in the stacked section on the southwestern profile, but an in-strike P-S converted reflection on the northeastern profile suggests a continuation of the fault on that side. Correlating the reflections with the resistivity profile acquired over the central parts of the southwestern seismic profile suggests that the fault zone diverges into two separate zones near the surface. An antiform can be seen on both profiles. The fold hinge line of the antiform is parallel to the suggested Tjellefonna fault, indicating that the folding and faulting may have a causal relationship.
5.3. Paper III: Reflection seismic investigations in the Dannemora area, central Sweden: Insights into the geometry of polyphase deformation zones and magnetite-skarn deposits

5.3.1. Summary

Two nearly perpendicular reflection seismic profiles were acquired in the Dannemora area. Profile 1 connects the Dannemora area with the Forsmark area and profile 2 crosses the active Dannemora mine. Figure 5.10 shows the location of the seismic profiles on geology and aeromagnetic maps. The objective of this study was to increase the knowledge of the crustal structures in the upper 3 km of the crust in order to improve the geological understanding of the Bergslagen region in general and specifically that near the Dannemora mineralization. Several shorter seismic profiles had been acquired in the Forsmark area that focused on the planned Swedish site for storage of spent nuclear fuel (Juhlin and Stephens, 2006). Structures at a greater distance from the Forsmark area are, however, poorly understood. In this summary I will focus on the part of Paper III to which I contributed i.e. the north-south profile (profile 1) that connects the Dannemora area with the Forsmark area. Structures in the Forsmark area are dominated by the high strain shear zones; the Eckafjärden Deformation Zone (EDZ), the Forsmark Deformation Zone (FDZ) and the Singö Deformation Zone (SDZ). These zones are characterized by a strong linear pattern on the magnetic maps and the linear structures continue further south to the central part of profile 1 (Figure 5.10). The southern part of profile 1 appears to lack structures both on the magnetic and geological maps, however, a large positive gravity anomaly and a magnetic high anomaly are centered on the southern end of profile 1. The gravity anomaly seems to be cut off by the SVFZ or ÖDZ. Figure 5.11 shows unmigrated stacked (a) and migrated stacked (b) seismic sections of profile 1. Profile 1 clearly shows a package of reflectivity (I1) in the southern end. These reflections, and their correlation with the large gravity anomaly, suggest these reflections are from the top of a large mafic intrusion.
Figure 5.10. (a) Geological map of the Dannemora area. Black line shows the location of seismic reflection profiles 1 and 2 and those acquired at the Forsmark site (Juhlin and Stephens, 2006). CMP lines are shown using blue color. Dashed line shows the inferred location of the Österbybruk Deformation Zone (ÖDZ) near the Dannemora mine. (b) Total field aeromagnetic map of the Dannemora area (Stephens et al., 2009). Red lines, show the location of seismic reflection profiles 1 and 2 and those acquired at the Forsmark site (Juhlin and Stephens, 2006) are shown in black. Dashed blue lines show the Singö Deformation Zone (SDZ), Eckafjärden Deformation Zone (EDZ) and the Forsmark Deformation Zone (FDZ). Dashed turquoise line shows a possible ductile shear zone forming a boundary between two structural domains. Reflections observed from profiles 1 and 2 are projected to the surface, and annotated. Maps courtesy of the Geological Survey of Sweden.
Figure 5.11. (a) Unmigrated stacked section along profile 1. (b) Migrated stacked section along profile 1. A series of north dipping reflections (I1, I2, F1, F2, F3) and a reflective zone P1 are shown in the southwestern and central parts of the section. A south dipping reflection (P2) is also observed at the northernmost part of the section. P1 approximately correlates with the structural domain boundary outlined in turquoise in Figure 5.10b.

Above the I1 reflectivity, there is a strong reflection, I2. It is not clear what causes the I2 reflection since both the magnetic and geological maps lack any correlating features to where I2 projects to the surface. Reflections F1-F3 may represent a series of fault zones within the metagranodiorite or at the contact with the metavolcanic rocks. Especially reflections F2 and F3 show good correlation with the metavolcanic rocks. The volcanic rocks appear to be correlated with deformation zones in the area (the SVFZ and ÖDZ in Figure 5.10a). P1 and P2 are two reflective zones. P1 seems to be dominated by northeast dipping structures and may be associated with the strong lineament visible in the magnetic map (turquoise dashed line in Figure 5.10b). This lineament is subparallel to the linear magnetic pattern in the Forsmark area. This zone has a cross-cutting relationship with the SVFZ and ÖDZ. Reflective structures southwest of P1 are clear and have a dip direction towards the northeast. The reflective pattern northeast of P1 appears to be more complex with contrasting dip directions. This boundary coincides with a structural boundary between two domains. The northern domain consists of steeply dipping high ductile strain belts (Stephens et al., 2009). P2 is a reflective package poorly represented in the seismic data. One explanation for the poor coherency of these reflections is that there is a significant cross-dip
component present. The magnetic pattern is oblique to the acquisition line in the northern end of profile 1. Cross-dip analysis was conducted, but without any positive result, probably due to a too small aperture of the midpoint spread. The P2 reflectivity dips towards the southeast/southwest and projects to the surface north of profile 1, where a strong linear magnetic pattern, sub-parallel to the Forsmark Deformation Zone (FDZ), is present. The P2 reflectivity could, therefore, be associated with the deformation structures of the (FDZ). The FDZ, however, is interpreted as a subvertical deformation zone and as such is difficult to image with the reflection seismic method (Sharifi Brojerdi et al., 2013). The P2 reflections may instead be related to south or south east dipping fracture zones similar to those observed by Juhlin and Stephens (2006), in the Forsmark area. The structures along profile 1 are characterized by northwest dipping reflections except for at the northern end where structures have a southerly dip direction (P2 reflections). This is in contrast to the trend of reflections in the Forsmark area, where reflections mostly show south or south-westerly dips (Juhlin and Stephens, 2006; Sharifi Brojerdi et al., 2013). Sharifi Brojerdi et al. (2013) interpreted these structures and a south dipping reflective zone as possibly a part of a positive flower structure. The structures imaged on the northern part of profile 1 and south of the EDZ and FDZ could in that case form the other part of this structure. A tectonic sole fault of the suggested flower structure is, however, not imaged on profile 1.

![Figure 5.12. 3D volume obtained from swath 3D processing of part of profile 2 demonstrating the main strike, which is N25° E, of reflection R1, related to the Dannemora iron ore at shallow depth.](image)
Profile 2 imaged the ÖDZ as a southeast dipping structure, using cross-dip analysis. Microstructure analyses indicate the last ductile movement along ÖDZ as a reverse or with an east-side-up component. Other prominent reflections imaged on profile 2 were related to structures correlated with the Dannemora mineralization. Swath 3D processing of profile 2, shown in Figure 5.12, using additional cross-profile source points imaged the structures, in the vicinity of the mine in 3D and indicated these structures to strike N25°E and extend to a depth of 2.2 km, which is 1.2 km deeper than the known depth of the ore body.

5.3.2. Conclusions
Two nearly perpendicular reflection seismic profiles were acquired across the Dannemora mine and between the Dannemora area and the Forsmark area. The east-west profile imaged structures correlated with the Dannemora mineralization, by swath 3D imaging, using additional cross-profile source points. The ÖDZ was also imaged with 3D structural information revealed using cross-dip analysis. The north-south profile imaged several mostly north or northeast dipping structures that may be related to fault zones or fracture zones. Cross-profile apertures were too small for cross-dip analysis to be useful, but magnetic lineaments oblique to the seismic profile and the weak coherency of some reflections suggest that a cross-dip component may exist. Reflectivity in the southwestern part is clear and northeast dipping while reflectivity in the northeastern part is complex with contrasting dip directions. The two different reflective areas correlate with a structural boundary between two tectonic domains. The northern domain consists of steeply dipping high ductile strain belts. The structures in the northern Dannemora area show oppositely dipping structures as compared to the structures in the Forsmark area further north.

5.4. Paper IV: High-resolution 3D reflection seismic investigation over a quick-clay landslide scar in southwest Sweden
5.4.1. Summary
Landslides are one of the most common natural disasters worldwide, killing more than hundreds of people every year and damaging buildings and infrastructure worth billions of dollars (Petley, 2011). One type of landslide, which often occurs in the northern hemisphere in areas that were covered by Pleistocene glaciations, is quick-clay landslides. These landslides are particularly common along the shorelines of the Göta River in southwestern Swe-
den. The high resolution 3D reflection seismic investigation was part of an intensive characterization project that includes ground gravity, magnetics, controlled source radio-magnetotellurics, resistivity and reflection seismsics using both P- and S-wave sources and receivers (see Malehmir et al., 2013a for details). This was the first time reflection seismsics was used in a landslide site investigation in Sweden. Reflection seismsics was thought to be better at imaging, and especially delineating lateral discontinuities, in thin layers with low electrical conductivity embedded in highly conductive clay, compared with, for example electrical methods, that are commonly used in landslide site investigations. The 3D area is located above a c. 40 year old landslide scar. The fan shaped geometry of the scar indicates that the landslide was retrogressive, i.e. it started near the river and then it progressed towards the south. Figure 5.13 shows the location of the 3D area and receiver lines as well as seismic lines 4 and 5. The actual triggering mechanism is not known, although erosion from the river is assumed to have been important. The principal objective of the 3D reflection seismic data was to image the bedrock topography and layering in the normally consolidated sediments above. The sediments consist mainly of clays with thin layers of sandy-silty materials. A few boreholes drilled by the Swedish Geotechnical Institute (SGI) reached a sandy-silty layer at about 20 m below the surface (Löfroth et al., 2011). This layer was the main target for the 3D reflection seismic investigation, since quick clays commonly occur above this layer. None of the boreholes, however, reached below this layer. One of the challenges in this study was to successfully image this shallow target layer. The uppermost two to five meters of the subsurface consisted of a low velocity layer (250-350 m/s). The seismic data quality improved significantly when variations in this layer were removed by the application of refraction static corrections. Another important processing step was spectral weighting that was used in order to boost the amplitudes of higher frequencies. The higher frequency signals increase the vertical resolution of the data, and hence improve the chances of detecting small steps across seismic horizons. Figure 5.14 shows selected inlines from the 3D seismic cube. Three main horizons could be delineated in the seismic data. The deepest horizon (B1) could be correlated with the bedrock topography.
Figure 5.13. Locations of the 3D seismic area and receiver lines as well as seismic lines 4 and 5 (see Malehmir et al., 2013b) overlaid on the air photo from the study area and the Lidar (light detection and ranging) elevation map. Landslide scar is clearly seen in the northeastern corner of the 3D lines. R1 to R3 relate to reflections projected to the surface. F1 shows the location of a possible subsurface fault or step in the bedrock topography. Inset map of the landslide risk in Sweden. More reddish colors indicate higher risk for landslides. Lidar data and air photo are provided by Lantmäteriet.

The bedrock dips about 25° towards the north and extends nearly to the surface in the south close to where bedrock is exposed on the surface. Towards the north, the B1 horizon is interrupted. This is likely due to a step in the bedrock topography, as indicated on some inlines, or due to an increased dip of the bedrock topography. Forward modeling of gravity data collected along a short north-south profile across the center of the 3D area suggests a step of about 20 m of the bedrock interface. A bedrock interface 20 m deeper towards the north does not explain why the bedrock cannot be imaged. However, since the study area is located in a deformation zone it is likely that the step or increased dip of the bedrock topography is indicative of a fault and marks the boundary to perhaps severely faulted and/or fractured rock to the north. In Figure 5.15 the picked seismic horizons are displayed as depth maps. The uppermost horizon (S1) could be imaged across the entire 3D area. The S1 horizon is, however, interrupted across the center of the 3D area. This interruption correlates well with the suggested fault in the be-
The same interruption is also evident in the deeper seismic horizon (S2). The correlation between the interruptions in both imaged layers in the sediments with the bedrock topography suggests that the interruption correlates to a preexisting feature not caused by the movement of the landslide itself.

Figure 5.14. Selected inlines from the 3D migrated seismic cube. Depth to the coarse grained layer from the geotechnical boreholes (Löfroth et al., 2011) is indicated by blue vertical lines. Interpreted seismic horizons are marked in yellow (S1), orange (S2) and red (B1). Arrows indicate where some reflectivity is observed above the S2 horizon in the western part of the area. Steps in the bedrock topography, near the truncation of the B1 horizon, are observed in inlines 108, 112, 114, and 116.
Figure 5.15. (a) Elevation of the B1 horizon (the bedrock interface). B1 is only imaged in the southern part of the 3D area. The truncation of the horizon is marked with F1 and the location is compared with the S1 and S2 horizons in (b) and (c). (b) Elevation of the S2 horizon. (c) Elevation of the S1 horizon. Black dots indicate where seismic horizons could be picked. Disturbances in the seismic horizons S1 and S2 (outlined in purple) correlate with F1 and with the southern extension of the landslide. Landslide scar marked with black line for reference. Black arrow indicates the transition between flat and west dipping reflectivity in the S1 horizon. This transition correlates with the western scar boundary. Purple dots indicate locations of possible high pore-water pressure build up (see text for explanation).

The interruption in the uppermost S1 horizon also correlates well with the extension of the landslide scar, indicating that this feature may have played a role in triggering the landslide and or limiting the extension of the landslide. The S1 horizon follows the bedrock topography close to the surface towards the south, suggesting the possibility for increased fresh water infiltration in this direction. In Figure 5.16 a schematic interpretation is shown. At the interruption in the S1 horizon it is likely that the interaction between fresh water and the clays will be more effective and, hence, increasing the amount of quick clays. Since the clays also have much lower permeability a higher pore water pressure may build up at this location. Erosion from the river in combination with an increased pore water pressure may also have triggered the landslide. A higher amount of quick clays above the disturbed zone in the S1 horizon could have acted as a weak zone, where a break of the landslide retrogression would more easily occur.
Figure 5.16. Schematic interpretation. The bedrock interface is dipping north and is interrupted at F1, due to a step or increased dip that may be related to a fault within the Götaälv Zone. Reflections R2 and R3 may be related to the sediment/bedrock interface on the northern shore of the Göta River. Marine clays above the bedrock are interrupted by two horizons (imaged by the 3D seismic data). The uppermost horizon is a coarse-grained layer (confirmed by geotechnical drillings, Löfroth et al., 2011). Rain water could infiltrate the coarse-grained layer from the south where the layer reaches close to the surface, but also water from the Göta River could infiltrate the coarse-grained layer from the north. The fresh water can leach the marine clays situated above the coarse-grained layer, forming quick clays. The southern end of the landslide scar correlates with an interruption in the seismic reflections from the coarse-grained layer. Clays cutting off the coarse-grained layer at this location could cause a concentration of water outflow from the coarse-grained layer into the clays, which could cause an increased pore-water pressure above. The clays cutting the coarse-grained layer may also have been more affected by fresh water and perhaps more quick clays formed above this location. A high amount of quick clays at this zone may have formed a weakness in the unit that could limit the extension of a retrogressive landslide towards the south. Sketch is not drawn to scale.

5.4.2. Conclusions

This study shows the potential of using 3D reflection seismics as a complement prior to/after drilling and other geophysical methods when performing landslide site investigations. The good resolution of the shallow target layer, located about 20 m below the surface, is attributed to good surface conditions during the acquisition and also to the use of refraction static corrections in order to remove the effects of the uppermost low velocity layer. The detailed 3D seismic image of the permeable layer revealed a discontinuity in this layer. This discontinuity could be correlated with a similar discontinuity in a deeper layer and also to a possible fault in the bedrock. The discontinuity also correlates with the extension of the landslide, suggesting that it may have played a role in triggering or in limiting the extension of the landslide.
6. Conclusions

The presented papers show different applications of seismic methods to study deformation zones in crystalline hard rocks, and how more information can be extracted from relatively small crooked 2D seismic lines, with the aid of easily implemented processing tools. Different non-conventional seismic data processing techniques were used. The azimuthal binning procedure and cross-dip analysis allowed 3D information to be extracted from the UDZ data set and significantly enhanced the interpretational possibilities in Paper I. New structural information about the Ullared Deformation Zone at depth was obtained. In Paper II integration of resistivity profiles and reflection seismic data as well as seismic travel time modeling made interpretation of the Tjellefonna fault zone, in the Møre-Trøndelag Fault Complex in central Norway, possible and in Paper III, new structural information was obtained in the important mining district of Bergslagen in central Sweden. Paper IV showed the possibility to use the 3D reflection seismic method for landslide studies. In favorable surface conditions and with careful seismic data processing a seismic horizon about 20 m below the surface was imaged. The 3D seismic data allowed for reliable interpretations of structures in the normally consolidated sediments as well as for the bedrock topography down to a depth of approximately 80 m below the surface. The reflection seismic method should perhaps be considered more often, as a complement to electrical methods, in very near surface investigations, especially when the subsurface consists of clays with high electrical conductivity that may mask thin layers with low electrical conductivity.

Some new data have been acquired since the publications or submission of Papers I-IV. MT (magnetotelluric) data were recorded across a profile perpendicular to the strike of the UDZ (Paper I), but slightly offset and oblique to the reflection seismic profile acquired in 2007. The MT data clearly show a boundary between significantly different resistive units correlating with the boundary between units A1 and A2 interpreted from the reflection seismic data or possibly between the A1 and C units (M. Bastani, personal communication, 2013). The geometry of the boundary is similar in both MT and reflection seismic data. The B unit appears to be correlated with a unit with similar resistivity as the A1 unit.

Within the quick-clay landslide project at Fråstad (Paper IV) more seismic lines were acquired in 2013, especially a 1.7 km long profile across the Göta River coinciding with the central recording line in the 3D area. Also
three new boreholes were drilled for downhole physical property measurements and geophysical investigation. These boreholes were located outside the 3D area, but the drilling confirmed that both layers observed in the seismic data in the normally consolidated sediments above the bedrock consisted of sand/silt/gravel layers. Quick clays were sometimes observed above these layers. The uppermost layer was thin, about 5 m thick, but in the western part the layer was about 10 m thick (A. Malehmir, personal communication, 2013) and could, therefore, possibly be thick enough to produce reflections from both the upper and the lower boundary.

At last I will give some thoughts from my experience up to now in seismic processing in hard rock environments. Spending time carefully on basic processing steps (i.e. geometry setup, trace editing and muting and first arrival picking) will pay off later. The refraction statics are very important for the end result both in hard rock, but also in areas with near surface unconsolidated sediments (Paper IV). When acquiring seismic data, gaps due to poor subsurface conditions or fragile constructions etc., may occur. The reduced fold in the data may be acceptable, but the limitation can instead come from missing first arrivals and a poor refraction static calculation. Such gaps should, if possible, be avoided perhaps using a less powerful source in some sections or by acquiring a few shots on a side road or parallel road. Acquisition of shots on a side road or parallel road will also be useful for increasing the midpoint spread of crooked line data, enabling better 3D coverage.
7. Summary in Swedish


Artikel fyra (Paper IV) behandlar 3D reflektionsseismiskt data från ett område i södra Sverige som berörts av ett kvicklara skred. Detta område ligger inom en förkastningszon och de strukturer som återfinns i de lösa avlagringarna ovanpå berggrunden kan ha påverkats av förkastningar i berggrunden. Den stora utmaningen med de reflektionsseismiska data som presenteras i artikel fyra är att den mest intressanta strukturen ligger på ett djup av cirka 20 meter. Bra ytförutsättningar vid inhämtandet av data samt nog-
grann processering är en förutsättning för att en bra bild från så ytnära struk-
turer skall kunna fås.

Sammantaget har forskningen som presenteras här bidragit till en ökad
förståelse för de strukturer som återfinns på djupet i flera intressanta områ-
den som tidigare var outforskade. Denna forskning kan således användas av
geologer och geofysiker som arbetar i dessa områden för att sätta samman de
upptäckter som görs vid senare geologiska karteringar samt vid ytterligare
geofysiska kartläggningar till en geologiskt hållbar modell. De processe-
ringsmetoder som använts kan rekommenderas för framtida forskare som
arbetar med reflektionsseismiskt data från krokiga profiler i kristallina berg-
områden. Den processeringsteknik som använts och de resultat som uppnåtts
för de ytnära strukturerna vid kvicklara området i södra Sverige visar att
reflektionsseismik bör användas vid liknande tillfällen i större utsträckning i
framtiden. Reflektionsseismik kan framgångsrikt komplettera andra geofy-
siska mätningar vid kartläggning av exempelvis skredområden.
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During my almost seven years at Uppsala University in the Reflection Seismic Group many people have contributed to my studies and to my well being. I have found that the same principle that comes from physical training also applies to PhD studies. During training you break down your body (brain) and when we rest we can grow stronger. Therefore, coffee breaks are an essential ingredient in a progressive training plan. This is where we can recharge and come back with new ideas and new solutions to problems we encountered. The time we spend at home with family and friends is also of highest priority. Many times the solution to a problem has been found on my way home, while riding my bicycle through the forest. I have been very unsuccessful at taking the work home, which is probably good, since you should try to avoid over training. At times the pressure at work has been tough, with approaching deadlines and multiple tasks that need immediate attention. I thank my family for taking me through those times. But the training is of course also important. If you don't do the proper exercises you will not get stronger during your rest. The first trainer I would like to acknowledge is Håkan Sjöström. He played an important role in directing me towards bedrock geology and eventually also towards shear zones. He supervised my Masters thesis on kinematic analysis of ductile and brittle/ductile deformation zones in the Oskarshamn area. Eventually I took, what appeared to be, a large jump to reflection seismic studies. In the beginning it seemed like Chris Juhlin was a bit of a miracle worker, able to find reflections were you could only see noise. With the sometimes direct guidance and sometimes by subtle hints he eventually taught me some of his tricks. It is not, however, only geophysics I have learned from Chris. Time management and the process of writing, as well as my English, are other things I am happy to have improved over these years with a lot of credit to Chris for that. My co-supervisor Alireza Malehmir has also provided great support and mentoring, especially during the last years, when we have collaborated closely in the Dannemora and the Fråstad landslide projects. Both Chris and Alireza understand the importance of recreational times. I thank you both for the opportunity to see a bit more of the world, especially the nice city of Vienna, the casinos in Las Vegas and the Great Barrier Reef (Australia). I would also like to thank my earlier co-supervisor Laust Pedersen and our chief of fieldwork Hasse Palm. The diversity of our geophysics group has been very positive for my well being. The topics on the coffee breaks and the field work
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A doctoral dissertation from the Faculty of Science and Technology, Uppsala University, is usually a summary of a number of papers. A few copies of the complete dissertation are kept at major Swedish research libraries, while the summary alone is distributed internationally through the series Digital Comprehensive Summaries of Uppsala Dissertations from the Faculty of Science and Technology.