Inversion of shallow amplitude anomalies in multi-channel seismic reflection data from the Reinga Basin

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Abstract

The Reinga Basin is a relatively unexplored sedimentary basin located to the northwest of Northland; it is regarded as the northern continuation of the Northland Basin. The Reinga Basin is bounded to the west by the compressional West Norfolk and Wanganella Ridges. To the northeast, the basin is bounded by Reinga Ridge and the Vening-Meinesz Fracture Zone. In 2008-2009, New Zealand Petroleum and Minerals, jointly with CGG Veritas, acquired two 2-D multi-channel petroleum speculation seismic surveys, Stratus-2D and Reinga-09, within the Reinga Basin. To date, no bottom-simulating reflections (BSRs) indicating the presence of gas hydrates have been conclusively identified in either survey. However, several areas of anomalously high amplitude and unresolved polarity have been identified within the probable gas hydrate stability zone (GHSZ). Prestack inversion on prestack Kirchhoff time-migrated CMP gathers using the Hampson-Russell Suite has been undertaken on these anomalous amplitude areas to determine the nature of these strong reflections. Due to the lack of well control in the Reinga Basin, background velocity models required for the prestack inversion were calculated using several methods; (1) travel-time inversion, (2) stacking velocities, and (3) stacking velocity pseudo-wells. Prestack inversion results from a portion of the STRATUS-2D seismic suggest that most areas of anomalous amplitude are associated with igneous intrusions. P-wave impedance inversions suggest that two of the anomalies may be caused by the presence of shallow (100-200 ms below seafloor), stratigraphically controlled free gas migrating updip along an unconformity.
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Chapter 1

Introduction

1.1 Overview

The Reinga Basin is a relatively unexplored sedimentary basin located northwest of New Zealand (Figure 1.1). Essentially a northward continuation of the Northland Basin, the Reinga Basin is bounded to the northeast by Reinga Ridge and the Vening-Meinesz Fracture Zone (a left stepping dextral transform fault), and to the west by Wanganella Ridge and West Norfolk Ridge (Herzer and Mascle, 1996; Herzer et al., 1997; Mortimer et al., 2007; Uruski, 2010). The Reinga Basin extends approximately 500 km northwest of Cape Reinga where it is closed off by the meeting of the West Norfolk, Wanganella and Reinga Ridges, resulting in a basin area of ~150 000 km$^2$ (Stagpoole et al., 2009). Water depths within the basin range from 500 to 2000 m, but reach as shallow as ~80 m above Wanganella/West Norfolk Ridge in the far northwest of the basin. Gravity modelling indicates that the crust underneath the Reinga Basin is ~15-20 km thick, compared to ~20-30 km for the surrounding West-Norfolk and Reinga Ridges, indicating crustal thinning under the Reinga Basin of approximately 35-40% (Herzer et al., 1997). The Northland Allochthon extends at least 100 km northwest into the basin from the Northland Peninsula (Stagpoole et al., 2009; Uruski, 2010). Two modern multi-channel seismic surveys have been collected in the Reinga Basin recently: the Reinga-09 survey in 2009 and the STRATUS-2D line in 2008. These surveys consist of ~5000 linear km of industry quality data which is tied to the Waka Nui-1 well in the Northland Basin (Stagpoole et al., 2009). Initial interpretation of the Reinga-09 seismic survey indicates that sediment thickness may reach up to 7 km (Stagpoole et al., 2009).

The 7 km thick Cretaceous-Cenozoic sediment package within the basin has numerous inferred potential sources rocks, and Eocene and Neogene compressional tectonic events, indicate that the Reinga Basin is one of the most prospective of New Zealand’s frontier petroleum basins (Stagpoole et al., 2009).
Figure 1.1 - Location and bathymetry map of the SW Pacific/Tasman Sea area and the Reinga Basin showing the location of the STRATUS-2D-Sequence 2 seismic line in black and red, major bathymetric features in the area, and dredge sample locations after Herzer et al. (1997) shown as symbols. The portion of the STRATUS line indicated in red was chosen for prestack inversion using the Hampson-Russell Suite software for P-wave impedance, S-wave impedance and density. The location of faults associated with the Vening-Meinesz Transform are indicated in grey after Bache et al. (2012) (Bathymetry provided by NIWA).
1.2 Purpose

The Reinga Basins contains a thick sedimentary sequence, and numerous amplitude anomalies have been observed on new modern seismic data sets within the basin, the Reinga-09 and STRATUS-2D surveys. Some of these amplitude anomalies are at appropriate depths below the seafloor to be within the potential gas hydrate stability zone (GHSZ) for the Reinga Basin.

The purpose of this study is to:

(1) Identify areas of anomalous seismic amplitude in the Reinga Basin based on the publicly available Reinga-09 and STRATUS-2D multichannel seismic surveys.

(2) Determine the nature of any anomalous amplitude areas identified using seismic interpretation and prestack seismic inversion methods; in particular to determine whether there is any methane hydrate, or indicators of potential methane hydrate in the Reinga Basin.

(3) Test the viability of industry standard inversion methods in areas with no/limited well control and outcrop using the Hampson-Russell Suite of inversion software.

1.3 Background geology

1.3.1 Basement geology

The number of physical samples recovered from the Reinga Basin is limited, with nearly all samples dredged from uplifted ridges: the West Norfolk, Reinga and Wanganella ridges, which form the margins of the basin (Herzer et al., 1997; Mortimer et al., 1998; Mortimer et al., 2010). The basement geology for the Reinga Basin is inferred from these sparse samples, geophysical evidence, in particular the location of the Junction Magnetic Anomaly, extrapolation from limited well data in the nearby Northland and Taranaki Basins, and outcrop and drill data in western Northland (Isaac, 1994; Herzer et al., 1997; Mortimer et al., 1997; Mortimer et al., 1998; Uruski and Stagpoole, 2004). The Reinga Basin is situated west of the Junction Magnetic Anomaly, which was inferred by Hatherton and Sibson (1970) to represent the Dun Mountain Ophiolite Belt - Maitai Terrane (Hatherton and Sibson, 1970); indicating that at least the southeastern portion of the Reinga Basin is likely to be underlain by a basement of Murihiku Supergroup rocks (Campbell et al., 2003) (Figure 1.2). The presence of the Murihiku Supergroup in western Northland and west of the North Island has been
confirmed by direct sampling in several petroleum exploration wells both onshore and offshore; Waka Nui-1 in the Northland Basin (Milne and Quick, 1999; Mortimer et al., 2009; Stagpoole, 2011), Waimamaku-1 and 2 in onshore western Northland (Hornibrook et al., 1976), Awhitu-1, a water bore in west Auckland, and numerous wells in the Taranaki Basin (Mortimer et al., 1997).

Dredged samples from West Norfolk and Wanganella Ridges range in composition from gabbro to biotite granite, with some samples showing a distinct foliation, indicating that they are probably best termed as orthogneiss (Mortimer et al., 1998). The strong linear positive magnetic anomalies associated with the West Norfolk and Wanganella Ridges and the geochemistry of the dredged samples indicate that they are probably correlatives of the Brook Street Terrane and the Median Tectonic Zone (MTZ) (Mortimer et al., 1998). The location of Brook Street Terrane and MTZ rocks to the west of Murihiku Supergroup rocks beneath the Reinga Basin is consistent with the current understanding of the New Zealand Basement Terranes, i.e., following the same arrangement as in the southern South Island (Mortimer, 2004).

Unlike the western and southern areas of the Reinga Basin, the basement rocks occurring beneath the northern portion are enigmatic, mainly due to the presence of the Vening-Meinesz transform, the offset of which is poorly understood, and the lack of samples of the basement units available from below Reinga Ridge.

1.3.2 Reinga Basin stratigraphy

Interpretation of the Reinga-09 seismic survey by Stagpoole et al. (2009) indicates that the Reinga Basin contains a Cretaceous-Cenozoic sedimentary sequence up to 7 km thick. Two
main efforts to interpret the sedimentary and tectonic history of the Reinga Basin using seismic stratigraphy have been published (Herzer et al., 1997; Bache et al., 2012).

Herzer et al. (1997) interpreted all of the available data in the Reinga Basin at the time of publishing, approximately 8900 km of predominantly single-channel seismic data, but also some low fold multichannel seismic data. The highest fold data analysed was 700 line km of a 96 channel survey collected by the Australian Geological Survey Organisation (Herzer et al., 1997). Recent work by Bache et al. (2012) revised the tectonic history of the Reinga basin using approximately 20 000 km of new modern high-fold industry seismic data from the Reinga (Reinga-09 and STRATUS-2D), Northland, Deepwater Taranaki and Taranaki Basins. The broad data coverage and seismic ties to the producing Taranaki Basin enabled regional reflections to be tied to wells and correlated throughout the basins (Bache et al., 2012). The Reinga-09/STRATUS-2D survey is directly tied to the Waka Nui-1 well in the Northland Basin, and broadly ties six other wells in the Taranaki Basin. The Waka Nui-1 well, drilled in 1999, lacks a sonic log or checkshot survey. However, several other geophysical logs were collected while drilling (LWD) including gamma ray and neutron density (Milne and Quick, 1999). As velocity information is required to make an accurate well tie to seismic data (so that well data can be confidently interpolated throughout a basin), Stagpoole (2011) developed a velocity model for the Waka Nui-1 well based on stacking velocities for the various seismic surveys which tie to the well, and confirmed its correlation to the seismic data using the LWD logs.

1.3.2.1 Stratigraphy

In this section, I provide a brief overview of the published stratigraphy of the Reinga Basin, predominantly from the most recent work by Bache et al. (2012). Detailed seismic facies descriptions, seismic stratigraphy and age interpretations for the Reinga Basin can be found in Herzer et al. (1997) and Bache et al. (2012).

Bache et al. (2012) defined eight seismic stratigraphic units above acoustic basement within the Reinga Basin using seismic stratigraphic methods after Vail et al. (1977) on all available 2-D seismic data in the Reinga, Northland and Taranaki Basins, and well control from the Northland and Taranaki Basins. These seismic stratigraphic units have been correlated, where possible to formations sampled in tied wells, or from potential correlative formations in the Taranaki Basin where reflections do no tie to any wells (Bache et al., 2012). The eight seismic stratigraphic units defined by Bache et al. (2012) are termed U1-U8, and are bounded by basal unconformities UB1-UB7 (i.e. the basal surface of U2 is UB1 and the basal surface of U3 is
UB2)(Figure 1.3). Which define five phases of development for the Reinga Basin: Phase 1 – Late Cretaceous extension, Phase 2 – Late Cretaceous-Eocene sedimentary infilling, Phase 3 – Late Eocene contraction, Phase 4 – Oligocene-Early Miocene subsidence, and Phase 5 – Middle Miocene contraction. These tectonic interpretations agree well with previous interpretation of the basin by Herzer et al. (1997). However, the timing of the early Cenozoic compressional event (Herzer et al., 1997), has been revised from early Miocene to Late Eocene.

Figure 1.3 – Interpreted seismic line from the Reinga Basin showing the stratigraphic interpretations of Bache et al. (2012). Note that only units U1 to U7 are labelled, and U1 and U2 are not differentiated (after Figure 5. Bache et al. (2012)).

Units U1 and U2 fill extensional sub-basins, and are characterised by divergent seismic reflections and sub-parallel reflections respectively. U2 reflections onlap onto U1, which is not present in all sub-basins and it may be difficult to differential between the two units. The upper contact of U2, UB2 is erosional in the southeast of the basin and appears as a high-amplitude reflection in the northwest. U1 and U2 are unable to be directly tied to any wells due to the West Norfolk Ridge basement high, which isolates these units from direct correlation with units in neighbouring basins. However, correlation to seismic units in the Deepwater Taranaki Basin which are tied to two wells in the Taranaki Basin indicate that U1 and U2 may represent a deltaic coal-rich sequence overlain by transgressive Cretaceous marine sequence (Taranaki Basin equivalent – Pakawau Group-Rapoki and North Cape formations)(Bache et al., 2012).

U3 and U4 are characterised by parallel seismic reflections which are folded in the northwest of the basin. U4 is distinguishable from U3 on the basis of its high-amplitude. U3 may onlap and bury some topographic highs within the basin; however U3 and U4 are observed folded
into Wanganella Ridge in the west of the basin. U3 and U4 are inferred to correlate with the Turi Formation, which has been observed in both the Taranaki and Northland Basins. Bache et al. (2012) infer ages of Paleocene-Early Eocene and Early to Late Eocene for U3 and U4 respectively. U1 to U4 are folded and eroded by a planation surface on the southern portion of West Norfolk Ridge (Figure 1.3).

U5 is confined to depressions formed by the folding of seismic units U1 to U4. The undeformed sub–horizontal reflections of U5 onlap on to the deformed reflections of U4, this basal onlap surface is termed UB4. U6 is defined by the basal contact, UB5, truncating U5 reflections and the UB4 surface, and the presence of volcanic edifices at the base of the seismic unit.

U6 reflections may onlap basin margins, and is not present on the Reinga, West Norfolk or Lord Howe ridges, indicating that they had already formed bathymetric highs at the deposition of U6. Some of the planation surfaces mentioned earlier that affect these bathymetric highs have some areas overlain with U6 indicating its deposition occurred after planation, or possibly coevally with the planation of these surfaces. The UB5 unconformity approximately coincides with the ‘P1’ reflection of Herzer (1995) and Herzer et al. (1997) in the Northland Basin, which is inferred to represent the base of an Oligocene condensed carbonate sequence (Herzer, 1995; Herzer et al., 1997; Bache et al., 2012). This ‘P1’ reflection is easily tied to wells in the Taranaki Basin, on the basis that a velocity inversion occurs at the location of the ‘P1’ reflection, which ties to the base of Oligocene limestones (Herzer et al., 1997). In the Waka Nui-1 well, the ‘P1’ reflection coincides with the base of the 185 m thick Oligocene Tikorangi Formation, which occurs directly above the first downhole occurrence of Eocene foraminifera (Bache et al., 2012). The condensed carbonate sequence, which is represented by the ‘P1’ reflection, occurs throughout much of New Zealand as a limestone/greensand sequence spanning the Oligocene and coincides with a major hiatus in the Southwest Pacific stratigraphic record, referred to as “the Marshall Paraconformity” (Carter, 1985; Fulthorpe et al., 1996; Herzer et al., 1997). Bache et al. (2012) interpret that the UB4-U5-UB5 sequence is of late Eocene age (UB5 falls below or at the base of the Tikorangi Formation in the Arika-1 well), and the Oligocene Tikorangi Formation is the basal unit in the U6 seismic unit. U6 coincides with the Tikorangi, Taimana, and Lower Maunganui Formations, in all wells used by Bache et al. (2012) giving U6 an age range from Oligocene to Early Miocene.

U7 may be distinguished from U6 in the centre of the Reinga Basin by a package of high-amplitude reflections. However, along the northeastern flank of Wanganella Ridge, U7 is
strongly unconformable and onlaps onto U6, which is included in the sequence tilted on the ridge. This indicates that U7 was deposited directly after the uplift of Wanganella Ridge (Bache et al., 2012). Bache et al. (2012) indicate that U7 represents the Reinga Basin equivalent of the Middle-Late Miocene Moki and Mohakatino Formations, which consist of sandstone-dominated turbidites and deepwater volcanoclastics respectively. Whilst Bache et al. (2012) do not provide an interpretation of U8, presumably it represents the Reinga Basin equivalent of the Plio-Pleistocene Giant Foresets Formation of the Taranaki Basin (Hansen and Kamp, 2006). However, U8 lacks the distinctive clinoform reflections associated with the Giant Foresets Formation, and probably represents hemipelagic drift in the centre of the basin.

In the extreme southeast of the Reinga Basin, the stratigraphy may be disrupted by the presence of the Northland Allochthon, which is thought to extend at least 100 km into the basin (Uruski, 2010). The Northland Allochthon is an allochthonous assemblage of Cretaceous to Oligocene rocks that extends across the majority of the Northland Peninsula and a significant distance offshore to the west (Ballance and Spörli, 1979; Malpas et al., 1992; Isaac, 1994). The allochthonous assemblage consists of ophiolitic massifs and a late Cretaceous to Oligocene sedimentary sequence that were obducted in the Late Oligocene onto autochthonous Tertiary sediments, and are unconformably overlain by Miocene sediments of the Waitemata Group (Ballance and Spörli, 1979; Malpas et al., 1992).

1.3.2.2 Waka Nui-1 well
The only well directly tied to the Reinga Basin with seismic data is the Waka Nui-1 well, which is located 250 km southeast of the Reinga Basin in the Northland Basin. The Waka Nui-1 well was spudded in April 1999 in 1455 m of water within the Northland Basin (Conoco, 2000). Waka Nui-1 reached a total depth of 3682 m bottoming out in Murihiku Group sandstones, with the majority of the sequence consisting of siltstones, claystones, and marls (summarised in Table 1.1)(Milne and Quick, 1999). No hydrocarbons were encountered during drilling, except for thin coal beds within the Murihiku Basement (Conoco, 2000). An $^{40}Ar/^{39}Ar$ age of 158 ± 2 Ma was obtained for a diabase sill encountered at 3602 m within Murihiku coal measures (Foland, 1999).
### Table 1.1 - Formation summary of Waka Nui -1 (after Milne and Quick, 1999)

<table>
<thead>
<tr>
<th>Formation</th>
<th>Age</th>
<th>Top (m)</th>
<th>Thickness (m)</th>
<th>Lithology</th>
</tr>
</thead>
<tbody>
<tr>
<td>No returns</td>
<td></td>
<td>1478.9</td>
<td>673.1</td>
<td>No returns</td>
</tr>
<tr>
<td>Rotokare Group (undifferentated)</td>
<td>Recent - Early Pliocene</td>
<td>2152.0</td>
<td>19.0</td>
<td>Marl</td>
</tr>
<tr>
<td>Rotokare Group, Ariki Formation</td>
<td>Early Pliocene - Middle Miocene</td>
<td>2171.0</td>
<td>176.0</td>
<td>Marl with minor interbedded Volcaniclastics</td>
</tr>
<tr>
<td>Wai-Iti Group, Mohakatino Formation</td>
<td>Early Miocene</td>
<td>2347.0</td>
<td>178.0</td>
<td>Interbedded Siltstone, Sandstone and Volcaniclastics</td>
</tr>
<tr>
<td>Wai-Iti Group, Taimana Formation</td>
<td>Early Miocene</td>
<td>2525.0</td>
<td>54.0</td>
<td>Interbedded Siltstone/ Volcaniclastics</td>
</tr>
<tr>
<td>Ngatoro Group, Tikorangi Formation</td>
<td>Late Oligocene - Late Eocene</td>
<td>2579.0</td>
<td>185.0</td>
<td>Limestone (mudstone)</td>
</tr>
<tr>
<td>Moa Group, Calcareous Claystone</td>
<td>Late Eocene</td>
<td>2764.0</td>
<td>59.0</td>
<td>Marl</td>
</tr>
<tr>
<td>Intra-Moa Group Marl</td>
<td>Middle- Early Eocene</td>
<td>2823.0</td>
<td>128.0</td>
<td>Argillaceous Limestone, shaling downwards.</td>
</tr>
<tr>
<td>Moa Group, Upper Turi Formation</td>
<td>Late Paleocene</td>
<td>2951.0</td>
<td>141.0</td>
<td>Claystone</td>
</tr>
<tr>
<td>Moa Group, Turi Formation, Waipawa Black Shale Member</td>
<td>Late Paleocene</td>
<td>3092.0</td>
<td>26.0</td>
<td>Claystone</td>
</tr>
<tr>
<td>Moa Group, Lower Turi Formation</td>
<td>Early Paleocene</td>
<td>3118.0</td>
<td>356.0</td>
<td>Claystone</td>
</tr>
<tr>
<td>Kapuni Group, transgressive Sandstone</td>
<td>Early Paleocene</td>
<td>3474.0</td>
<td>70.5</td>
<td>Conglomerate</td>
</tr>
<tr>
<td>Murihiku Group, weathered Coal Measures</td>
<td>Middle Jurassic</td>
<td>3544.5</td>
<td>31.5</td>
<td>Claystone</td>
</tr>
<tr>
<td>Murihiku Group, Coal Measures</td>
<td>Middle Jurassic</td>
<td>3576.0</td>
<td>75.0</td>
<td>Interbedded Siltstone, Sandstone, Coal with a thin diabase sill (3602 m)</td>
</tr>
<tr>
<td>Murihiku Group, Volcanics</td>
<td>Middle Jurassic</td>
<td>3651.0</td>
<td>31.0</td>
<td>Tuff, Sandstone</td>
</tr>
</tbody>
</table>

### 1.3.3 Igneous activity

#### 1.3.3.1 Late Oligocene – Early Miocene

Several authors have reported the presence of Late Oligocene-Miocene volcanic activity within the Reinga and Northland Basins (Isaac, 1994; Herzer, 1995; Herzer et al., 1997; Stagpoole et al., 2009). These conical volcanic bodies are predominantly located in the southeast of the Reinga Basin and have been buried by onlapping strata of the U6 and younger seismic units (Herzer et al., 1997; Bache et al., 2012). The bases of these volcanic edifices are located directly above the UB5/P1 reflection, and where data quality is high, reflections within these volcanic units downlap on the UB5/P1 surface, indicating that eruption occurred after the deposition of the Oligocene condensed carbonate sequence. The restriction of these volcanic edifices to the southeast of the basin and their Oligocene-Miocene age suggests a Northland Arc correlation (Herzer et al., 1997). This is consistent with seismic stratigraphy of volcanic bodies in the Northland Basin directly south east of the Reinga Basin by Herzer (1995), which gave ages of 22 to 16 Ma for Northland Basin volcanoes.
1.3.3.2 Pliocene-Pleistocene

Herzer et al. (1997) also report possible Pliocene intrusive and extrusive igneous bodies associated with Wanganella Ridge in the northwest of the Reinga Basin. These volcanic bodies appear as detached anticlines or diapiric bodies within the uplifted wedge of Wanganella Ridge. Herzer et al. (1997) also infer that a short horizontal reflection in the Pliocene/Pleistocene sequence on the AGSO 114-04 seismic line above a slight zone of upwarped reflections to be a sill or flow (this will be discussed further in Chapter 3, see Figure 3.24). Based on the limited data available to Herzer et al. (1997), they infer that these igneous bodies appear to reduce in size and density away from Wanganella Ridge suggesting volcanism has some association with the uplift of Wanganella Ridge. Some small conical seafloor peaks also occur in the Wanganella Ridge area, one of which has been dredged (ORSTOM dredge GO353) recovering late Pliocene intraplate alkaline basalts and hyaloclastites (Figure 1.4) (Monzier and Vallot, 1983; Herzer et al., 1997; Mortimer et al., 1998). To date, ORSTOM dredge GO353 is the only sample collected from a volcanic edifice that protrudes through sediment cover in the Reinga Basin (Figure 1.4). Whole rock $^{40}$Ar/$^{39}$Ar dating of a fresh olivine alkalic basalt (GNS sample –P57143) from the GO353 dredge gave a statistically good Pleistocene age of $2.27 \pm 0.02$ Ma (Mortimer et al., 1998). Whilst these Plio-Pleistocene igneous bodies show an association with the uplifted Wanganella Ridge, Herzer et al. (1997) prefer a hotspot origin for these Pliocene volcanics due to their young age relative to the inferred Early to Middle Miocene uplift of Wanganella Ridge (Herzer et al., 1997; Bache et al., 2012).

Figure 1.4 – Line drawing of the AUS204 seismic line across West Norfolk Ridge, Wanganella Ridge and the Reinga Basin showing the location of the GO353 dredge site, which sampled Pleistocene olivine alkali basalt (modified from Figure 5b of Herzer et al. 1997).
1.3.4 Basin evolution

The Reinga Basin was formed initially by crustal extension and thinning in the Cretaceous resulting in crustal thinning of 35-40% (Herzer et al., 1997). Extension, graben formation and deposition of the U1 syn-rift deposits probably ceased at the inception of seafloor spreading in the Tasman Sea at around 85-80 Ma (Phase 1 – Figure 1.5) (Bache et al., 2012). Thermal subsidence followed until the late Eocene depositing U2-U4 in marine conditions, U3-U4 are correlatives of the Turi Formation, a dark-coloured marine mudstone (Phase 2) (Bache et al., 2012). Compression occurred in the northern part of the basin in the Late Eocene, folding U1-U4, and uplifting them to shallow levels to be eroded (planation surfaces of Bache et al. 2012) and subsequent deposition of U5 (Phase 3). This is followed by Oligocene to Miocene subsidence and Northland Arc volcanism in the southern portion of the basin depositing seismic unit U6, which includes the Tikorangi Limestone (Phase 4) (Herzer, 1995; Herzer et al., 1997; Bache et al., 2012). Compression occurred again in the Middle Miocene resulting in the uplift of Wanganella Ridge preceding the deposition of U7 and U8 (Phase 5). Alkaline basalt composition intrusive and extrusive volcanism occurred along a northeast trend on the east side of Wanganella Ridge in the Pliocene/Pleistocene (Herzer et al., 1997; Mortimer et al., 1998).

Figure 1.5 - Tectonic evolution of the Reinga Basin based of seismic stratigraphy by Bache et al. 2012. Phase 1 – Extension during the late Cretaceous due to rifting of Zealandia from Australia. Phase 2 – late Cretaceous to Eocene thermal subsidence and deep marine sedimentation. Phase 3 – Late Eocene contraction resulting in the folding of U1 to U4 and uplift Reinga and West Norfolk ridges. Phase 4 – Oligocene to Miocene basin wide subsidence, Northland Arc volcanism in the southeast of the basin. Phase 5 – Middle Miocene contraction resulting in the uplift of Wanganella Ridge. (modified from figures 2 and 9 in Bache et al. 2012).
1.4 Thesis structure

This thesis has been divided into 5 chapters which outline the theory controlling amplitude anomaly formation and some examples, multi-channel seismic processing methods and initial interpretations, basic seismic inversion theory, seismic inversion results, and a summary discussion on the finding of this thesis, their significance and limitations.

Chapter 2 provides an overview of the theory of seismic reflection amplitude from vertically incident reflections to pore fluid substitution, as well as providing several geological models which cause anomalous amplitude reflections which are applicable to the Reinga Basin. These include; gas hydrate, igneous intrusions, and sand intrusions.

Chapter 3 outlines the preserved amplitude processing flow applied to the STRATUS-2D-002 line, and initial interpretation of areas of anomalous reflection amplitude.

Chapter 4 outlines basic seismic inversion theory, requirements and theory for full waveform inversion using the Hampson-Russell Suite inversion software, background impedance model generation and presents the results of six inversion runs. The resulting inverted impedances are discussed for each anomaly defined in Chapter 3 to provide more constraint on the lithologies which are causing these areas of anomalous amplitude.
Chapter 2

Amplitude anomalies

In reflection seismology, an amplitude anomaly is defined as a localised increase or decrease in reflection amplitude compared to the background amplitude (Sheriff, 2002). There are various causes for the presence of amplitude anomalies, ranging from geometric or velocity focusing, actual reflections from the underlying geology, for instance hydrocarbon accumulations or igneous intrusions, or processing artifacts. This chapter discusses some of the theory of seismic reflection amplitudes, touching on elastic rock parameters and several geological scenarios in which amplitude anomalies may form, which may have relevance to the Reinga Basin, including free gas accumulation, bottom simulating reflections (gas hydrates), igneous intrusions, and sand intrusions.

2.1 Seismic reflection amplitude theory

Seismic reflection amplitude is a function of the acoustic impedance contrast (reflection coefficient) between the lithology above and below a reflecting boundary (Yilmaz, 2001). Acoustic impedance of any material is the product of P-wave velocity and density;

\[ Z = \rho_1 V_1, \]  

(2.1) (Sheriff, 2002)

where \( Z \) is acoustic impedance, \( \rho \) is density (g/cc) and \( V \) is P-wave velocity (m/s).

The theoretical vertical reflection coefficient (and the reflected energy if the source energy is accurately known) for a normally incident plane wave can be calculated as the ratio of the difference in acoustic impedance between the upper and lower layer and the sum of the impedance of both layers;

\[ R = \frac{\rho_2 V_2 - \rho_1 V_1}{\rho_2 V_2 + \rho_1 V_1} = \frac{Z_2 - Z_1}{Z_2 + Z_1}, \]  

(2.2) (Sheriff, 2002)
where $R$ is the reflection coefficient, $\rho_2$ and $\rho_1$ are the densities of the lower and upper layers, respectively, and $V_1$ and $V_2$ are the P-wave velocities of the upper and lower layers, respectively.

In ideal elastic media, the sum of the reflected and transmitted energy at any given boundary must be equal to the energy that is incident on the boundary. The ideal transmission coefficient will be:

$$T = 1 - R,$$

(2.3)

where $T$ is the transmission coefficient and $R$ is the reflection coefficient of the reflecting boundary. In reality, however, some energy will be dispersed by irregularities on the reflecting boundary and some energy will be attenuated by conversion to heat, and non-elastic deformation (permanent deformation) of the geology in the near field.

Therefore, considering a boundary between perfectly elastic media with a zero-offset (vertical) acquisition geometry where we know original source energy and account for spherical spreading of the source energy (for which we need to know the depth to the reflector), we could determine the reflection coefficient of the reflector based on the ratio of initial source energy to recorded energy. In this ‘perfect’ case, we can also calculate the velocity of each layer because we know the depth to the reflector. In a real situation, however, even for this simple case we do not know the depth to the reflector, therefore we are unable to account for spherical spreading of the source energy without making an assumption of the velocity of the upper layer to estimate depth.

For non-vertically incident reflections, the situation becomes more complex. In almost all real-world cases, recorded reflected energy will not be normally incident on a reflector. For example, standard 2-D marine seismic reflection acquisition consists of a hydrophone array up to 12 km long with a near offset of ~130 m, this would result in reflection raypaths for a reflector at 2 km depth to range between ~1° and 70°. This results in two major changes to the determination of recorded reflected energy; (1) reflections recorded at higher angles will have longer path lengths, resulting in more attenuation, and will have ‘delayed’ arrival times compared to a normally incident reflection, and (2) non-normal reflection of a P-wave at a reflection surface will result in wave type conversions: i.e. some P-wave energy will be converted into S-wave energy. The first problem is relatively easy to deal with, progressively longer raypaths to each receiver result in a hyperbolic increase in travel-time (normal moveout - NMO). Where data are converted into common midpoint (CMP) gathers i.e., all traces from different shot and receiver pairs for which the reflection point is at the same
location in the subsurface, a normal moveout (NMO) correction can be applied to the data (Yilmaz, 2001). This NMO correction accounts for the extra travel time taken for a reflection from a point on a horizontal reflector to reach receivers with increasing offset, thereby flattening reflections on the CMP gather. By calculating an NMO correction we are also able to place some constraints on the velocity structure of the subsurface, enabling us to better account for spherical divergence of the source energy and place broad constraints on the P-wave velocity within the subsurface.

The second issue with non-vertical reflections and calculating their associated reflection coefficients is wave-type conversion, which is far more difficult to account for than the increasing travel times mentioned previously. If a P-wave is incident at an angle to the reflection surface, varying amounts of energy will be distributed as: transmitted P-wave energy (Tpp), reflected P-wave energy (Rpp), transmitted S-wave energy (Tps) and reflected S-wave energy (Rps) which can be expressed via Snell’s Law:

\[
P_i = \frac{\sin \theta_{pp}}{V_{p1}} = \frac{\sin \theta_{ps}}{V_{s1}} = \frac{\sin \theta_{rp}}{V_{p2}} = \frac{\sin \theta_{rs}}{V_{s2}},
\]

(2.4) modified from (Sheriff, 2002)

where \( P_i \) is the incident P-wave energy, \( V_{p1} \) and \( V_{s1} \) are the P-wave and S-wave velocities in the upper layer, \( V_{p2} \) and \( V_{s2} \) are the P-wave and S-wave velocities in the lower layer, \( \theta_{pp} \) and \( \theta_{ps} \) are the angles of reflection of the P-wave and S-wave respectively and \( \theta_{rp} \) and \( \theta_{rs} \) are the angle of refraction of the transmitted P-wave and S-wave respectively (Figure 2.1). The ratio of reflected, transmitted and converted energy for a given angle of incidence and elastic parameters across a reflecting interface can be modelled exactly using a set of equations derived in the early 20\(^{th}\) century by Karl Bernhard Zoeppritz, known as the Zoeppritz’s equations (Sheriff, 2002). Several modern approximations of the Zoeppritz equations exist, such as the Aki and Richards (1980) and Shuey (1985) approximations.
In marine seismic methods we can only record P-wave data as S-waves are unable to propagate through the water column. Therefore the conversion of incident P-wave energy into differing proportions of P- and S-wave energy, dependent on the angle of incidence, will express itself in the recorded data as a change in recorded seismic amplitude with offset. Where attenuation is correctly accounted for, the Zoeppritz equations and their approximations can be used to model physical rock properties (Vp, Vs and density) based on the variation in recorded P-wave amplitude with increasing angle, which will have components of both the P-wave impedance contrast across the reflection boundary and the amount of conversion to S-wave energy (dependent on angle and S-wave impedance contrast) at the reflecting surface.

The Hampson-Russell Suite inversions discussed in Chapter 4 use a joint inversion scheme based on a modified version of the Aki and Richards approximation of the Zoeppritz equations (Hampson et al., 2005).
2.2 Elastic rock properties

Elastic rock properties provide the fundamental link between seismic data and the underlying geology. They govern how seismic energy travels through the subsurface. Changes in these elastic constants results in reflection, refraction and attenuation of seismic energy. Using Amplitude Variation with Offset (AVO) and inversion methods which are based on these fundamental elastic parameters, we can attempt to extract elastic rock property data from seismic data and use it to infer the lithologies causing reflections in the seismic data.

2.2.1 Bulk elastic parameters

Propagation of seismic energy in geological media is physically governed by the elastic parameters of the media involved. For small non-permanent deformations (i.e. elastic) such as a far-field seismic signal, Hooke’s Law is valid which means that strain is proportional to stress (Sheriff, 2002). P-wave velocity is dependent on the bulk modulus of compressibility, \( K \); the bulk shear modulus, \( \mu \); and the bulk density, \( \rho \); of the material. S-wave velocity is only dependent on the shear modulus, and the bulk density of the material. The bulk modulus of compressibility is defined as the dilatational volume change under an isotropic stress field:

\[
K = \frac{\Delta P}{\Delta V/V}, \quad (2.5) \quad \text{(Sheriff, 2002)}
\]

where \( \Delta P \) is the pressure change and \( \Delta V/V \) is the dilatational volume change (i.e. change in volume with respect to the original volume).

The shear modulus is defined by the amount of shear displacement that occurs for a given application of a shear force:

\[
\mu = \frac{\Delta F/A}{\Delta L/L}, \quad (2.6) \quad \text{(Sheriff, 2002)}
\]

where \( \Delta F \) is the applied shear force, \( A \) is the cross sectional area over which the force is applied and \( \Delta L/L \) is the ratio of change in length to original length.

P-wave velocity, \( V_p \), is defined by the following equation:

\[
V_p = \sqrt{\frac{K + 4\mu/3}{\rho}}, \quad (2.7) \quad \text{(Sheriff, 2002; Schon, 2004)}
\]

S-wave velocity, \( V_s \), can be defined as:
Note that density, \( \rho \), perhaps counter-intuitively, is in the denominator in these two equations. Whilst both P-wave and S-wave velocities for a material can be defined by \( K \), \( \mu \) and \( \rho \) above, there are three other elastic parameters which are useful for defining elastic rock properties.

1. Young’s Modulus, \( E \), which is essentially the stretchability of a rod of material:

\[
E = \frac{\Delta F/A}{\Delta L/L},
\]

where \( \Delta F/A \) is stress (the force per unit area) and \( \Delta L/L \) is the ratio of change in length to original length.

2. Lame’s Constant, \( \lambda \), is:

\[
\lambda = K - \frac{2\mu}{3},
\]

3. Poisson’s ratio, \( \sigma \), is the ratio of longitudinal to transverse strain:

\[
\sigma = \frac{\Delta W/W}{\Delta L/L},
\]

where \( \Delta W/W \) is the ratio of contraction when a rod is elongated and \( \Delta L/L \) is the amount of elongation that occurs.

Poisson’s ratio can vary from 0 to 0.5; \( \sigma = 0.5 \) for fluids. A Poisson ratio of 0.25 is termed a Poisson solid, which indicates that the shear modulus, \( \mu \), is equal to Lame’s constant, \( \lambda \).

(2.12) (Sheriff, 2002)

If accurate values of P- and S-wave velocity and density can be obtained, then other elastic constants, such as Poisson’s ratio, can be determined which may be able to be used to infer changes in pore fluid composition, i.e. gas filled sands vs. brine filled sands, which may cause areas of anomalous amplitude.

### 2.2.2 Fluid substitution

As shown in Section 2.2.1 above, P- and S-wave velocities are dependent on the bulk modulus of compressibility, the shear modulus, and the bulk density of the material (equations 2.7 and 2.8 respectively). However in porous materials such as a sandstone, the
porosity and fluids filling the pore spaces will have a significant effect on the bulk modulus of compressibility and the bulk density. In a simple case, where we only look at the effect of porosity on bulk density, a water saturated quartz sandstone with 30% porosity has 30% of the total volume with a lower density than if the sample had no porosity, therefore the bulk density of the sample will be lower in this case, 2.19 g/cm$^3$ opposed to 2.7 g/cm$^3$. Ignoring the effect of porosity on the bulk modulus and shear modulus, this would result in an ~10% increase in P-wave velocity for a sandstone with 30% porosity compared to a non-porous sandstone with a bulk modulus of 10 GPa and a shear modulus of 5 GPa. However, in reality we see a decrease in P-wave velocity with porosity as the whole rock bulk and shear moduli also decrease with increasing porosity. The effect of pore fluids and porosity on the elastic moduli for isotropic drained porous media can be defined using a simplified formulation of Gassmann’s equation (Gassmann, 1951; Berryman, 1999; Han and Batzle, 2004; Simm, 2007). Berryman (1999), using the assumptions of Gassmann’s equation, indicates that pore fluid composition has no effect on the shear modulus (i.e. $\mu_{dry} = \mu_{sat}$), therefore the only effect of pore fluid composition on S-wave velocity is due to the change in bulk density. The bulk modulus, however, is dependent on the porosity and fluid composition as:

$$K_{sat} = K_{dry} + \Delta K_{dry} \quad (2.13) \quad (Han \ and \ Batzle, \ 2004)$$

$$\Delta K_{dry} = \frac{K_0(1 - \frac{K_{dry}}{K_0})^2}{1 - \frac{\Phi K_{dry}}{K_0} + \frac{\Phi K_\Phi}{K_f}} \quad (14) \quad (Han \ and \ Batzle, \ 2004)$$

where $K_{sat}$ is the wet bulk modulus, $K_{dry}$ is the dry frame bulk modulus (i.e. dry rock), $K_0$ is the mineral bulk modulus and $K_f$ is the pore fluid bulk modulus. $\Phi$ is the porosity of the sample. The fluid bulk modulus is proportional to the bulk modulii of each fluid phase if there is more than one, and their proportions; i.e.:

$$\frac{1}{K_f} = \frac{S_w}{K_w} + \frac{1-S_w}{K_h} \quad (2.15) \quad (Simm, \ 2007)$$

where $S_w$ is the proportion of water (water saturation), and $K_w$ and $K_h$ are the bulk modulii of water and hydrocarbons respectively. This is the Reuss average, which assumes that the fluid phases are uniformly mixed, if a patchy mix is assumed then the bulk modulus should be calculated as the Voigt average, i.e. Han and Batzle (2004).

The Han and Batzle (2004) formulation of Gassmann’s equation above shows the saturated bulk modulus is dependent on the porosity, pore fluid bulk modulus, mineral bulk modulus
and the dry frame bulk modulus (avoiding the use of the less tangible pore space stiffness modulus, $K_{\Phi}$, by relating it into the dry frame bulk modulus via the following relationship:

$$K_{\Phi} = \frac{\phi}{K_d} \frac{1}{K_0}$$

(2.15) (Simm, 2007))

Any change to pore space fluid will result in a change in P-wave velocity. However, as fluid composition has no effect on the shear modulus, apart from changing the bulk density, the S-wave velocity will stay relatively constant. However, because P-wave velocity is significantly changed by pore fluid composition, changes in pore fluid will result in a change in the $V_p/V_s$ ratio of the rock (and therefore Poisson’s ratio) which may result in anomalous amplitudes (Savic et al., 2005). If the change in $V_p/V_s$ is significant it may be measurable using AVO and waveform inversion techniques and used to indicate the presence of different pore fluids, i.e. water vs. gas as it is sensitive to pore fluid composition (Sheriff, 2002).

### 2.3 Examples of amplitude anomalies

#### 2.3.1 Bottom Simulating Reflections (BSR), gas hydrate and free gas

Gas hydrate is a solid substance composed of gas molecules, usually methane (but sometimes higher hydrocarbons) contained within a rigid framework of water molecules (Kvenvolden et al., 1993). Global gas hydrate accumulations are thought to represent $\sim 2 \times 10^{13} \text{ m}^3$ of natural gas (Fohrmann and Pecher, 2012). However these estimates vary by several orders of magnitude (Riedel et al., 2010). Gas hydrates are pressure and temperature sensitive, only forming at low temperatures and moderate pressures. At a maximum depth of $\sim 1000 \text{ m}$ below the seafloor, depending on water depth and the local geothermal gradient, gas hydrate is no longer stable and free gas will be present instead (Kvenvolden et al., 1993). This strong contrast between high velocity gas hydrate and low velocity free gas below results in the formation a strong negative polarity reflection known as a bottom simulating reflection (BSR) (Figure 2.2). The BSR parallels the seafloor as is caused by the temperature sensitive phase boundary between gas hydrate and free gas. These bottom simulating reflections generally have strong negative polarity (with respect to the seafloor) and may crosscut stratigraphy and have associated dim zones (Kvenvolden et al., 1993; Pecher et al., 1996).

Pure methane hydrate has a P-wave velocity of $\sim 3800 \text{ m/s}$, a shear wave velocity of $\sim 1950 \text{ m/s}$ and a density of $\sim 0.9 \text{ g/cm}^3$ (Gabitto and Tsouris, 2010; Riedel et al., 2010). This
is in strong contrast to the physical properties of unconsolidated sediments that we expect to find in the gas hydrate stability zone, with P-wave and S-wave velocities of ~1600-1800 m/s and 300-800 m/s respectively and densities of ~ 1.8 g/cm$^3$ (Riedel et al., 2010). This large impedance contrast between gas hydrate bearing sediment and no-hydrate/free gas bearing sediment is likely to result in strong negative amplitude anomalies. Amplitude anomalies associated with the base of the gas hydrate stability zone are likely to be represented by a velocity (or impedance profile) that shows high velocity zones (hydrate) underlain by a low velocity zone (free gas). Amplitude anomalies associated only with free gas are likely to exhibit low velocities with the surrounding velocities following the local background velocity trend.

Pecher et al. (1996) carried out 1-D full waveform inversion of a bottom simulating reflection (BSR) on the Peruvian Margin. The resulting inverted P-wave velocity profiles show a low velocity zone at the BSR where the average velocity drops from 2150 m/s to 1700 m/s. This velocity drop was interpreted to represent several percent of free gas below the base of gas hydrate stability (Pecher et al., 1996).

Mienert and Posewang (1999) present ocean bottom hydrophone velocity modelling from the Storegga Slide on the Norwegian Shelf. Two bottom simulation reflections and two possible gas chimneys were identified from seismic profiles over the hydrophone locations. Velocity modelling of the ocean bottom hydrophone data indicated that the interpreted hydrate bearing layers from the seismic profiles have P-wave velocities of ~1800 m/s and gas bearing layers have velocities as low as 1300 m/s.

Large areas of gas hydrate have also been identified in New Zealand’s Exclusive Economic Zone (EEZ), in particular along the Hikurangi Margin and Offshore Fiordland (Crutchley et al., 2007; Crutchley et al., 2010a; Crutchley et al., 2010b; Pecher et al., 2010; Fohrmann and Pecher, 2012). Possible gas hydrate has also been inferred in the Taranaki Basin through examination of high-amplitude anomalies in the predicted hydrate stability zone (Ogebule and Pecher, 2010).
2.3.2 Igneous intrusions

Igneous intrusions, in particular basalt sills, have excellent potential to cause anomalously high amplitude reflections due to high impedance contrast with the surrounding sediment. For example, in a situation where a thin but seismically resolvable basalt sill ($V_p=5180$ m/s and $\rho=2.81$ g/cm$^3$ (Schreiber and Fox, 1977)) is situated within a thick sand bed ($V_p=2400$ m/s and $\rho=2.2$ g/cm$^3$), the sill will have an upper surface reflection coefficient of approximately 0.47. Given that a ‘normal’ strong reflection such as the seafloor reflection has a reflection coefficient of ~0.2, this situation would result in a very strong positive reflection.

Igneous intrusions, in particular sills, have been interpreted in multichannel seismic data in sedimentary basins by numerous authors, e.g. Offshore Senegal (Rocchi et al., 2007; Hansen et al., 2008), Vøring and Møre Basins (Planke et al., 2005), Faroe-Shetland Basin (Trude et al., 2003; Hansen et al., 2004) and the Rockall Trough (Thomson, 2005; Hansen and Cartwright, 2006). These igneous intrusions within sedimentary basins are commonly sills, which are often saucer-shaped and are generally imaged as high-amplitude negative-polarity (reverse of seafloor) reflections (Hansen and Cartwright, 2006; Hansen et al., 2008). Not uncommonly, saucer shaped sills may have associated forced folds above them, where the overlying sediment has bulged up to accommodate the addition of material beneath in the sill (Trude et al., 2003; Hansen and Cartwright, 2006). These forced folds are characterised by convex upward continuous but potentially offset reflections reaching their highest deflection above a sill. These forced folds are often bounded by inward dipping reverse faults (Figure 2.3). The diameter of these forced-folds above sills in the Rockall Trough and Faeroe-Shetland Basin commonly range between 2 and 8 km (Trude et al., 2003; Hansen and Cartwright, 2006).
Whilst sills are easily identifiable in multichannel seismic data due to their high impedance contrast, they may often be difficult to properly interpret as thin sills may often have a thickness that is less than the vertical seismic resolution (Rayleigh Resolution $\lambda/4$, although $\lambda/2$ is generally required to resolve top and bottom reflections from a thin layer, where $\lambda$ is the dominant wavelength of the seismic data (Sheriff, 2002; Planke et al., 2005). Planke et al. (2005) suggest a detection limit as small as $\lambda/10$. A basalt sill with a P-wave velocity of 5.5 km/s will have a dominant wavelength of ~92 m (assuming 60 Hz dominant frequency); therefore it will be resolvable if it is thicker than 23 - 46 m and detectable if it is thicker than 9.2 m. The long dominant wavelength of high-velocity materials, such as basaltic sills will also cause the recorded reflections to exhibit low frequency ‘thick’ reflections. The negative polarity reflections from these sills suggests that strong reflections are occurring from the lower surface of these sills (high impedance over low impedance), perhaps the upper surfaces are more irregular causing more scattering and result in upper sill reflections without anomalously high amplitudes.
2.3.3 Sand intrusions

The intrusion of a relatively high impedance sand body into a muddy unit is likely to produce a moderately strong amplitude anomaly. If we take a sand body ($V_p=2400 \text{ m/s}$ and $\rho=2.2 \text{ g/cm}^3$) and intrude it into a clay-rich mudstone ($V_p=1850 \text{ m/s}$ and $\rho=2.3 \text{ g/cm}^3$), the resulting reflection coefficient from the upper surface of the sand body will be approximately 0.1. Whilst this is not a high reflection coefficient compared to the basalt sill example above, it will still result in a moderately strong reflection.

Sand intrusions have been reported from both 2-D and 3-D seismic data in various basins, including the Faeroe-Shetland Basin (Shoulders and Cartwright, 2004), the Tampen Spur region (Huuse and Mickelson, 2004) and the South Viking Graben (Huuse et al., 2004), several of which have been directly sampled through exploration wells.

Conical discordant reflections with similar acoustic properties, geometry, and scale have been recognised within Eocene claystone packages over large areas of the North Sea petroleum basins, where they have been tied to wells and interpreted as giant conical sandstone intrusions up to 50 m in thickness. Shoulders and Cartwright (2004) identified >100 conical amplitude anomalies from ~ 15 000 line km of 2-D seismic data and ~2600 km$^2$ 3-D seismic data within the Eocene to Oligocene interval appearing not to cross the locally folded Intra-Neogene unconformity. These conical anomalies show no significant variation in size and aspect, with all showing inward dipping conical shaped high amplitude reflections ~500 to 1000 m in diameter with 10-50° dips and opposing polarity upper and lower reflections.

Shoulders and Cartwright (2004) interpreted these anomalies to represent high-impedance sand bodies within low-impedance clays with the discordant nature indicating an intrusive origin. They also indicate that intrusion resulted in folding above the intrusion and deformation of the paleoseafloor where intrusion occurred at shallow levels.

Huuse et al. (2004), and Huuse and Mickelson (2004), also report sand intrusions within Eocene mudstones from the South Viking Graben and Tampen Spur areas of the North Sea, respectively. Both authors present conical seismic anomalies interpreted to be sand intrusions, and further confirm these features to be injected sands with exploration well data. These anomalies in the Tampen Spur region range from hundreds of metres to several kilometres across; the conical anomaly geometries dip inward at 25-40° and may cross cut up to 300 m of strata (Figure 2.4) (Huuse and Mickelson, 2004). Intersection of some of these anomalies with hydrocarbon exploration wells has sampled 50-60 m thick sandstones within polygonal faulted mudstones (Huuse et al., 2004; Huuse and Mickelson, 2004). Whilst intruded into
polygonal faulted areas, these sandstone intrusions appear to form independently, except where polygonal faults are favourably oriented. In some cases polygonal faults are crosscut by the conical amplitude anomalies.(Huuse et al., 2004; Huuse and Mickelson, 2004).

Figure 2.4 - Discordant conical amplitude anomalies in the Tampen Spur area after Huuse and Mickelson (2004), Figure 8a. Note that all anomalies show high amplitude, a “V” shape in 2-D and do not extend above the reflection at ~1500 ms.
3.1 Introduction

From late 2008 to early 2009 New Zealand Petroleum and Mineral and CGGVeritas acquired two regional prospectivity multi-channel seismic surveys within the Reinga Basin, the STRATUS-2D survey and the Reinga-09 (REI09) survey (Figure 3.1). These surveys provide the only high-fold MCS data within the Reinga Basin, with all data previous to this being relatively low fold with either low numbers of channels or large shot and group spacings (e.g. the Mobil72 (single-channel), AGSO114 (120-channel - 50 m shot interval -25 m group interval) and RE93 (single-channel) surveys) (Herzer et al., 1997; Sutherland et al., 2012). The majority of the Reinga-2009 survey is privately housed in the CGGVeritas data library; however, Reinga-2009 lines 002, 005, 012, 020 and 028 are in the public domain. The entire set of STRATUS-2D sequences is publically available (Figure 3.2). For detailed information on the acquisition of all historic seismic data located in the Reinga Basin and the Tasman Frontier see Sutherland et al. (2012). All publicly available commercially processed migrated stacks from the Reinga and STRATUS surveys were examined to determine the presence of any amplitude anomalies. However, only STRATUS-2D sequence-002 was selected for further processing and interpretation due to the presence of several shallow seismic amplitude anomalies. The STRATUS-2D sequence 002 line will be referred to as STRATUS-2D-002 throughout this thesis.

This chapter outlines the preserved amplitude processing flow applied to the STRATUS-2D-002 line for both inversion and initial interpretations (Section 3.3). I will also present some initial observations and interpretations of six amplitude anomalies identified on STRATUS-2D-002 based on a series of different processing results, including; offset-limited stacks (Section 3.4.5), instantaneous amplitude (Section 3.4.6) and different length AGC windowing (Section 3.4.4).
Figure 3.1 – Location map for the entire Reinga-09 and STRATUS-2D surveys in the Reinga Basin, both publically available lines and private data housed by CGGVeritas are shown. The STRATUS-2D line is shown in green and the publically available Reinga-09 lines are indicated in black. All nearby well locations are shown, including Waka Nui-1 in the Northland Basin (modified after Stagpoole et al., 2009).

Figure 3.2 - Academic and public domain industry seismic data coverage within the Reinga Basin excluding the Mobi72 survey. This includes the publically available Reinga-09, STRATUS-2D, RE93, AGSO 123 and AGSO114 Surveys. The black and red line is STRATUS 2D-002 which is used in this study. Preserved amplitude processing was applied to the red section to ensure relative amplitudes were maintained for full waveform inversion (bathymetry provided by NIWA).
### 3.2 Data acquisition

The STRATUS-2D survey was acquired by CGGVeritas onboard the *Pacific Titan* to industry standard specifications (Table 3.1). A 4140 cubic inch 3-string gun array operating at 2000 psi and a depth of 6 m was used as the source. Data were collected with a 37.5 m shot interval. Data were recorded for 12 s using a 7950 m Sercel Seal digital streamer with 12.5 m group spacing and 636 active groups. The streamer was kept at a depth of 7±1 m during acquisition.

<table>
<thead>
<tr>
<th>Shot Point Spacing</th>
<th>37.5 m</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number of groups</td>
<td>636</td>
</tr>
<tr>
<td>Recording System</td>
<td>Sercel Seal – Digital – Solid Streamer</td>
</tr>
<tr>
<td>Group Spacing</td>
<td>12.5 m</td>
</tr>
<tr>
<td>Hydrophones per group</td>
<td>8</td>
</tr>
<tr>
<td>Streamer Length</td>
<td>7950 m</td>
</tr>
<tr>
<td>Near Offset</td>
<td>131 m</td>
</tr>
<tr>
<td>Record Length</td>
<td>12 s</td>
</tr>
<tr>
<td>Sampling Rate</td>
<td>2 ms</td>
</tr>
<tr>
<td>Low-Cut</td>
<td>Effectively 4.3 Hz</td>
</tr>
<tr>
<td>Hi-Cut</td>
<td>200 Hz@ 370 dB/Octave</td>
</tr>
<tr>
<td>Start-of-data delay</td>
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</tr>
<tr>
<td>Streamer Depth</td>
<td>7 m</td>
</tr>
<tr>
<td>Gun depth</td>
<td>6 m</td>
</tr>
<tr>
<td>Gun volume</td>
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</tr>
<tr>
<td>Gun strings</td>
<td>3</td>
</tr>
<tr>
<td>Gun Pressure</td>
<td>2000 Psi</td>
</tr>
<tr>
<td>Auxiliary Channels</td>
<td>36</td>
</tr>
<tr>
<td>Theoretical Fold</td>
<td>106</td>
</tr>
<tr>
<td>Noisy channels</td>
<td>Channel: 312 (removed as a channel edit during processing)</td>
</tr>
</tbody>
</table>

### 3.3 Processing methods

Seismic reflection processing was carried out using GNS Science’s Globe *Claritas* seismic processing suite (Ravens, 2004). Processing consisted of data formatting and editing, addition of geometry to the data headers, velocity modelling, amplitude recovery, stacking and migration. For data destined to be used for prestack waveform inversion, the processing flow was designed to limit any changes to the relative seismic amplitude, by avoiding the use of processes such as Automatic Gain Control (AGC), and frequency-wave number filters (FK-filters). A summary diagram of the processing flow can be found on page 38.
3.3.1 Data formatting and editing

Raw shot records initially in SEGD format were converted into Claritas-SEGY (CSGY), the non-standard SEGY file format used internally in Claritas (Ravens, 2004). Conversion to CSGY also includes the addition of basic header information such as SHOTID and CHANNEL to the shot record headers. All 36 auxiliary channels were removed and noisy traces (channel 312) listed in the observers logs were removed by zeroing all trace amplitudes (Brook et al., 2008a).

The number of shots used for further processing for inversion were limited to focus effort on features of interest and minimise processing time. The shot range chosen for further processing was SHOTID 3000-5350.

3.3.2 Static corrections

Two static corrections were applied to the data; a start of data delay (SOD) and a gun cable static correction. The SOD static correction for the STRATUS-2D data was -50 ms, i.e., the data were moved 50 ms up trace. A gun cable static correction of -8.6 ms was applied to shift the recording datum of the data from the depth of the gun and streamer, 6 m and 7 m respectively, to the sea surface. The shift was calculated by the following equation (3.1):

$$ gun\ cable\ static = \frac{(\text{gun depth} + \text{streamer depth})}{V_p\ in\ water} = \frac{(6\ m + 7\ m)}{1500\ m/s} = 8.6\ ms $$

3.3.3 Geometry

Initially, processing geometry was defined from a UKOOA (UK Offshore Operators Association) .p190 formatted survey file, containing the geographic location of all shot points and associated receivers during acquisition. A crooked-line geometry was defined with a 6.25 m CMP spacing consisting of 4.4 m inline bins centred around each CMP location (Figure 3.3). The geometry was developed in a Cartesian system where an initial CMP location is defined [i.e. CMP 100 (0,0)] as well as the location of a shot [i.e. Shot 1500 (0,0)] and the numbering and locations of CMP bins are defined relative to the original CMP-Shot location definitions. All recorded traces fell into an assigned CMP bin. The geometry was later updated during the shot interpolation and offset regularisation stage.
3.3.4 Shot gather muting

An offset-dependent mute was handpicked on shot gathers to apply as a front mute to attenuate direct arrival energy and refracted energy within shot records. Mute functions were generally picked for every 100th shot record, but in areas of high seafloor gradient, mute functions were picked for every 50th shot record. The time to which data were muted on each trace was stored in the trace header SOURCE_WATER to enable the same mute to be reapplied in later processing steps.

3.3.5 Shot interpolation and offset regularisation

A time domain offset regularisation was applied to interpolate data at locations where original noisy data had been removed or muted and to regularise the trace geometry to a 12.5 m trace spacing for an offset range of 130 to 8067.5 m (130-8067.5/12.5 m) for subsequent processes. Regularised traces were interpolated from windows of five live traces with a fourth order polynomial interpolation with a constant NMO correction of 1700 m/s. The NMO correction was removed after interpolation.
Any spatial aliasing in the data was minimised by decreasing the shot point to CMP ratio (Gülünay, 2003). The shot interval was interpolated to 12.5 m from an original spacing of 37.5 m, i.e., the shot interval was reduced to a third of the original spacing. Whilst the “standard” technique for limiting spatial aliasing is applying a broad wave number filter (k-filter) to an NMO corrected shot gather and dropping every other trace following Yilmaz (2001); such a method was not applied during this processing flow. However, various wave number filters were tested with wave number cut-offs as broad as $K = 0.7$ to $0.2$. All wave number filter tests resulted in significant removal of primary energy. As this results in the loss of relative amplitude preservation, making the data unsuitable for full waveform inversion, no wave number filters were applied in the final processing flow.

### 3.3.6 Amplitude recovery

Amplitude recovery was carried out by applying a spherical divergence gain correction to the data. Spherical divergence gain recovery (Claritas module: SPHDIV) was applied to the data to enhance imaging of deeper reflections by correcting for the apparent loss of incident energy (Figure 3.5). Signal energy at a given point decreases with distance from the source point due to the spherical expansion of the signal wave front away from the source, i.e., the same energy is spread over a greater area (ignoring attenuation)(Sheriff, 2002)(Figure 3.4). Signal attenuation via spherical spreading is proportional to the inverse square of distance from the source point in a homogeneous medium, as the surface area of a sphere is derived from the following equation:

\[
Surface\ Area = 4\pi r^2
\]  

Spherical divergence gain recovery corrects for spherical spreading by applying a time varying gain to each trace that increases proportionally with the square of the distance from the source for each shot. As the data are in time domain, distances used for the gain correction are approximated using travel times and stacking velocities.

No other gain corrections such as automatic gain control (AGC) were applied as they may change relative trace amplitudes making the data unsuitable for inversion because of the non-linear way in which they affect the amplitude of seismic traces.
Figure 3.4 – Example of spherical divergence energy spreading from a point source (in this case located at a depth of $r=6$) at $r = 0, 2, 4$, and 6. The energy at a given point on the wavefront at a given radius is proportional to $\frac{1}{4\pi r^2}$ of the initial energy.

Energy at reflection point of wave front:

$$E_r = \frac{E_{\text{initial}}}{4\pi r^2}$$

<table>
<thead>
<tr>
<th>$r$</th>
<th>$E_r$</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>$\frac{1}{4\pi 2^2} \approx 0.02$</td>
</tr>
<tr>
<td>4</td>
<td>$\frac{1}{4\pi 4^2} \approx 0.004$</td>
</tr>
<tr>
<td>6</td>
<td>$\frac{1}{4\pi 6^2} \approx 0.002$</td>
</tr>
</tbody>
</table>

Figure 3.5 - (top) Front muted shot record with no gain correction; SHOT 3500 and (bottom) Front muted shot record with spherical divergence gain correction applied; SHOT 3500 – showing significantly increased amplitudes with depth.
3.3.7 Prestack filtering

The STRATUS-2D data were recorded with an analogue low-cut filter of 4.3 Hz and a high-cut filter at 200 Hz with a taper of 370 dB/octave. However, some random noise may be added to the data after the analogue band-pass as an artefact of some processing techniques and from electrical noise during recording. Band-pass filtering, with filter corner frequencies of 5-10-100-130 Hz was carried out using the Claritas module FDFILT to remove unwanted high and low frequency noise, caused by swell noise on the streamer, ship noise and any electrical noise introduced during acquisition (Figure 3.6).

FDFILT is a zero phase trapezoidal band-pass filter; each trace is converted into the frequency domain using a fast Fourier transform. Amplitudes of frequencies outside the defined band-pass are zeroed, and inside the filter are left unchanged, a taper is applied to the edges of the filter to limit edge effects. Each trace is then returned to time domain using an inverse fast Fourier transform. The phase spectrum of each trace remains unchanged (Ravens, 2004).
3.3.8 Velocity analysis

Stacking velocities were interactively picked using the Claritas velocity picking tool CVA (Claritas Velocity Analysis). Velocities were picked on unmigrated CMP gathers primarily using the constant velocity gather (CVG) and semblance plot tools, with quality control predominantly carried out on the fly by viewing NMO-corrected CMP gathers and ensuring that reflections were properly flattened. The data were also restacked at regular intervals for quality control. The unmigrated CMP velocities were then used as the velocity model for application of a shot-domain prestack Kirchhoff time migration. NMO-corrected CMP gathers were output from the migration and an inverse NMO correction was applied using the premigration velocities. A second pass of velocity picking was carried out on these migrated CMP gathers using the same methods as the first pass picking on unmigrated CMP gathers. Second pass velocities were predominantly picked using semblance plots and NMO corrected gathers, as migration of the data resulted in very ‘clean’ semblance plots. These second pass post-migration velocities were used for the final prestack migration of the data (Figure 3.7).
3.3.9 Migration

Prestack Kirchhoff time migration was carried out in the shot domain to shift reflections to their true subsurface location and to collapse diffractions. Prestack migration was used instead of the less resource intensive dip moveout correction (DMO) and zero-offset migration approach as the output data are still in the prestack domain and therefore retain amplitude information with offset. As a result, the data may be used for AVO analysis and prestack waveform inversion. Whilst only DMO could have been applied and the resulting CMP gathers would still retain relative amplitudes and have the correct dip, diffractions would not have been collapsed, and cross cutting reflections would not have been dealt with as rigorously applying a prestack time migration (Yilmaz, 2001). Prestack migration was carried
out on filtered, offset regularised and shot interpolated shot gathers (Section 3.3.5) which had geometry information in the headers. A migration range of 4000 m was chosen for the final migration using the second pass velocity model, calculated from previously migrated CMP gathers. A migration angle of 65° was used to ensure all dipping events were correctly migrated. The data were output as 107 fold CMP gathers with a constant offset range in all CMPs of 130 – 8080 m with a 75 m trace interval. A NMO stretch mute of 80% was applied by the migration module (i.e. if a trace was stretched by 80% or greater between two NMO functions that portion of the trace was muted), and then a handpicked front mute was also applied to ensure only flat reflections were retained with minimal far offset tails. Migration quality control was achieved by producing stacks of the migrated and muted CMPs and by checking that all CMP gathers were correctly flattened when the front mute was picked. No residual NMO was applied as all CMP gathers were flattened using the post migration velocity model mentioned above.

3.3.10 Other processes

3.3.10.1 AGC (Automatic Gain Correction)
Automatic Gain Correction (AGC) was applied to the stacked PSTM data to enhance weak reflections for initial interpretations. Such corrections were not part of the main processing flow. Three different AGC windows were employed; 250 ms, 500 ms and 1000 ms. Shorter AGC windows result in higher relative amplitudes for weak reflections; however, they also limit the total amplitude variability in the data.

3.3.10.2 Angle stacking and limited-offset stacks
Limited angle stacks were produced in the Hampson-Russell software from angle gathers by stacking together particular angle ranges from the gather. A radon de-noise filter was applied to the prestack time-migrated CMP gathers before they were transformed in to angle gathers. Angle stacks with angle ranges of 0-45°, 0-5°, 5-10°, 10-15°, 15-20°, 20-25°, 25-30°, 30-35°, 35-40°, and 40-45° were produced. Limited offset stacks were also produced in Globe Claritas from migrated CMP gathers. These stacks are presented later in the chapter and were limited to offset ranges of 0-1000 m, 1000-2000 m and 2000-3000 m.
Figure 3.8 - Processing flow for STRATUS-Sequence002 - showing all processes applied in Globe Claritas, RAYINVR, and the Hampson-Russell Suite – Blue shaded boxes indicate processed data output presented within this thesis.
3.4 Anomaly observations

Preliminary interpretations of the modern Reinga Basin seismic data were carried out on all the industry processed angle stacks and pre-stack time migrations for the Reinga-2009 and STRATUS-2D surveys (these are not shown in this thesis). Only Sequence 2 from the STRATUS-2D survey was selected for reprocessing, using the processing flow detailed in the previous section as it contains numerous amplitude anomalies. Whilst the data destined to be used for inversion were processed to ensure relative amplitudes are retained, further processing was carried out on stacked sections such as AGC and angle stacking to improve reflection continuity and determine if there were any broad AVO effects for initial interpretations of the STRATUS-2D-002 line. Where possible, age interpreted horizons after Bache et al. (2012) have been used in the initial interpretations.

Five differently processed seismic sections of the STRATUS-2D-002 line were used for initial interpretations:

1. Near trace stacks,
2. Full-offset stacked prestack time migrated gathers (PSTM),
3. PSTM with AGC applied (1000 ms, 500 ms, and 250 ms AGC windows),
4. Offset limited stacks (0-1000 m, 1000-2000 m, 2000-3000 m),
5. Instantaneous amplitude (for each of the PSTM and offset limited stacks).

These are presented in the text as enlarged sections for each anomaly. Please note that the full-offset PSTM stack of each anomaly is presented in each figure as part a, except for the near trace stack. Full seismic sections showing these different processing flows and the location of Figures 3.10 through 3.22 are included as Plate 1. All seismic section enlargements of amplitude anomalies are oriented left-side northwest, right-side southeast.

Six major amplitude anomalies were identified from STRATUS-2D-002:

1. Anomaly A, a near seafloor very high amplitude positive reflection between CMP 18432 and 19403 (~6 km),
2. Anomaly B, high amplitude negative-polarity disturbed reflections below a fault bounded anticline with apparent dim amplitudes and velocity pull-up below between CMP 19571 and 19982 (~2.5 km),
3. Anomaly C, negative-polarity high amplitude reflection associated with anomaly B between CMP 19473 and 19570 (~600 m),
4. Anomaly D, the same as Anomaly B (~2.95 km),
5. Anomaly E, disturbed high amplitude reflection with apparent dim spots below between CMP 25260 and 25735 (~2.95 km),

6. Anomaly F, stratigraphically controlled negative polarity reflection, that extends from the southeastern bounding fault of Anomaly B updip along an erosion surface for approximately 2.9 km (CMP 19991 to 20451) losing amplitude to the southeast and becoming ‘patchy’ until CMP 21100 (a further ~4.2 km). The patchy portion of Anomaly F consists of well separated ~10-20 CMP (~65-130 m) weak-to-moderate amplitude anomalies.

Anomalies B, C, D and the lowest part of Anomaly E coincide with the late Early-Miocene Horizon UB6 of Bache et al. (2012)(Figure 3.9). The UB6 horizon is folded by the anticlines above Anomalies B and D. Anomaly F is confined to an erosion surface which appears to onlap the uppermost deformed reflection above Anomaly B and erodes the crest of the uppermost deformed reflection above Anomaly D.

3.4.1 Structures observed on STRATUS-2D-002

3.4.1.1 Faults

Faults on STRATUS-2D-002 are predominantly small (~5-10 ms throw) normal faults, which in general do not reach to within 150-200 m of the seafloor (approximately the UB9 surface). Whilst the majority of faults are these small extensional normal faults, there are two exceptions: (1) three relatively large reverse faults which bound the anticlines above Anomalies B and D, and (2) Wanganella and West Norfolk ridges to the northwest of the STRATUS-2D line which have uplifted along a pre-existing normal faults in response to Miocene compression (Bache et al., 2012). Interpretation of the full offset migrated stack indicates that the anticline above Anomaly B is reverse fault bounded on both its northwest and southeast sides. The fold above Anomaly D appears to be bounded by a reverse fault only to the northwest. Offsets on these bounding faults are on the order of 50 ms, and they do not cut any reflections above the UB9 onlap surfaces which bound the uppermost deformed reflections above the anomalies. Several small normal faults occur within the crest of the Anomaly D anticline.
Figure 3.9 - Annotated PSTM section of the STRATUS-2D-002 line. Anomalies B and D are indicated with blue letters B and D respectively. The pink horizon represents the erosion surface discussed in the text. The dark blue horizon represents the UB6 horizon after Bache et al. (2012), and the green horizon represents the UB9 onlap surface defined from this study. The anomaly B and D bounding faults are indicated in black, with their sense of motion shown.
3.4.1.2 Folds

Only two folds are imaged on STRATUS-2D-002, occurring above Anomalies B and D. These folds are broad open dome shaped anticlines, with limited extent, only extending as far as the bounding faults around Anomalies B and D mentioned above. The maximum amplitude of these anticlines is ~100-200 ms at the hinge of the fold. A similar anticline with high-amplitude reflections beneath is also visible on AGSO 114-04 (see Figure 3.24) in close proximity to Anomaly D. These folds show maximum deflection at the maximum thickness of Anomalies B and D beneath them, suggesting that there may be some relationship between the formation of these anticlines in an otherwise extension dominated setting and the underlying amplitude anomalies.

3.4.2 Near trace stack

Diffractions occur where a reflector discontinuity is present, i.e., where a reflector is cut by a fault or fractures and where a reflector is smaller than the first Fresnel zone of the data such as a irregular reflector topography (Knapp, 1991; Yilmaz, 2001). Therefore significant amounts of diffracted energy in the seismic data may indicate high degrees of reflector surface irregularity, reflector inhomogeneity, or reflector termination (either from visible faults, fractures or pinchouts).

A near trace stack of STRATUS-2D-002 was produced by selecting the near trace (offset ~130 m) from every shot gather in the survey, and subsequently stacking the selected traces. No NMO was applied and only broad frequency filtering; therefore this near trace stack is equivalent to single-channel seismic data set with a constant offset. Near trace stacks were used to qualitatively determine whether there was a significant amount of diffracted energy associated with the observed amplitude anomalies (Figure 3.10).

Anomalies B and C (especially B) show intense diffractions. In contrast, whilst Anomaly D has a very similar form to Anomaly B (i.e. bright reflection below an anticline), it shows very little diffracted energy. Small diffractions also occur around small normal faults. Each of the anomalies is discussed in terms of the characteristic diffractions found to be associated with them in more detail below and shown in Figure 3.10.
**Anomaly A**
Anomaly A has few associated diffractions, but all amplitudes below the anomaly are very dim, indicating low transmission through the feature. There are possible small diffractions between the seafloor and the top of anomaly A. However, these may be due to reverberation between the highly reflective Anomaly A reflector and the seafloor, or a processing artefact.

**Anomalies B, C and F**
Anomalies B and C are highly diffracted, exhibiting very large and strong diffractions that have comparable amplitudes to the first multiple seafloor reflection. These diffractions have diffraction tails up to 1.75 s long. Most of these come from Anomaly B; however there are definite diffractions with crests centred on Anomaly C and associated with the Anomaly B bounding faults. Anomaly F displays small ~50-100 ms diffractions below its continuous section and similar diffractions centred on high amplitudes in the patchy area of the anomaly.

**Anomaly D**
Anomaly D exhibits low amplitude diffractions, only showing some weak diffraction directly below the central flat high amplitude reflection, approximately 200-300 ms long. Small diffractions can also be seen associated with the bounding fault.

**Anomaly E**
Anomaly E exhibits a large number of diffractions associated with all high amplitude reflections; most of these high-amplitude diffractions are relatively large (i.e. <400 ms long).
3.4.3 Prestack Time Migrated Stack (PSTM)

A full-offset PSTM stack of the STRATUS-2D-002 line was produced by stacking all offsets from prestack Kirchhoff time migrated CMP gathers. In the full-offset PSTM stack all diffractions should be collapsed and all dipping reflections should be migrated into their true subsurface position, with no cross-cutting reflections occurring. Full-offset PSTM stacks of each anomaly are presented as part a. in Figures 3.11 to 3.18.

3.4.3.1 AGC applied to PSTM stacks

Automatic gain control (AGC) enhances weak amplitudes by equalising the energy in an entire trace, allowing for the interpretation of weak reflections. To apply AGC, the amplitude of each sample is calculated, then each sample in the trace is scaled by a scaling factor so that the average amplitude in a particular window is constant (usually equal to 1) down the trace (Ravens, 2004). Three different AGC windows were applied: 250, 500 and 1000 ms. Smaller AGC windows result in more consistent amplitudes in the stack, i.e. relatively weak reflections can become indistinguishable from strong reflections.
**Anomaly A (Figure 3.11)**

The 250 ms AGC stack shows that there are concordant reflections in the dim zone beneath Anomaly A, which appear to be continuous. This indicates that the low amplitude beneath Anomaly A is most likely due to poor transmission through the anomaly.

![Figure 3.11](image)

*Figure 3.11 – Anomaly A – PSTM with various AGC operator lengths: (a) No AGC, (b) 1000 ms, (c) 500 ms, (d) 250 ms. Below Anomaly A there is a pronounced dim zone in the PSTM, no AGC stack, this is probably due to low transmission through the highly reflective Anomaly A. In the AGC stacks, in particular the 250 ms AGC it is clear that the reflections just below Anomaly A are continuous extending to the NW Anomaly B bounding fault, which is indicated by the black line).*
Anomalies B, C and F (Figure 3.12)

All of AGC windows suggest that there is a continuous negative-polarity relatively low-frequency reflection at the top of Anomaly B. There is also strong indication, especially in the 250 ms AGC stack, that Anomaly C may be a continuation of this reflection at the top of Anomaly B. Whilst anomalies B and C appear to be continuous in these stacks, Anomaly C displays a higher frequency character than the Anomaly B reflection suggesting that if they are caused by the same reflector, either a change in the reflector (i.e. thickness or elastic properties) or a tuning effect must occur. Anomaly F shows no change between the different AGC operator lengths.

![Figure 3.12](image)

**Figure 3.12** – Anomalies B, C and F – PSTM with various AGC operator lengths: (a) No AGC, (b) 1000 ms, (c) 500 ms, (d) 250 ms. Note that there is a clear continuous reflection at the top of Anomaly B in all of the stacks, the centre of the anomaly however is characterised by chaotic reflections. Note that Anomaly C appears to be a continuation of the upper reflection from anomaly B.
**Anomaly D (Figure 3.13)**

The distinct saucer shaper morphology of Anomaly D becomes apparent in the 500 ms and 250 ms AGC stacks, as well as a short parallel reflection below Anomaly D to the southeast. Both reflections are negative polarity with respect to the seafloor.

*Figure 3.13 – Anomaly D – PSTM with various AGC operator lengths: (a) No AGC, (b) 1000 ms, (c) 500 ms, (d) 250 ms. The saucer shape of Anomaly D becomes apparent in the 500 ms and 250 ms AGC stacks, a short parallel reflection is apparent beneath the southeastern portion of Anomaly D.*
Anomaly E (Figure 3.14)

Dim zones remain below the upper areas of anomaly E with all AGC windows except for the 250 ms window, which appears to show some reflection continuation in these blank zones.

Figure 3.14 - Anomaly E – PSTM with various AGC operator lengths: (a) No AGC, (b) 1000 ms, (c) 500 ms, (d) 250 ms. Note that the upper part of Anomaly E is stratigraphically controlled
3.4.3.2 Offset-limited PSTM stacks

Offset-limited stacks are similar to producing angle-limited stacks of the data. Three offset range stacks were produced from the PSTM gathers; 0-1 km, 1-2 km, and 2-3 km these have equivalent angle ranges of 0-14°, 14-26° and 26-36° respectively for a reflection at 2 km depth. Limited offset stacks were used to determine whether any of the amplitude anomalies exhibited any strong amplitude with offset (AVO/AVA) effects, which, in turn, could be used to improve lithological interpretation of the data. Angle stacks were also produced with the Hampson-Russell Suite software with angles ranges of 0-5°, 5-10°, 10-15°, 15-20°, 20-25°, 25-30°, 30-35°, 35-40° and 40-45°. However, these stacks are not presented as they did not provide further insight beyond that which is presented here.

Anomaly A (Figure 3.15)

Anomaly A does not show any strong AVO effects in the offset range presented.

Figure 3.15 – Offset-limited PSTM stacks – Anomaly A - (a) full stack, (b) 0-1000 m stack, (c) 1000-2000 m stack, (d) 2000-3000 m stack.
Anomalies B, C and F (Figure 3.16)

In the near offset stacks, 0-1000 m and 1000-2000 m, reflections from anomaly B are relatively chaotic. However, the full stack and particularly the far offset stack, 2000-3000 m, indicate that there is a strong reflection (blue) at the top of this chaotic zone spanning the width of the anticline. This reflection is also visible in the nearer offset ranges, but is usually less continuous. A relatively strong continuous reflection (pink) is well imaged at the central base of the chaotic zone in the 0-1000 m and 1000-2000 m stacks. This basal reflection is also visible in the full stack, however it is much weaker. The chaotic nature of the Anomaly B reflections is probably due to poor stacking of the data as it is a relatively 3-D structure and therefore probably contains out of plane reflections.

In the full offset, 0-1000 m and 2000 m stacks, Anomaly C appears as a 600 m wide negative polarity (peak-trough-peak) reflection with relatively constant amplitude throughout the offset ranges. In the 2000-3000 m stack the reflection character of Anomaly C changes to peak-trough-peak-trough; however, the reflection appears to retain the same wavelength.

Figure 3.16 - Offset-limited PSTM stacks – Anomalies B, C and F - (a) full stack, (b) 0-1000 m stack, (c) 1000-2000 m stack, (d) 2000-3000 m stack. Note the strong continuous reflection (blue) at the top of chaotic zone in Anomaly B which is present in the PSTM full stack and the 2000-3000 m stack, but not in the 0-1000 m or 1000-2000 m stacks; and the strong basal reflection (pink) in the 0-1000 m and 1000-2000 m stacks. Anomaly C retains relatively constant amplitude; however it appears that the polarity of the Anomaly C reflection changes with increased angle from a trough-peak-trough negative reflection to a trough-peak-trough-peak pair of reflections in the 2000-3000 m stack. This may be a tuning effect at Anomaly C, or may potentially be due to poor stacking at high offset ranges.
Anomaly D (Figure 3.17)

In the 0-1000 m stack Anomaly D appears as a closely spaced (~40 ms from peak to peak) pair of high-amplitude positive polarity reflections with a horizontal, but upturned at the edges morphology. These upturned tips do not appear to be processing artefacts and may represent the actual reflector shape. Anomaly D becomes easily definable as a single trough in the 2000-3000 m stack. The reflection is horizontal and distinctly dips shallowly inwards at the edges of the reflection. The Anomaly D reflection terminates at the reverse faults which bound the overlying anticline. The 2000-3000 m stack shows a strong increase in the amplitude of the Anomaly D reflector and the small reflection beneath it compared to other offset ranges. However, the wavelength of the reflection also appears to become larger, potentially indicating that the stack is poor in the far offset due to upturned reflection tails that were not correctly flattened by the NMO correction (perhaps due to the true dip of the structure).

Figure 3.17 - Offset-limited PSTM stacks – Anomaly D - (a) full stack, (b) 0-1000 m stack, (c) 1000-2000 m stack, (d) 2000-3000 m stack. Anomaly D displays a strong increase in amplitude in the 2000-3000 m stack, as well as an apparent decrease in frequency of the reflection. Note the upturned tips of the Anomaly D reflection.
Anomaly E (Figure 3.18)

Both the upper and lower portions of Anomaly E appear to increase in amplitude relative to the background amplitude with increasing offset. The lower area of Anomaly E shows relatively high-amplitude negative-polarity reflections with gentle concave-upwards morphologies at all offsets.

Figure 3.18 - Offset-limited PSTM stacks – Anomaly E - (a) full stack, (b) 0-1000 m stack, (c) 1000-2000 m stack, (d) 2000-3000 m stack. The Anomaly E reflections seem to increase in amplitude relative to the background reflections with offset.

3.4.4 Instantaneous amplitude

Instantaneous amplitude is proportional to the square root of the total energy of the seismic signal at a given sample, and is calculated by taking the square root of the signal squared plus the Hilbert transform of the data squared at a given sample;

$$R(t) = \sqrt{x^2(t) + y^2(t)}, \quad (3.3)(Yilmaz, 2001)$$

where R(t) is the instantaneous amplitude, x(t) is the seismic signal and y(t) is the Hilbert transform of the seismic signal (a Hilbert transform is equivalent to applying a 90° phase shift to the recorded data (Yilmaz, 2001)). Instantaneous amplitude approximates reflectivity in the data, therefore enhances the interpretation of amplitude anomalies (Taner et al., 1979).
Anomalies A, C, and E display very high instantaneous amplitudes in all offset ranges. Anomalies B and D, however, display weaker more diffuse high instantaneous amplitudes which are significantly weaker in the 1000-2000 m offset range. Anomaly E also shows some weakening at the 1000-2000 m offset range. Anomaly F consistently loses amplitude with offset.

Anomaly A (Figure 3.19)
Anomaly A appears to show no significant changes in amplitude with increasing offset. However the amplitude of Anomaly A appears to become slightly more consistent in the 2000-3000 m stack compared to the other stacks and the Anomaly A reflection also becomes more continuous.

Figure 3.19 – Instantaneous amplitude offset-limited stacks – Anomaly A - (a) PSTM; full stack, (b) PSTM; 0-1000 m stack, (c) PSTM; 1000-2000 m stack, (d) PSTM; 2000-3000 m stack. Note the ‘low’ amplitude gap in the Anomaly A reflection that appears to become more continuous in the 2000-3000 m stack.
Anomaly B, C and F (Figure 3.20)

Anomaly C decreases in amplitude from 0-1000 m to 1000-2000 m, but gains amplitude in the 2000-3000 m stack. Anomaly B has scattered high amplitudes in the 0-2000 m offset range, but a strong low-frequency reflection is apparent at the top of anomaly B in the 2000-3000 m offset range. This relatively continuous reflection at the top of Anomaly B is also apparent in the AGC stacks presented previously. The low frequency nature of Anomaly B in the far offset stack may be due poor NMO correction at far offset i.e. non-hyperbolic reflections or out of plane reflections as Anomaly B is likely to be a relatively 3-D structure. Anomaly F loses amplitude with offset throughout the offset range presented.

Figure 3.20 – Instantaneous amplitude offset-limited stacks – Anomaly B, C and F - (a) PSTM; full stack, (b) PSTM; 0-1000 m stack, (c) PSTM; 1000-2000 m stack, (d) PSTM; 2000-3000 m stack. The continuous reflection at the top of Anomaly B, which can be interpreted in the AGC and limited offset stacks, shows significant amplitude in the 2000-3000 m offset range. Anomaly F consistently loses amplitude with offset.
Anomaly D (Figure 3.21)

The Anomaly D reflection appears to increase in amplitude and continuity with offset, and the upturned reflection tips can be clearly defined at the northwestern end of the reflection in the 1000-2000 m offset range.

Figure 3.21 – Instantaneous amplitude offset-limited stacks – Anomaly D - (a) PSTM; full stack, (b) PSTM; 0-1000 m stack, (c) PSTM; 1000-2000 m stack, (d) PSTM; 2000-3000 m stack. The Anomaly D reflection shows high amplitude at 1000-2000 m offset, with distinctly upturned ‘tips’ at the ends.
**Anomaly E (Figure 3.22)**

The lower portion of Anomaly E appears to lose amplitude and continuity with offset. However, the amplitude of the upper portion remains relatively constant with respect to the background amplitude.

![Figure 3.22 - Instantaneous amplitude offset limited stacks – Anomaly E](image)

**Figure 3.22** – Instantaneous amplitude offset limited stacks – Anomaly E - (a) PSTM; full stack, (b) PSTM; 0-1000 m stack, (c) PSTM; 1000-2000 m stack, (4) PSTM; 2000-3000 m stack. The lower portion of Anomaly E loses amplitude with offset.

### 3.5 Seafloor reflection coefficients

Seafloor reflection coefficients (see definition of reflection coefficient in Section 2.1) were calculated using the ratio of peak multiple energy to peak primary energy of the seafloor in a near trace stacked section to which a spherical divergence gain correction had been applied with a constant velocity of 1500 m/s. Since the multiple seafloor reflection ray path should only be through the water column, a constant velocity is appropriate. The spherical divergence gain correction was applied to account for spherical spreading of the source energy, as the path length for the multiple reflection is twice that of the primary reflection. The seafloor and seafloor multiple amplitudes were picked in two ways: (1) horizons representing the seafloor and seafloor multiple were picked and the peak amplitude was extracted from a 5 ms window around the horizons; (2) the same horizons were used, but the peak amplitude was extracted from a 200 ms window around them.
Typical strong seafloor reflection coefficients are around 0.2, hard seafloor reflection coefficients may be as much as 0.3 (Yilmaz, 2001). For the majority of the seafloor in both methods the seafloor has a relatively constant and typical reflection coefficient of ~0.2 (Figure 3.23). The exception is where the seafloor is dipping and multiples are shallow, i.e. the rises around West Norfolk Ridge, because multiples become harder to pick and multiple amplitudes become unreliable due to scattering and non-vertical raypaths. However, the 200 ms method has a very high reflection coefficient associated with Anomaly A, which has an approximate average value of 0.5 (up to ~0.7), although the 5 ms method only shows a coefficient of ~0.2. This is because Anomaly A is less than 100 ms below the seafloor, and therefore is included in the 200 ms window. If we assume a reflection coefficient of 0.5 for a seawater/Anomaly A reflection (Anomaly A is very shallow at ~50 ms and probably covered with young water-saturated sediments, so this is a reasonable approximation) and a seawater impedance of 1552.5 g/cm³·m/s (V_p = 1500 m/s, ρ = 1.03 g/cm³), the impedance of Anomaly A will be approximately 4660 g/cm³·m/s. If the reflection coefficient is ~0.7 then the approximate Anomaly A impedance will be 8800 g/cm³·m/s. These impedance values are very high for what is likely for unconsolidated water-saturated sediments.

**Figure 3.23** – Reflection Coefficients for the entire STRATUS-2D-002 seismic line (overlaid on PSTM stack) – Reflection coefficients were calculated using two methods, (1 – blue) the peak amplitude was selected from a 5 ms window centred on picked seafloor and multiple horizons and (2 – red) the peak amplitude was selected from a 200 ms window centred on the same picked seafloor and multiple horizons. For the majority of the seafloor in both methods the seafloor has a relatively constant and typical reflection coefficient of ~0.2 except where the seafloor is dipping and multiples are shallow i.e. the rises around West Norfolk Ridge. However, the 200 ms method calculates a very high reflection coefficient associated with Anomaly A, which has an approximate average value of 0.5 (up to ~0.7), although the 5 ms method only indicates a coefficient of ~0.2. This is because Anomaly A is less than 100 ms below the seafloor and has an amplitude that is far higher than the seafloor reflection.

### 3.6 Anomaly interpretations

Six major amplitude anomalies were identified in STRATUS-2D-002, Anomalies A-F. These have been divided in to 5 distinct types: (1) laterally extensive thin (only one wavelet) reflections with very high positive amplitude and strong dimming below (Anomaly A); (2) concave-upwards chaotic to continuous high amplitude reflections with associated anticlines above and velocity pull ups below (Anomalies B and D); (3) very high amplitude reflections
with negative polarity relative to the seafloor (Anomaly C); (4) short chaotic high-amplitude shallowly dipping reflections with associated strong attenuation below (Anomaly E); and (5) moderate to low amplitude negative polarity reflections which may be patchy in places (Anomaly F).

<table>
<thead>
<tr>
<th>Anomaly Type</th>
<th>A</th>
<th>B and D</th>
<th>C</th>
<th>E</th>
<th>F</th>
</tr>
</thead>
<tbody>
<tr>
<td>Anomaly A</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Amplitude</td>
<td>Very high</td>
<td>High</td>
<td>Very high</td>
<td>Moderate to high</td>
<td>Moderate to low</td>
</tr>
<tr>
<td>Apparent Polarity</td>
<td>Positive</td>
<td>Negative</td>
<td>Negative</td>
<td>Negative??</td>
<td>Negative</td>
</tr>
<tr>
<td>Velocity Pull-up</td>
<td>No</td>
<td>Yes ~200 ms</td>
<td>No</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>Structural Association</td>
<td>No</td>
<td>Domed fold above, fault bounded</td>
<td>Association with NW Anom. B fault</td>
<td>Small faults</td>
<td>SE Anom. B bounding fault</td>
</tr>
<tr>
<td>Stratigraphically controlled</td>
<td>yes</td>
<td>no</td>
<td>yes</td>
<td>Erosion surface horizon</td>
<td>Erosion surface horizon</td>
</tr>
<tr>
<td>Coherent</td>
<td>Yes</td>
<td>Chaotic to partially</td>
<td>Yes</td>
<td>Partially</td>
<td>Becomes patchy</td>
</tr>
<tr>
<td>Extent</td>
<td>~6 km</td>
<td>~2.5 km and ~3 km respectively</td>
<td>~600 m</td>
<td>Various over ~3 km</td>
<td>~2.9 km (7.1 total)</td>
</tr>
<tr>
<td>Anomaly Type</td>
<td>1</td>
<td>2</td>
<td>3</td>
<td>4</td>
<td>5</td>
</tr>
</tbody>
</table>

**Anomaly A**

Anomaly A is a very shallow very high-amplitude positive reflection that occurs ~50 ms below the seafloor to the NE of Anomaly B, and extends for approximately 6 km. Reflection coefficient calculations indicate that the anomaly has an impedance between 4660 to 8800 g/cc.m/s. Strong amplitude dimming is apparent below Anomaly A. Whilst the reflection coefficient impedances are relatively low, we suggest that the very high amplitude, positive polarity, broad extent and close association with Anomaly B and C indicate that Anomaly A is probably igneous in origin, either a thin sill or flow.

**Anomaly B and D**

Anomalies B and D, with their high-amplitude negative reflections and overlying broad anticlines appear to be very similar to igneous intrusions, 2-4 km in diameter, reported from 3-D seismic data in the Rockall Trough by Hansen and Cartwright (2006) (see Figure 2.3). Hansen and Cartwright (2006) infer that these concave upwards high amplitude reflections imaged beneath dome shaped anticlines are saucer-shaped sills, which have forced up the overlying strata during intrusion forming the overlying anticlines. The high degree of diffraction from Anomaly B suggests that the Anomaly B reflector is irregular or discontinuous.
A similar feature to Anomaly B and D was imaged on the AGSO 114-04 seismic line which crosses STRATUS-2D-002 near to perpendicular at the southeastern end of anomaly D. The feature on AGSO 114-04 occurs approximately 1.3 km SW (however, the accuracy of the geometry is unknown) of Anomaly D, which was originally interpreted by Herzer et al. (1997) to be a compressional pop-up (Figure 6a in Herzer et al. (1997)). There is strong similarity between Anomaly D and the feature on AGSO 114-04, which shows shallow inward-dipping high-amplitude negative-polarity reflections that display a distinct symmetrical ‘saucer’ shape approximately ~5 km in lateral extent with a gentle dome-shaped anticline overlying it which only extends as wide as the high amplitude reflections (Figure 3.24). This suggests that the feature on AGSO 114-04 and Anomaly D may be similar features, or the same feature, although determination of the accuracy of the 114-04 navigation would be required to confirm this.

The similarity of Anomalies B and D (and the feature on AGSO 114-04) in size (3-4 km) and reflection character (negative polarity, saucer shaped) to the saucer-shaped intrusions of Hansen and Cartwright (2006) and the close proximity of Anomaly B to the Pliocene basalt sample GO353, leads us to infer that Anomalies B and D represent igneous saucer-shaped sills, that have intruded at shallow depths (~500 m? Below the seafloor) resulting in the formation of overlying anticlines which deformed the paleoseafloor.

This igneous intrusion interpretation is reinforced by the presence of velocity pull-ups beneath Anomalies B and D, indicating that the velocities are higher than expected, i.e. they are causing reflections to occur earlier than expected beneath these anomalies. Several isolated anomalously high velocity reflections were encountered during velocity analysis below Anomaly B (stacking velocities of ~2000 m/s compared to ~1900 m/s for the same reflection, ~200 m apart), these only occurred where ray paths are likely to have travelled through Anomaly B.
Anomaly C

In several of the post stack processing schemes applied to STRATUS-2D-002, in particular the full offset stack, 2000-3000 m stack and the 1000 ms AGC stack, it appears that Anomaly C may be a continuation of Anomaly B. However, this interpretation raises two issues; (1) Anomaly C in general appears to be more consistent and have higher amplitude and frequency than Anomaly B suggesting a difference in the reflector causing it and (2) the location of Anomaly C with respect to the NW Anomaly B bounding fault.

If Anomaly B is a saucer-shaped sill as inferred above, and Anomaly C is a continuation of that intrusion, the change in reflection character between anomalies B and C to higher amplitude and apparently higher frequency may be a tuning effect. Tuning effects will occur when a reflector is approaching its detection limit thickness and the upper and lower surface reflections will constructively (and destructively) interfere (Sheriff, 2002). If anomalies B and C are caused by the same intrusion then the anomaly C reflector must be closer to its detection limit than Anomaly B, resulting in a tuned high-amplitude, relatively high-frequency reflection.

Anomaly C occurs directly northwest of the NW Anomaly B bounding fault, on the downthrown side of the fault, but occurs stratigraphically higher than Anomaly B and appears continuous. This suggests that either, Anomaly C is unrelated to Anomaly B and the apparent continuity is coincidental, or that intrusion causing anomalies B and C must have occurred
syntectonically with the folding of the anticline overlying Anomaly B and slip occurring on the NW bounding fault.

Anomalies B and C are most likely to be caused by the same intruded igneous body due to the continuous nature of the anomalies across the NW Anomaly B bounding fault, and that the higher amplitude and frequency of Anomaly C is caused by a tuning effect. This suggests that Anomaly C is the thin outer rim of a saucer-shaped igneous intrusion and Anomaly B represents the thicker flatter inner sill (e.g. Thomson and Hutton, 2004 and Polteau et al., 2008).

**Anomaly E**
The occurrence of Anomaly E (in particular; the upper part of the anomaly) is stratigraphically controlled by an erosional surface above the UB9 onlap surface. This surface erodes the crest of the Anomaly D anticline, and in the location of Anomaly E is very shallow at 30-90 ms below the seafloor. Anomaly E occurs below this surface with an apparent negative polarity (with respect to the seafloor), and has strong associated amplitude dimming below. In the 250 ms window AGC, reflections appear to be continuous through this blanked zone. The high degree of diffraction from the upper part of Anomaly E suggests that the upper Anomaly E reflector is irregular or discontinuous.

The lack of disturbance and lack of velocity pull-up below Anomaly E, strong amplitude dimming, probable negative polarity and stratigraphic control by an erosion surface suggest that Anomaly E is caused by a free gas accumulation at the erosion surface mentioned above, trapped by the overlying sediment drape.

**Anomaly F**
Anomaly F shows strong structural and stratigraphic associations. Anomaly F is stratigraphically controlled updip along the erosion surface horizon above the UB9 surface mentioned in the previous section, which erodes the anticline above anomaly D and onlaps onto the anticline above Anomaly B and is not cut by the faults that bound Anomaly B. Anomalous negative amplitudes start above the SE Anomaly B bounding fault, losing amplitude to the southeast, and becoming patchy further updip. The moderate negative polarity of Anomaly F suggests that it may be caused by the presence of free gas migrating updip along the erosion surface horizon. This interpretation of stratigraphically controlled free
gas is reinforced by a consistent amplitude decrease in the anomaly updip and the patchy nature of the anomaly at its southeastern extent. The association of higher amplitudes at the down dip end of Anomaly F, near the SE bounding fault of Anomaly B, suggests that the fault may be facilitating gas migration from depth which is subsequently trapped at the erosion surface forming Anomaly F.
Chapter 4

Seismic waveform inversion

4.1 Introduction

Six distinct amplitude anomalies were identified in STRATUS-2D-002, anomalies A to F (see Chapter 3, in particular Table 3.2). These anomalies range from high-amplitude, continuous, positive-polarity reflections near the seafloor (Anomaly A) to fault-bounded, negative-polarity reflections beneath broad anticlines (Anomalies B and D). Whilst some inference of the lithologies causing these anomalies has been made through seismic interpretation and comparison to similar physically sampled seismic features in the literature, a more rigorous approach is to apply seismic inversion techniques in an attempt to directly extract elastic rock properties from the data.

In this chapter, an industry standard inversion software package, CGGVeritas’ Hampson-Russell Suite software, is used to jointly invert for P-wave impedance ($Z_p$), S-wave impedance ($Z_s$) and density ($\rho$) from time migrated CMP gathers in an area with no well control to attempt to determine the lithologies causing the amplitude anomalies identified on the STRATUS-2D-002 seismic line.

4.2 Inverse theory

Seismic inversion is the process of extracting information about elastic rock properties from seismic data, based on the travel-time, amplitude and phase information contained within a seismogram (Sheriff, 2002). Inversion of geophysical data in general can be defined as a method for producing a geological model that adequately fits the observed data (Treitel and Lines, 2001). All inversion methods are non-unique, and there may be an infinite number of models that fit the observed data. Therefore a good initial low frequency ‘guess’ model must be used to avoid local minima in the inversion mis-fit function, to obtain the most realistic model (Virieux and Operto, 2009). Whilst difficult in practice, in general the more observed parameters that we are trying to fit with inversion, the more likely we are to obtain a realistic
geological model because the inverted model must be able to adequately fit all of the observed parameters. For instance if we were to carry out joint gravity and P-wave travel-time inversion on a crustal data set, the inverted geological model must fit both the observed gravity data as well as the measured travel-times, which will exclude many of the possible models that would fit either the observed gravity or travel-time data individually but do not fit both data sets (Lelièvre et al., 2012). The amount of \textit{a priori} knowledge available to guide the inversion is generally key to obtaining a geologically ‘real’ and reasonable model; i.e., if we have bore hole data or core, we may already have a good indication of rock types and their P-wave velocities in the areas of interest allowing us to start with a realistic model.

A simplistic view of seismic waveform inversion is that we are trying to remove the source wavelet from the recorded seismogram to leave us with a reflectivity series of the subsurface, in essence trying to do the inverse of the seismic convolution model below:

\[
\text{Earth Reflectivity } \ast \text{ Wavelet + Noise} = \text{Seismic Trace}
\]

where \(\ast\) indicates the convolution operator. Instead of the final result being the recorded data (seismic trace) we want to instead use the recorded data and remove the effect of the input wavelet to quantitatively say something about the reflectivity of the earth.

At first glance it would appear that this is a relatively trivial problem, that we could simply remove the wavelet from the seismic trace by convolving the seismic trace with the inverse of the wavelet to extract the reflectivity series (± some noise). However, in practice it isn’t that simple for several reasons; (1) the source wavelet is often not accurately known, (2) noise is random and falls well within the range of recorded seismic frequencies (5-200 Hz), and may have amplitudes similar to or even greater than primary signal, and (3) the seismogram is band-limited due to band-limited data collection (from both the source and receivers), therefore numerous models may be able to fit the data as low frequency impedance trends are not available in the data to guide the inversion. For some inversion methods e.g. travel-time inversion some of these considerations are not an issue, but those methods can only invert for velocity information, and provide no information on the other elastic properties, such as impedance, Poisson’s ratio and density. In theory prestack waveform inversion (using phase, amplitude and travel time) should be able to accurately extract any elastic property of the subsurface. However, in reality we are limited by; the lack of low frequency data, random noise, computational power, the ability to obtain true amplitude recovery in data processing, and the ability to obtain an accurate estimation of the source wavelet.
Raytracing travel-time inversion, such as Zelt and Smith’s (1992) RAYINVR uses picked travel times for particular reflections and refractions and an initial layered model, and applies a least-squares inversion to minimise the misfit between the modelled travel-times and the picked travel times to give P-wave velocity information (if S-wave phases are present then S-wave inversion can also be undertaken separately). Ray-tracing inversion methods can provide information about the velocity structure but do not give information on other rock properties, such as impedance or density. Ray-tracing methods are generally most useful for large-scale problems such as crustal refraction surveys, as all phases that are inverted for travel time must be defined in the starting model. Although in practice only large-scale velocity structure can be easily obtained using ray-tracing methods, they do provide the ability to ignore noise in the data (unless reflections are entirely obscured), by defining and picking travel time curves for the reflections that are used for inversion. However, the picking of phases for travel-time inversion may be quite subjective, especially with noisy data, as the user has to decide what phase each arrival represents.

Waveform inversion, either poststack or prestack, seeks to produce a subsurface model that when used to produce synthetic seismic traces, will adequately match preserved amplitude processed field data (Warner et al., 2013). In theory, waveform inversion should be able to produce a subsurface model of any physical property that affects seismic energy propagation in the subsurface; i.e. V_p, V_s, and ρ (see Chapter 2 – Sections 2.2.1 and 2.2.2) (Warner et al., 2013). Generally, due to the high model resolution required (the modelled data must have sufficient resolution to match the real data), only P-wave velocity or impedance is inverted for in all but the simplest cases, due to the high computational requirements (Warner et al., 2013). The majority of current waveform inversion methods use a damped least-squares optimisation method to minimise the mis-fit between the real and synthetic data (Virieux and Operto, 2009). Without low frequency (<1 Hz) variation, these waveform inversion methods do not reliably obtain accurate results especially at high frequencies (Virieux and Operto, 2009). Therefore it is critical to accurately define a background impedance model for any waveform inversion method.

Another significant issue affecting successful waveform inversion is data conditioning prior to inversion. Input data need to be processed to obtain true relative amplitude, i.e., attenuation must be accounted for, and high signal-to-noise ratios need to obtained, for which some standard processing methods may be unable to be applied (e.g., f-k filtering) as they will effect relative amplitudes.
Poststack waveform inversion methods can only account for one component of the wavefield at a time, i.e., data can be inverted to obtain P-wave impedances, but not account for S-wave conversion, and vice versa. However, the ability of the stack to attenuate random noise enables high signal-to-noise ratios to be obtained, and the relatively small size of data volumes compared to prestack data make poststack inversion computationally simpler. However, stacking also attenuates any amplitude variation with offset (AVO) information that may be present in the prestack data, so wavetype conversions will not be accounted for, and therefore only one elastic parameter can be inverted for at a time.

Prestack inversion methods, on the other hand, retain this AVO information and inversion methods based on the linearised approximations of the Zoeppritz equations, such as the Aki and Richards approximation, can be used to jointly extract multiple rock properties, such as P- and S-wave impedance, density and porosity (Simmons and Backus, 1996). However, the major disadvantage of these methods is the several order of magnitude increase in data involved in the computation (and therefore processing time), as well as using a processing scheme that ensures relative amplitudes are maintained at all offsets and, determining how to correctly weight the effect of the different elastic parameters that are being inverted for, to ensure that the global misfit minima is reached representing a ‘real’ subsurface model (Virieux and Operto, 2009).

4.3 Hampson-Russell Suite inversion

Full-waveform inversion was carried out on STRATUS-2D-002 using the commercial Hampson-Russell Suite inversion software, which is marketed by CGGVeritas. Both model-based poststack acoustic impedance inversion and joint prestack inversion were undertaken using the Hampson-Russell Suite software.

4.3.1 Hampson-Russell prestack inversion

Prestack inversion for P-wave and S-wave impedance and density, was carried out using the Hampson-Russell Suite software. Three main pieces of information are required, (1) a prestack seismic volume consisting of NMO-corrected CMP gathers, (2) velocity and density control in the form of a well log to define a background earth model and (3) a set of picked horizons to guide the initial model interpolation. As the Reinga Basin has no direct well
control, several methods were employed to produce synthetic ‘wells’ for velocity control, such as ray-tracing forward modelling and stacking velocity pseudo-wells.

Unlike poststack inversion, which ignores the assumption that wet clastic rocks should show a linear relationship between $V_p$ and $V_s$ (i.e. Castagna’s relationship), and that there should be some relationship between $V_p$ and density (i.e. Gardner’s relationship), joint prestack inversion provides some coupling between these related variables (Gardner et al., 1974; Castagna et al., 1985; Hampson-Russell, 2007). Hampson-Russell prestack inversion applies a modified linearised version of the Zoeppritz equations, the ‘simultaneous inversion equation’, which is a modification of the Fatti modification of the Aki-Richards Equation (Fatti et al., 1994; Simmons and Backus, 1996; Aki and Richards, 2002; Hampson et al., 2005; Hampson-Russell, 2007). For derivations and detailed information on the algorithms applied see (Fatti et al., 1994; Simmons and Backus, 1996; Buland and Omre, 2003; Hampson et al., 2005; Hampson-Russell, 2012)

Linearised Zoeppritz equations, such as the modified Aki-Richards equation relate elastic properties across reflecting boundaries using the ratio of the difference across the boundary to the average properties in the layers adjacent to the boundary, i.e. $\frac{\Delta \rho}{\rho}$, $\frac{\Delta V_p}{V_p}$ and $\frac{\Delta V_s}{V_s}$ where $\Delta \rho, \Delta V_p$ and $\Delta V_s$ are the difference in density, P-wave velocity and S-wave velocity across the reflecting boundary, respectively, and $\rho, V_p$ and $V_s$ are the average density, P-wave velocity and S-wave velocity of those layers (Simmons and Backus, 1996). The Aki-Richards equation and its derivatives are reasonably accurate where $\frac{\Delta \rho}{\rho}$ and $\frac{\Delta V}{V}$ terms are less than $\sim 0.1$ (Hampson-Russell, 2007).

The small reflectivity approximation assumed in the Hampson-Russell inversion algorithm is:

$$R_{pi} = \frac{Z_{pi+1} - Z_{pi}}{Z_{pi+1} + Z_{pi}} \approx \frac{\Delta Z_{pi}}{2Z_{pi}},$$

where $i$ is the $i^{th}$ interface between layer $i$ and $i+1$, this approximation holds when reflection coefficients are less than 0.1 (Hampson et al., 2005; Hampson-Russell, 2012).

The ‘simultaneous inversion equation’ applied in the Hampson-Russell prestack inversion routine can be expressed as:

$$S_\theta = \tilde{c}_1 W_\theta D L_p + \tilde{c}_2 W_\theta D \Delta L_s + W_e c_3 D \Delta L_D,$$  \hspace{2cm} (4.1) \hspace{1cm} \text{(Hampson et al., 2005)}

where $\tilde{c}_1 = \frac{1}{2} c_1 + \frac{1}{2} k c_2 + m c_3$ and $\tilde{c}_2 = \frac{1}{2} c_2$, and where $c_1 = 1 + tan^2 \theta,$

$$c_2 = -8 \left( \frac{V_s}{V_p} \right)^2 \tan^2 \theta \text{ and } c_3 = 0.5 \tan^2 \theta - 2 \left( \frac{V_s}{V_p} \right)^2 \sin^2 \theta; S_\theta \text{ is the trace reflectivity, } W_\theta \text{ is}$$
the angle dependant wavelet, \( D \) is the derivative matrix and \( L_P, L_S \) and \( L_D \) are the logs of the P-impedance, S-impedance and density vectors, respectively.

It is an unstable and memory intensive process to solve these equations using matrix inversion methods. Therefore, an iterative approach is applied using the conjugate gradient method starting from the initial P-wave impedance model (Hampson et al., 2005).

### 4.3.2 Hampson–Russell model-based poststack inversion

Poststack acoustic impedance (P-wave impedance) inversion was carried out using the Hampson-Russell Suite software, to be used as a benchmark for determining the reliability of subsequent prestack inversions. The background P-wave impedance model was derived from interval velocities for the entire line and the application of Gardner’s relationship (to estimate corresponding densities) producing a background impedance section for the entire STRATUS-2D-002 line.

Poststack inversion works to minimise the misfit for each seismic trace between the real data and synthetic data produced from the impedance model. A constrained model-based inversion method was used, which sets hard constraints on the amount in which the P-wave impedance for a given sample can change from the initial model, following the relationship below:

\[
I(i) = I_0(i) \pm X I_{AV},
\]

(Hampson-Russell, 2012)

where \( I(i) \) is the final impedance at sample \( i \), \( I_0(i) \) is the original estimated impedance at sample \( i \), \( I_{AV} \) is the average impedance of the input constraint, \( I_0 \). \( X \) is the maximum allowable percentage change. The Hampson-Russell model-based inversion algorithm is iterative and only allows gradual changes in impedance from the original impedance model during each iteration to prevent noise driving the inversion to a local minimum (Hampson-Russell, 2012).

### 4.4 Background models and synthetic well log generation

All full waveform seismic inversion schemes require a background earth model to provide low frequency information to stabilise the inversion, and to enable correct scaling of the synthetic data (Virieux and Operto, 2009). Ideally the background model will contain real low frequency information measured independently (e.g., from well data) for each of the parameters which we are inverting. For instance, this might be done by taking the low frequency component of wireline sonic and gamma density by taking a large window running
average of the data to obtain relatively robust background P-wave and S-wave velocity, and density profiles. However, in the Reinga Basin, there is no nearby well control for the STRATUS-2D-002 line. The only well that ties to the data is located over 500 km to the southeast in the Northland Basin, the Waka Nui-1 well, which lacks post-drilling wireline data (Stagpoole, 2011). No sonic logging or even a checkshot survey was undertaken on Waka Nui-1; therefore other methods need to be applied to create background impedance models for the STRATUS-2D-002 line.

The only elastic parameter for which low frequency information can be reasonably obtained directly from multi-channel seismic data is P-wave velocity in the form of stacking velocities and their associated interval velocities. S-wave velocity and density must be inferred from the P-wave velocity data using established global empirical relationships, such as the Castagna and Gardner relationships, which calculate S-wave velocity and density respectively from P-wave velocities (Gardner et al., 1974; Castagna et al., 1985).

All background models produced use the Castagna and Gardner relationships defined below to calculate S-wave velocity and density from P-wave velocities. S-wave velocity and density were held constant in the water column in all background models.

Castagna et al. (1985) defined a relationship between P- and S-wave velocities for siliciclastic water-saturated mud rocks, based on in-situ sonic and seismic field measurements from various studies. The relationship is defined as:

$$V_p = 1.16V_s + 1.36,$$  \hspace{1cm} (4.3)

where $V_p$ and $V_s$ are the P-wave and S-wave velocities in km/s.

Gardner et al. (1974) define the following relationship between P-wave velocity and density as an average relationship from a large number of laboratory and field measurements in saturated sedimentary rocks:

$$\rho = 0.31V_p^{0.25},$$  \hspace{1cm} (4.4)

where $\rho$ is density in g/cm$^3$, and $V_p$ is P-wave velocity in km/s.

Three methods were used to define low frequency P-wave velocity background models from which impedance and density models were calculated: (1) stacking and interval velocities for the entire line, (2) pseudo-well logs derived from individual CMP stacking and interval velocities, and (3) pseudo-well logs derived by shot raytracing using the established
RAYINVR travel-time inversion code (Zelt and Smith, 1992). Background impedance models were calculated in the Hampson-Russell Suite as the product of velocity and density models.

4.4.1 Interval velocity background model

Stacking velocities used to correct for normal moveout during seismic processing provide an RMS velocity model for the entire seismic line with control points located along the strongest reflections. RMS stacking velocities were extracted for every CMP from the Claritas.nmo velocity file created during processing. Interval velocity profiles were then calculated for each CMP using the Dix Equation (Sheriff, 2002):

\[ V_{\text{int}}(n) = \sqrt{\frac{V^2 t_n - V_{n-1}^2 t_{n-1}}{t_n - t_{n-1}}} \]

(4.5)

where \( V_{\text{int}}(n) \) is the interval velocity for a layer defined as lying between two control points in the velocity model, \( V \) is the RMS velocity to the reflector at the base of a given layer, and \( t \) is the two-way travel time to the reflector.

Interval S-wave velocity and density models were calculated from the interval P-wave velocity model using the Castagna and Gardner relationships (see Section 4.3). These velocity and density models were then used to create background P- and S-wave impedance models for the inversion. Background models created using interval velocities for the entire seismic line, typically show smoother impedance changes across the model.

4.4.2 Interval velocity pseudo-well background model

P-wave RMS velocity profiles for individual CMP gathers in desired pseudo-well locations were extracted manually from the Claritas.nmo velocity file and converted to interval velocities using the Dix equation (Sheriff, 2002). Corresponding S-wave velocity and density logs were calculated using the empirical relationships of Castagna et al. (1985) and Gardner et al. (1974) respectively.

Subsequently, P-wave and S-wave impedance logs were calculated from the velocity and density logs. Impedances from the pseudo-well were extrapolated across the entire STRATUS-2D-002 guided by several hand-picked horizons including the seafloor. Background models created using this horizon guided method cause impedances across the
profile to be strongly controlled by the pseudo-well and guiding horizons used, and impedances will tend to remain constant within a layer.

4.4.3 RAYINVR pseudo-well background model
Low frequency 1-D velocity models were calculated via ray-tracing of individual shot gathers using the Zelt and Smith (1992) RAYINVR Travel-Time inversion code. Travel times for five phases (the seafloor plus four other reflections) were picked using the interactive GUI ZPLOT. Ray-traced travel times were iteratively forward modelled to match the picked travel times, rather than carrying out least-squares inversion due to difficulties with obtaining a stable inversion with a 1-D velocity model (Figure 4.1). All layers in the velocity model were assumed to have no lateral or vertical velocity gradients. As before, S-wave velocity and density were calculated using the empirical relationships of Castagna et al. (1985) and Gardner et al. (1974), respectively.

Subsequently, P-wave and S-wave impedance logs were calculated from the velocity and density logs. Again, impedances from the pseudo-well were extrapolated across the entire STRATUS-2D-002 guided by hand-picked horizons, including the seafloor, which in this case correspond to the reflections analysed in RAYINVR. Background models created using this horizon guided method cause interval velocities across the profile to be strongly controlled by the pseudo-well and guiding horizons used, and impedances will tend to remain constant within a layer.

4.5 Hampson-Russell Suite data pre-processing
Seismic data were input to the Hampson-Russell Suite as NMO-corrected, time-migrated CMP gathers. As all areas of interest were contained within the top 4500 ms of data, the total trace length was limited to 4500 ms to avoid the seafloor multiple reflection.

Prior to inversion, CMP gathers were supergathered using a 5-CMP sliding window. A radon-domain de-noise filter was applied to minimise any unflattened arrivals and attenuate random noise. After radon filtering, a broad bandpass filter was applied with 5/10/70/90 Hz corner frequencies, and the filtered CMP supergathers were transformed into the angle domain using an interval velocity model derived from processing stacking velocities for the STRATUS-2D-002 line. Angle gathers with a maximum angle of 50° were produced with various binning parameters used for different prestack inversion runs. Inversion runs using angle ranges of 0-
45° binned into 5 bins, 0-50° binned into 20 bins, and 0-50° into 40 bins are presented below. Data destined for poststack inversion were stacked using all offsets from the filtered CMP gathers to provide a single poststack volume.

**Figure 4.1** – RAYINVR ray-tracing plot for Shot 11401 (equivalent to CMP location 17480). The upper plot shows the ray tracing P-wave velocity layer boundaries in black and the coloured lines show traced ray paths for particular arrivals. The lower plot shows the corresponding calculated travel-time curves in black for each layer. The coloured ticks correspond to the picked travel-times for each arrival from Shot 11401. These show a good fit to the calculated arrivals.
4.6 Inversion results

A representative subset of six inversion runs is presented in this section from a total of approximately 50 inversion runs undertaken. One of these is a model-based poststack inversion, and the other five are prestack inversions with different starting models and inversion parameters. The six examples shown here provide a broad overview of all of the different inversion runs that were computed.

All inversions use 150 ms window length wavelets, which were statistically estimated from the data. For prestack inversions, three angle-dependent wavelets were statistically estimated from the data. These cover near, middle, and far angle ranges, and help to account for decreasing frequencies at far-offsets caused by NMO stretch during NMO correction.

Figures comparing the starting P-wave impedance models to inverted P-wave impedances and the poststack impedance inversion for the entire STRATUS-2D-002 line are presented as Figures 4.2 to 4.6. The inverted poststack P-wave impedances are shown as part a of Figures 4.7 to 4.10 so direct comparisons can be made between all different inversion runs.

The colour scale of all P-wave impedance figures is clipped between 1500 and 10 000 g/cm³·m/s to ensure that detail can be seen in areas of non-anomalous reflections in all inversion runs. The measured impedance values reported in the text are not affected by this clipping.

Only P-wave impedance results are presented, as they can be directly compared to the poststack inversion results. S-wave impedance and density inversion results tended to follow the P-wave impedance. This is not unexpected, as empirical relationships such as the Castagna’s and Gardner’s relationships indicate that there should be some relationship between these parameters. Also there are no independently derived S-wave impedance and density starting models, only models that were derived from P-wave velocity using Castagna and Gardner’s relationships respectively.
Figure 4.2  (a) Interval model - 5 bin prestack inversion. (top) - Starting P-wave impedance model calculated from stacking velocities for the entire STRATUS-2D-002 line. (bottom) - Inverted P-wave impedance result for the entire STRATUS-2D-002 line. Note that the background model also shows increased P-wave impedance around Anomaly B.

(b) Model-based poststack inversion. (top) - Starting P-wave impedance model calculated from stacking velocities for the entire STRATUS-2D-002 line. (bottom) - Inverted P-wave impedance result for the entire STRATUS-2D-002 line. Note that the background model also shows increased P-wave impedance around Anomaly B.
Figure 4.3 - (a) Interval model prestack inversion. (top) - Starting P-wave impedance model calculated from stacking velocities for the entire STRATUS-2D-002 line. (bottom) - Inverted P-wave impedance result for the entire STRATUS-2D-002 line. Note the smooth impedance variation throughout the line.

(b) CMP 18000 model prestack inversion. (top) - Starting P-wave impedance model extrapolated from P-wave impedances from a pseudo-well located at CMP 18000. (bottom) - Inverted P-wave impedance result for the entire STRATUS-2D-002 line. Note the sharp changes in background impedance, and the low impedance zone that has been extrapolated across the entire line.
Figure 4.4 - (a) Interval model prestack inversion. (top) - Starting P-wave impedance model calculated from stacking velocities for the entire STRATUS-2D-002 line. (bottom) - Inverted P-wave impedance result for the entire STRATUS-2D-002 line. Note the background model also shows increased P-wave impedance around Anomaly B. (b) Model-based poststack inversion. (top) - Starting P-wave impedance model calculated from stacking velocities for the entire STRATUS-2D-002 line. (bottom) - Inverted P-wave impedance result for the entire STRATUS-2D-002 line. Note the background model also shows increased P-wave impedance around Anomaly B.
Figure 4.5 - (a) Interval model - covariance=0.5 prestack inversion. (top) - Starting P-wave impedance model calculated from stacking velocities for the entire STRATUS-2D-002 line. (bottom) - Inverted P-wave impedance result for the entire STRATUS-2D-002 line. Note the high similarity between this run and the poststack inversion.

(b) Model-based poststack inversion. (top) - Starting P-wave impedance model calculated from stacking velocities for the entire STRATUS-2D-002 line. (bottom) - Inverted P-wave impedance result for the entire STRATUS-2D-002 line. Note that the background model also shows increased P-wave impedance around Anomaly B.
Figure 4.6: (a) RAYINVR model prestack inversion. (top) Starting P-wave impedance model extrapolated from P-wave impedances calculated from seismic velocities for the entire STRATUS-2D-002 line. Inverted P-wave impedance result for the entire STRATUS-2D-002 line. Note significantly lower overall impedance and smooth variation throughout the line.

(b) Model-based poststack inversion. (top) Starting P-wave impedance model calculated from stacking velocities for the entire STRATUS-2D-002 line. Inverted P-wave impedance result for the entire STRATUS-2D-002 line. Note increased P-wave impedance around Anomaly B.
4.6.1 Model-based poststack inversion (acoustic impedance inversion)

Model-based poststack acoustic inversion was carried out using a starting P-wave impedance model based on processing interval velocities, and densities calculated using the Gardner (1974) relationship (see Section 4.4.1). Poststack inverted P-wave impedances can be used as a benchmark for determining whether P-wave impedances calculated from the different prestack inversions were realistic.

**Inversion Parameters:**

- **Starting model:** P-wave impedance model calculated from stacking velocities using Gardner’s equation (Section 4.4.1).
- **Method:** Hard-constraint, model-based (see Section 4.3.2)
- **Iterations:** 50
- **Block size:** 2ms
- **Prewhitening:** 1%
- **Updating:** Zp
- **Constraints:** lower constraint, 30%; upper constraint, 90%

**Anomaly A (Figure 4.7)**

Anomaly A is characterised by consistent P-wave impedances of up to ~6500 g/cm³·m/s. Low impedance areas of ~2300 g/cm³·m/s occur above and below Anomaly A. These low impedance areas are probably artifacts, perhaps indicating that the Anomaly A reflection is close to its detection limit and shows some tuning effects.

**Anomalies B and C (Figure 4.8)**

Anomaly B reaches P-wave impedance values of up to 9500 g/cm³·m/s especially in the northwestern part of the anomaly. However, the majority of Anomaly B is relatively close to background impedances and is relatively incoherent.

Anomaly C is characterised by a high-impedance over low-impedance pair, with the upper section reaching impedances of 8500 g/cm³·m/s. The lower section has an impedance of 4300 g/cm³·m/s, which is slightly lower than the background impedance. The vertical lines
are artifacts introduced by the trace-by-trace application of the poststack inversion (see Figure 4.8).

**Anomaly D (Figure 4.9)**
The Anomaly D has relatively consistent P-wave impedance along the anomaly up to ~9080 g/cm³·m/s.

**Anomaly E (Figure 4.10)**
The lower area of Anomaly E reflections has P-wave impedance values up to 8500 g/cm³·m/s, whereas the upper area has impedances of 5300 g/cm³·m/s, which are slightly higher than the background impedance.

**Anomaly F (Figure 4.8)**
The Anomaly F reflection consists of a high impedance zone of ~4100 g/cm³·m/s over a low impedance zone at 2600 g/cm³·m/s. The background impedance around Anomaly F is ~3300 g/cm³·m/s.

**4.6.2 Interval model – 5-angle bins**
Prestack joint inversion for Zp, Zs, and density, using a background impedance model calculated from stacking velocities for the entire STRATUS-2D-002 line. An angle range of 0-45° was used, separated into five bins. This method may result in a more stable inversion as the signal-to-noise ratios should be higher due to stacking out of random noise in the broad angle bins used.

**Inversion Parameters:**
- **Starting Model:** Zp, Zs, and density starting models calculated from stacking velocities using Castagna’s and Gardner’s equations.
- **Gathers:** 5 bins, 0-45°
- **Method:** Prewhitening,
- **Iterations:** 50
- **Background Vs/Vp (Gamma):** 0.5
- **Zp prewhitening:** 2%, **Zs prewhitening:** 2%, **Density prewhitening:** 5%
- **Updating:** Zp, Zs and density

**Anomaly A (Figure 4.7)**
Anomaly A is a strong and consistent reflection in the 5-bin inversion with P-wave impedance ranging from 18 000 to 19 000 g/cm³·m/s.

**Anomalies B and C (Figure 4.8)**
Anomaly B has an impedance of up to 17 000 g/cm³·m/s. However, impedances along Anomaly B are very inconsistent.

The top layer of Anomaly C has an impedance of 16 000 g/cm³·m/s, the lower layer has an impedance of 1400 g/cm³·m/s. The background impedance at this depth is ~ 5000 g/cm³·m/s.

**Anomaly D (Figure 4.9)**
The Anomaly D reflection is strong but slightly less consistent than in the acoustic impedance inversion. P-wave impedances are up to 23 000 g/cm³·m/s in places. Some areas below the high impedance area are as low as 2000 g/cm³·m/s.

**Anomaly E (Figure 4.10)**
The upper reflection of the lower part of Anomaly E has an impedance of 12 000 g/cm³·m/s, the lower reflection has an impedance of 2000 g/cm³·m/s. The upper part of Anomaly E has P-wave impedances up to 11 000 g/cm³·m/s but some patches are as low as 1200 g/cm³·m/s.

**Anomaly F (Figure 4.8)**
The upper Anomaly F reflection has an impedance of ~6500 g/cm³·m/s, the lower reflection has an impedance of 1800 g/cm³·m/s. The background impedance around Anomaly F is ~3200 g/cm³·m/s.
4.6.3 CMP 18000 model

This run uses a background model calculated from interval velocities picked from a single CMP gather (CMP18000) to create a pseudo-well which were subsequently extrapolated across the entire section, guided by several horizons in the Hampson-Russell software. Direct comparison between the effects of different starting models, can be made between this run and the interval model run (Section 4.6.4) (Figure 4.3).

**Inversion Parameters:**

- **Starting Model:** Extrapolated CMP 18000 impedances, calculated using Gardner’s equation.
- **Gathers:** 20 bins, 0-50°
- **Method:** Prewhitening
- **Iterations:** 50
- **Background Vs/Vp (Gamma):** 0.5
- **Zp prewhitening:** 1%, **Zs prewhitening:** 5%, **Density prewhitening:** 10%
- **Updating:** Zp, Zs and density

**Anomaly A (Figure 4.7)**

Anomaly A has consistent P-wave impedance at ~17 000 g/cm³·m/s, some areas directly below Anomaly A have impedances as low as 1000 g/cm³·m/s.

**Anomalies B and C (Figure 4.8)**

Anomaly B is very patchy with no consistent high impedance area. P-wave impedances for the anomaly range from ~4600 to 9200 g/cm³·m/s.

The upper reflection of Anomaly C has an impedance of ~16 500 g/cm³·m/s and the lower reflection has an impedance of ~2800 g/cm³·m/s. The background impedance in the area is ~2300 g/cm³·m/s.

**Anomaly D (Figure 4.9)**

Anomaly D has a consistent P-wave impedance of ~18 000 g/cm³·m/s.
Anomaly E (Figure 4.10)
The upper reflection of the lower area of Anomaly E has an impedance of 12 000 g/cm³·m/s, the lower reflection has an impedance of 3700 g/cm³·m/s. The upper area of Anomaly E has upper and lower layer impedances of 6400 g/cm³·m/s and 2100 g/cm³·m/s respectively.

Anomaly F (Figure 4.8)
The background impedance around Anomaly F is ~3000 g/cm³·m/s. The upper reflection of the Anomaly F has an impedance of ~3600 g/cm³·m/s, the lower reflection has an impedance of 2700 g/cm³·m/s.

4.6.4 Interval model
This run uses background impedance models calculated from the stacking velocities for the entire STRATUS-2D-002 line. The same inversion parameters and angle gathers used for the CMP 18000 inversion (Section 4.6.3) are used for this inversion. Only the background model is different. Therefore, direct comparison can be made on the effects of different starting models on inversion.

Inversion Parameters:

- **Starting Model:** Zp, Zs and density interval models calculated from stacking velocities using Castagna’s and Gardner’s equations.
- **Gathers:** 20 bins, 0-50°
- **Method:** Prewhitening,
- **Iterations:** 50
- **Background Vs/Vp (Gamma):** 0.5
- **Zp prewhitening:** 1%, **Zs prewhitening:** 5%, **Density prewhitening:** 10%
- **Updating:** Zp, Zs and density

Anomaly A (Figure 4.7)
Anomaly A has consistent P-wave impedance of ~17 000 g/cm³·m/s. In some areas directly below Anomaly A P-wave impedances are as low as 1400 g/cm³·m/s.
**Anomalies B and C (Figure 4.8)**

Anomaly B is patchy with no consistent high impedance area, impedance ranges from 4000 to 10 000 g/cm³·m/s.

The upper reflection of Anomaly C has an impedance of ~15 000 g/cm³·m/s and the lower reflection has an impedance of ~2500 g/cm³·m/s. The background impedance in the area is ~2000 g/cm³·m/s.

**Anomaly D (Figure 4.9)**

Anomaly D is strong and consistent with a P-wave impedance of ~12 000 g/cm³·m/s.

**Anomaly E (Figure 4.10)**

The upper reflection of the lower area of Anomaly E has an impedance of 9200 g/cm³·m/s, the lower reflection has an impedance of 2600 g/cm³·m/s. The upper area has upper and lower layer impedances of 5500 g/cm³·m/s and 1700 g/cm³·m/s respectively.

**Anomaly F (Figure 4.8)**

The background impedance around Anomaly F is ~3300 g/cm³·m/s. The upper reflection of the Anomaly F has an impedance of ~3900 g/cm³·m/s, the lower reflection has an impedance of 2650 g/cm³·m/s.

**4.6.5 Interval model – covariance=0.5**

Background impedance models for this run were calculated from the stacking velocities for the entire STRATUS-2D-002 line. Instead of the prewhitening method used in the other inversion runs, in this run the covariance of the natural log of each parameter of interest (i.e., lnZp, lnZs, and lnρ) is set, indicating the expected range of each parameter (which also indicates confidence in the accuracy of each parameter)(Hampson-Russell, 2012). For this run all parameters were given an equal covariance of 0.5.
**Inversion Parameters:**

- **Starting Model:** Zp, Zs and density interval models calculated from stacking velocities using Gardner’s equation.
- **Gathers:** 0-40° in 15 bins
- **Method:** Covariance
- **Iterations:** 50
- **Background Vs/Vp (Gamma):** 0.5
- **Prewhitening:** 1%
- **Zp covariance:** 0.5, **Zs covariance:** 0.5, **density covariance:** 0.5
- **Updating:** Zp, Zs and density

**Anomaly A (Figure 4.7)**

Anomaly A has a consistent impedance of 6000 g/cm³·m/s, with no significant low impedance areas, excluding one small area with an impedances of 2000 g/cm³·m/s.

**Anomalies B and C (Figure 4.8)**

Anomaly B is variable, with P-wave impedances up to 8900 g/cm³·m/s.

The upper reflection of Anomaly C has an impedance of ~8700 g/cm³·m/s, the lower reflection has an impedance of 3100 g/cm³·m/s.

**Anomaly D (Figure 4.9)**

Anomaly D is moderately patchy with impedances reaching ~10 000 g/cm³·m/s. Some areas beneath Anomaly D have impedances as low as 3700 g/cm³·m/s.

**Anomaly E (Figure 4.10)**

The upper reflection in the lower area of Anomaly E has an impedance of 8800 g/cm³·m/s, the lower reflection has an impedance of 3500 g/cm³·m/s. The upper area of Anomaly E has impedances of up to 5400 g/cm³·m/s, but some patches are as low as 2000 g/cm³·m/s.
Anomaly F (Figure 4.8)
The upper Anomaly F reflection has an impedance of ~4000 g/cm$^3$·m/s, the lower reflection has an impedance of 2600 g/cm$^3$·m/s. The background impedance around Anomaly F is ~3000 g/cm$^3$·m/s.

4.6.6 RAYINVR model
This inversion run was carried out using a five-layer starting impedance model calculated using iterative forward modelling with RAYINVR of shot 11401 of the STRATUS-2D-002 line (equivalent to CMP location 17480). The resulting pseudo-well was extrapolated across the entire STRATUS-2D-002 line guided by horizons picked in the Hampson-Russell software.

Inversion Parameters:
- **Starting Model:** Zp, Zs and density interval models calculated by ray-tracing using RAYINVR and Castagna and Gardner’s equations.
- **Gathers:** 40 bins, 0-50°
- **Method:** Prewhitening
- **Iterations:** 50
- **Background Vs/Vp (Gamma):** 0.5
- **Zp prewhitening:** 1%, **Zs prewhitening:** 2%, **Density prewhitening:** 5%
- **Updating:** Zp, Zs and density

Anomaly A (Figure 4.7)
Anomaly A has a consistent high impedance of 7500 g/cm$^3$·m/s, surrounded by impedances as low as 1400 g/cm$^3$·m/s.

Anomalies B and C (Figure 4.8)
Anomaly B has a P-wave impedance of ~5000 g/cm$^3$·m/s. although it is not consistent along the entire anomaly.
The upper reflection of Anomaly C has an impedance of ~6600 g/cm$^3\cdot$m/s, the lower reflection has an impedance of 1700 g/cm$^3\cdot$m/s.

**Anomaly D (Figure 4.9)**

Anomaly D shows consistent P-wave impedances across the anomaly reaching ~9900 g/cm$^3\cdot$m/s. The background impedance in the area is ~5000 g/cm$^3\cdot$m/s.

**Anomaly E (Figure 4.10)**

The upper reflection of the lower area of Anomaly E has an impedance of 7700 g/cm$^3\cdot$m/s, the lower reflection has an impedance of 3400 g/cm$^3\cdot$m/s. The upper area of Anomaly E has impedances up to 5000 g/cm$^3\cdot$m/s, but some patches are as low as 2500 g/cm$^3\cdot$m/s. The background impedance is approximately 3500 g/cm$^3\cdot$m/s.

**Anomaly F (Figure 4.8)**

The upper reflection of Anomaly F has an impedance of ~3000 g/cm$^3\cdot$m/s, the lower reflection has an impedance of 2400 g/cm$^3\cdot$m/s. In this run Anomaly F is difficult to distinguish from the background impedance of ~2400 g/cm$^3\cdot$m/s.
Figure 4.7 - Anomaly A

P-wave impedance inversion comparison. The same colour scale is used for all inversions.

(a) Acoustic Impedance inversion - Anomaly A P-wave impedance of up to 6500 g/cm³·m/s.
(b) Interval model inversion - 5-bin - Anomaly A P-wave impedance of 18000-19000 g/cm³·m/s.
(c) CMP18000 model inversion - Anomaly A P-wave impedance is approximately 17000 g/cm³·m/s. Some areas directly below Anomaly A are as low as 1000 g/cm³·m/s.
(d) Interval model inversion - Anomaly A P-wave impedance is 17000 g/cm³·m/s.
(e) Interval model inversion - covariance 0.5 - Anomaly A P-wave impedance is consistent at ~6000 g/cm³·m/s.
(f) RAYINVR model inversion - Anomaly A P-wave impedance is consistent at 7500 g/cm³·m/s.
**Figure 4.8** - Anomaly B, C, and F P-wave impedance inversion comparison.

(a) Acoustic Impedance inversion - Anomaly B, C and F impedances are 9500, 8500 and 2600 g/cm³.m/s respectively.

(b) Interval model inversion - 5-bin - Anomaly B, C and F impedances are 17000, 16000 and 1800 g/cm³.m/s respectively.

(c) CMP18000 model inversion - Anomaly B, C and F impedances are 9200, 16500 and 2700 g/cm³.m/s respectively.

(d) Interval model inversion - Anomaly B, C and F impedances are 10000, 15000 and 2650 g/cm³.m/s respectively.

(e) Interval model inversion - covariance 0.5 - Anomaly B, C and F impedances are 8900, 8700 and 2600 g/cm³.m/s respectively.

(f) RAYINVR model inversion - Anomaly B, C and F impedances are 5000, 6600 and 2400 g/cm³.m/s respectively.
Figure 4.9 - Anomaly D

P-wave impedance inversion comparison. The same colour scale is used for all inversions.

(a) Acoustic Impedance inversion - Anomaly D impedance 6500 g/cm².m/s.

(b) Interval model inversion - 5-bin - Anomaly D impedance 23000 g/cm².m/s.

(c) CMP18000 model inversion - Anomaly D impedance 18000 g/cm².m/s.

(d) Interval model inversion - Anomaly D impedance 12000 g/cm².m/s.

(e) Interval model inversion - covariance 0.5 - Anomaly D impedance 10000 g/cm².m/s.

(f) RAYINVR model inversion - Anomaly D impedance 9900 g/cm².m/s.
Figure 4.10 - Anomaly E P-wave impedance inversion comparison. The same colour scale is used for all inversions.
(a) Acoustic Impedance inversion - Upper portion impedance 5300 g/cm³.m/s. Lower portion impedance 8500 g/cm³.m/s.
(b) Interval model inversion - 5-bin - Upper portion impedance 11000 g/cm³.m/s. Lower portion impedance 12000 g/cm³.m/s. (c) CMP18000 model inversion - Upper portion impedance 6000 g/cm³.m/s. Lower portion impedance 12000 g/cm³.m/s.
(d) Interval model inversion - Upper portion impedance 5500 g/cm³.m/s. Lower portion impedance 9200 g/cm³.m/s.
(e) Interval model inversion - covariance 0.5 - Upper portion impedance 5400 g/cm³.m/s. Lower portion impedance 8800 g/cm².m/s.
(f) RAYINVR model inversion - Upper portion impedance 5000 g/cm³.m/s. Lower portion impedance 7700 g/cm³.m/s.
4.7 Inversion interpretations

All inversion runs show similar P-wave impedance responses for all anomalies from background impedance values, i.e., the same areas show higher or lower than background impedances. However, the actual impedance values and in some cases the magnitude of relative impedances contrasts differ significantly, with the strongest control on impedances seeming to be due to the different starting models applied. As no independently measured impedances were used to create the background model, i.e. well data, all inverted impedances values are relative.

The only independent measure of P-wave impedance available for the data set is the approximate impedance of Anomaly A calculated from seafloor reflection coefficients, which ranges from 4660 to 8800 g/cm$^3$·m/s (Section 3.5). Whilst this is a very approximate measure of the impedance, it may help to rule out some inversion runs as unrealistic. The inversions which have P-wave impedance values for Anomaly A which fall within the impedances calculated from the reflection coefficients are: model-based poststack inversion, prestack inversion – covariance=0.5, and the RAYINVR model inversion.

The 5-bin prestack inversion run has consistently higher impedance values than the other inversion runs for all anomalies. The difference from the other inversion runs indicates that some amplitude variation with angle is probably being stacked out using this wider angle binning scheme. However, these wide-angle bins will also stack out some noise from the data, resulting in higher signal-to-noise ratios in the input gathers, and possibly more stable inversion.

Impedance values for typical rocks

Typical P-wave impedances ranges for a variety of rocks measured predominantly from petroleum well sonic-logs are presented below to provide a context for interpretation of the inverted P-wave impedances (Table 4.1).
<table>
<thead>
<tr>
<th>Rock Type</th>
<th>P-wave Velocity (m/s)</th>
<th>S-wave Velocity (m/s)</th>
<th>Density (g/cm(^3))</th>
<th>Z(_{p}) (g/cm(^3)·m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sedimentary</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sandstone</td>
<td>2800-4460</td>
<td>1590-2930</td>
<td>2.0-2.8</td>
<td>5600-12488</td>
</tr>
<tr>
<td>Shale</td>
<td>1830-5180</td>
<td>2.3-2.8</td>
<td></td>
<td>4209-14504</td>
</tr>
<tr>
<td>Limestone</td>
<td>3500-6400</td>
<td>2000-3700</td>
<td>2.3-2.9</td>
<td>8050-18560</td>
</tr>
<tr>
<td>Dolomite</td>
<td>3500-6500</td>
<td>1900-3600</td>
<td>2.4-2.9</td>
<td>8400-18850</td>
</tr>
<tr>
<td>Marl</td>
<td>2000-3000</td>
<td>740-1500</td>
<td>1.5-2.2</td>
<td>3000-6600</td>
</tr>
<tr>
<td>Sand, gravel</td>
<td>400-2100</td>
<td>150-600</td>
<td>1.4 (dry)-2.3(wet)</td>
<td>560-4830</td>
</tr>
<tr>
<td>Igneous</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Granite</td>
<td>4180-5850</td>
<td>2610-3300</td>
<td>2.750</td>
<td>11495-16088</td>
</tr>
<tr>
<td>Andesite</td>
<td>3970-6130</td>
<td>2440-3500</td>
<td>2.636</td>
<td>10465-16158</td>
</tr>
<tr>
<td>Diorite</td>
<td>4830-6370</td>
<td>2960-3350</td>
<td>2.823</td>
<td>13635-17982</td>
</tr>
<tr>
<td>Basalt</td>
<td>4560-6150</td>
<td>2370-3510</td>
<td>2.785-2.840</td>
<td>12700-17466</td>
</tr>
<tr>
<td>Gabbro</td>
<td>5970-6870</td>
<td>2700-3650</td>
<td>2.970-2.991</td>
<td>17730-20548</td>
</tr>
</tbody>
</table>

**Anomaly A**

Anomaly A maintains a consistent high impedance for all runs, with impedances ranging from 6000 to 19 000 g/cm\(^3\)·m/s. In some runs Anomaly A is bounded by very low impedance areas with impedances as low as 1000 g/cm\(^3\)·m/s. These low impedances are unrealistically low, being lower than that of seawater at ~1500 g/cm\(^3\)·m/s, and are probably an artefact of the very high reflection coefficient causing Anomaly A. The Hampson-Russell Suite inversion algorithm assumes that the small reflectivity approximation holds for the data being inverted (Section 4.3.1). This means that reflections coefficients ideally need to be less than 0.1; Anomaly A has an approximate reflection coefficient of 0.5 to 0.7, exceeding this limit significantly. Therefore, the inverted impedances are probably not reliable for Anomaly A.

Whilst some of the inverted impedance values are relatively low and within the realm of sedimentary rock impedances at ~6000 g/cm\(^3\)·m/s, the very high amplitude of Anomaly A, and its location are unlikely to be caused by a sedimentary unit, i.e. a sandstone bed within mudstones. The only feature that is likely to produce such a strong positive reflection is an igneous intrusion or flow, therefore the original interpretation of a shallow igneous intrusion may hold, which is consistent with some of the high impedance values obtained in some inversion runs of up to 19 000 g/cm\(^3\)·m/s.

**Anomaly B**

Anomaly B does not have consistent impedance along the anomaly. Impedances range from 5000 to 9000 g/cm\(^3\)·m/s within an inversion run and, in some runs, impedance reaches 17 000 g/cm\(^3\)·m/s. This inconsistent impedance is probably due to the likely 3-D nature of
Anomaly B and the associated high noise and poor NMO correction. Whilst these impedances are high, suggesting Anomaly B is caused by a high-velocity reflective body consistent with the basalt sill hypothesis, they are inconsistent between runs and the high reflectivity of Anomaly B probably make them unreliable (due to the small reflectivity approximation). Therefore the impedances for Anomaly B will not be used for further interpretation directly, aside from determining whether Anomaly B is continuous with Anomaly C.

The original interpretation of Anomaly B as a basalt sill is still considered the most likely lithology to cause the anomaly.

Anomaly C
All P-wave impedance inversion results confirm the hypothesis that Anomaly C is most likely a continuation of Anomaly B. Both the upper high-impedance and lower low-impedance zones of Anomaly C appear to be continuous with areas of similar impedance in Anomaly B. The high P-wave impedances, up to 16 000 g/cm³·m/s are consistent with a basalt sill interpretation. However, similar to Anomalies A and B the high reflectivity of Anomaly C may exceed the small reflectivity approximation used in the Hampson-Russell inversion algorithm, suggesting the inverted impedances are probably unreliable.

Anomaly D
P-wave impedance values range from 9000 to 23 000 g/cm³·m/s for Anomaly D. The 23 000 g/cm³·m/s impedance value is unrealistic; it could only be obtained from a gabbro or ultramafic unit, both of which are very unlikely at the location of Anomaly D. However, excluding the 23 000 g/cm³·m/s impedance value, the inverted impedances are consistent with a basalt sill interpretation. However, like Anomalies A, B and C, the reflection coefficient of Anomaly D is likely to higher than 0.1 and the small reflectivity approximation probably does not hold.

Some impedance values for Anomaly D fall within the reported sandstone impedance range. However, sandstones at the higher end of the sandstone impedance range are unlikely to occur in shallow sandstone intrusions, which will probably consist of relatively low impedance poorly consolidated sands, and therefore can probably be discounted as a likely cause for Anomaly D.
**Anomaly E**

The upper area of Anomaly E always shows higher than background impedances generally at 5500 g/cm\(^3\)·m/s (although the 5-bin inversion has values of 11 000 g/cm\(^3\)·m/s), which is inconsistent with the shallow free gas interpretation presented in Chapter 3. However, there are always associated lower than background impedances below these high impedance areas. Impedance values in these low impedance areas may be as low as 1700 g/cm\(^3\)·m/s. If these are not due to errors in the inversion, these values would be consistent with gas charged sediments.

The lower area of Anomaly E has moderately high P-wave impedances at 7000 to 12 000 g/cm\(^3\)·m/s. These impedances suggest that the lower part of Anomaly E also be caused by an igneous intrusion.

Relative to Anomalies A, B, C and D, Anomaly E has a low reflection coefficient, which may not exceed the small reflectivity approximation of the Hampson-Russell algorithm, and therefore may have reliable inverted impedances.

**Anomaly F**

In all inversion, runs Anomaly F is characterised by a paired P-wave impedance anomaly with a high impedance upper section and a low impedance lower section. In all runs, except for the RAYINVR model, Anomaly F is clearly distinguishable from the background impedance. Whilst not easily visually distinguishable in the RAYINVR model inversion, the upper and lower parts of the anomaly still have measurably higher and lower impedances, respectively, to the background impedance. However, they are very similar to the background impedance. The low impedance section of Anomaly F is probably is consistent with the interpretation of free gas causing Anomaly F. However, the presence of the upper high impedance area is more difficult to interpret, it could be either due to (1) the presence of gas hydrate, for which its formation would have to be limited strongly by the methane flux along the erosion surface controlling Anomaly F, or perhaps more realistically it could be due to (2) incorrect estimation of the phase of the source wavelet resulting in the formation of impedance side lobes or another artifact of the inversion process (Latimer et al., 2000). Further work would be needed to confirm either of these hypotheses. However, the close spatial association
of Anomaly F to inferred shallow igneous bodies (Anomalies B and C) suggests that the gas hydrate interpretation of the high-impedance area is unlikely as these intrusions (depending on their age) will perturb the local thermal gradient, resulting in shallowing of the gas hydrate stability zone.
Chapter 5

Discussion and conclusions

5.1 Overview

Seismic interpretation techniques including waveform inversion were used on the STRATUS-2D-002 seismic line in the Reinga Basin to investigate the nature of six major amplitude anomalies. The preferred interpretations of these anomalies range from shallow basalt sills (Anomalies A, B, C and D) to stratigraphically controlled shallow gas migrating updip from a reverse fault (Anomaly F).

No direct indicators of gas hydrate (e.g. BSR) were observed on STRATUS-2D-002 within the potential gas hydrate stability zone. However, Anomalies E and F suggest that there is fault controlled gas migration from deeper in the basin.

In general, full waveform inversion of amplitude anomalies identified on STRATUS-2D-002 was unsuccessful due to the very high reflection coefficients of the anomalies, which violated the small reflectivity approximation assumed in the Hampson-Russell Suite inversion algorithm. However, most inversion runs, with both differing parameters and starting models, show similar P-wave impedance responses for all anomalies examined.

This chapter is going to attempt to constrain the geological interpretations of the amplitude anomalies described in Chapters 3 and 4, the relative age of these anomalies, and discuss their significance to the evolution of the Reinga Basin and its hydrocarbon potential. The inversion methods applied to STRATUS-2D-002 and their limitations will also be discussed in Section 5.3. The findings of this study and potential future work will be summarised at the end of this chapter.
5.2 Amplitude anomalies

The preceding chapter included descriptions of the main amplitude anomalies imaged in this thesis. This section attempts to use that information to constrain the geological interpretations of these anomalous bodies. The preferred interpretation of the amplitude anomalies imaged on the STRATUS-2D-002 line are summarised in Table 5.1.

5.2.1 Interpretations of anomalous bodies

Anomaly A

The very high reflection coefficient and amplitude, broad extent and strongly attenuated amplitudes below Anomaly A can only be convincingly explained by the presence of a shallow igneous body, either a sill or flow. This interpretation is enforced by the close association to Anomalies B and C which are inferred to represent shallow basaltic sills. The preferred interpretation of Anomaly A is that it is caused by a flow rather than a sill as the anomaly is less than 50 m below the present seafloor and an intrusion of a 6 km long sill is unlikely at such a shallow depth.

There is no evidence of direct connection between Anomaly B and Anomaly A on STRATUS-2D-002. However, Anomaly B is likely to have similar dimensions into the plane of the seismic profile (i.e., it will have an approximate diameter of ~2.5 km); and perhaps the flow that formed Anomaly A was fed via a peripheral dyke from the Anomaly B sill located outside of the plane of the seismic profile.

Anomalies B and D

Anomalies B and D are inferred to represent shallow basalt saucer-shaped sills, which have folded the overlying sediment during intrusion. Although intruded sandstones can result in similar structures in seismic reflection data, (e.g., Huuse et al. 2004, and Huuse and Mickelson, 2004), Anomalies B and D are more likely to represent igneous intrusions due to their relatively flat reflection morphology (sand intrusions commonly have ~30° dips), very high amplitude, large associated velocity pull-up beneath them, and spatial association with basaltic volcanic units of potentially similar age in the Reinga Basin (e.g. Mortimer et al. 1998 and Herzer et al. 1997).
The similarity of the feature identified on the AGSO 114-04 line by Herzer et al. (1997) to be a fault bounded pop-up structure (see Figure 3.24) to Anomalies B and D (the presence of high amplitude negative-polarity reflections below a localised anticline of similar size), and spatial association to Anomaly D, indicate that it probably also represents a saucer-shaped sill with overlying associated folding rather that a faulted pop-up and potentially, depending on the accuracy of positioning of the AGSO 114-04 seismic line may actually be the same feature as Anomaly D.

The negative polarity of these anomalous reflections inferred to be basaltic sills suggests that strong reflections are only occurring from the lower surface of these sills (high impedance to low impedance), perhaps the upper surfaces are more irregular causing more scattering, resulting in upper sill reflections without anomalously high amplitudes. Negative-polarity reflections inferred to be sills are well represented in the literature, e.g., Hansen and Cartwright (2006), Hansen et al. (2008).

Anomalies B and D - Sill thickness and emplacement depth

Anomalies B and D are both overlain by anticlines, which have formed in response to the intrusion of the Anomaly B and D sills. Several studies suggest strong correlations exist between saucer-shaped sill thickness and diameter, and emplacement depth and overlying fold relief (e.g. Hansen and Cartwright (2006) and Polteau et al. (2008).

Hansen and Cartwright (2006) suggest that where sill thickness is above the \(\lambda/2\) resolution limit, a 1:1 ratio between sill thickness and fold relief exists. The relief on the Anomaly B and D anticlines is approximately 95 and 100 ms TWT, respectively. So if we take the velocity to be constant in this at 2000 m/s, that would correspond to fold reliefs of 95 and 100 m, respectively. If the 1:1 relationship of Hansen and Cartwright (2006) holds it would suggest that the sills forming Anomalies B and D are approximately 100 m thick.

Polteau et al. (2008) present a emplacement depth to sill diameter relationship based off several different data sets, including both igneous and intruded sand sills measured in seismic data, numerically modelled data and sills mapped on land. If we apply this relationship to anomalies B and D, which have diameters in 2-D of approximately 2.5 and 3 km, respectively, the resulting emplacement depths range from 440 to 1200 m, and 550 to 1400 m, respectively (Figure 5.1). These emplacement depth ranges are deeper than emplacement depths estimated directly from the data, measured for the top of the anomalies to the inferred paleoseafloor horizon (last reflection displaced by the bounding faults), which
suggest emplacement depths of 450 and 330 m, respectively if a velocity of 2000 m/s is assumed. However, the emplacement depth estimates measured from the data have numerous sources of potential error, including the velocity used for depth conversion and whether the true paleoseafloor horizon was picked.

**Figure 5.1** – Sill diameter to emplacement depth estimates after Polteau et al. (2008). The black lines indicate the upper and lower bounds of the constraining data. The blue lines represent the Anomaly B sill width and projected upper and lower emplacement depth estimates. The red lines represent the Anomaly D sill width and projected upper and lower emplacement depth estimates. Modified from Figure 4a in Polteau et al. (2008).

**Anomaly C**

Whilst the inverted impedance values for Anomaly C are unlikely to be reliable due to their high reflection coefficient (see Section 5.3.3 below), they do support the interpretation that Anomaly C is an extension of the Anomaly B sill. In all the inversion runs presented Anomaly C, appears continuous with the NW portion of Anomaly B, with no offset across the NW Anomaly B bounding fault. A continuous sill causing Anomalies B and C is a realistic interpretation, with Anomaly B forming an inner sill and Anomaly C forming an outer sill joined by an inclined dyke. e.g., Thomson and Hutton (2004) and Polteau et al. (2008).

**Anomaly E**

The upper area of Anomaly E most likely represents shallow free gas accumulation at the same erosion surface horizon that confines Anomaly F. However, similar to Anomaly F these
low impedance areas are also associated with areas of high impedance, the cause of which is somewhat difficult to determine but is probably an artifact of the inversion.

Interpretation of the lower part of Anomaly E remains enigmatic. The lower area of anomalous reflections coincide with the UB6 horizon, and show relatively high-amplitude negative-polarity reflections with very gentle concave-upwards morphologies. These negative polarity reflections are associated with relatively high P-wave impedance (up to 12 000 g/cm$^3$·m/s), and probably represent small basaltic sills, although in this case, unlike Anomalies B and D there is no associated deformation.

**Anomaly F**

Anomaly F is inferred to represent free gas migrating updip from the southeast Anomaly B bounding fault. This is predominantly based on the association of higher amplitudes near the fault that become weaker and then patchy updip along the erosion surface horizon, which controls its location. Inverted impedances for Anomaly F indicate the presence of a high-impedance layer over a low impedance layer relative to the background impedance. The presence of a lower than background impedance agrees well with the free gas interpretation of Anomaly F. However, the high impedance layer above is more difficult to interpret, and presents three main possibilities, either, (1) gas hydrates, (2) an inversion artifact, or (3) a sandy layer.

At less than 200 m below the seafloor (which is approximately 2000 m deep at its location), Anomaly F is likely to be well within the gas hydrate stability zone, and because of the high velocity of gas hydrate (see Section 2.3.1) we would expect a high impedance layer if gas hydrate was present. However, high impedance areas are only associated with the low impedance areas, which would suggest that if there is any hydrate present its formation is strongly dependent on methane flux along the erosion surface horizon, as the high impedance layer is not present elsewhere.

Another potential explanation for the high impedance layer in Anomaly F is that it is an artifact of the inversion process, in particular, the result of a poorly estimated source wavelet. If the phase component of the estimated wavelet is not correct, it may result in the formation of side lobes in the data, i.e. the development of areas with false impedance values around an anomaly due to the rotation of the wavelet from the actual phase in the data (Latimer et al., 2000). However, as all inversion runs appear to have this high and low impedance pair for
Anomaly F, and all of these runs have independently extracted statistical wavelets, it is unlikely that this high-low impedance pair is an artifact of poor wavelet estimation. However, this does not rule out any other types of inversion artifacts.

The high impedance areas of Anomaly F could potentially represent a sandy layer, which would have higher impedance that the mud-rich units that probably surround Anomaly F (see Table 1.1 – Waka Nui-1 well summary table). However, the patchy nature of the anomaly, especially at its southeastern end, would be difficult to explain by the deposition of a geologically feasible sand body in a deepwater setting.

The preferred interpretation of Anomaly F is that it represents stratigraphically controlled free gas. The close spatial association of Anomaly F to the southeast bounding fault of Anomaly B and the decreasing amplitude and patchy nature of Anomaly F away from this fault suggest that gas migration to the erosion surface which confines Anomaly F, is occurring from deeper in the basin along the southeast Anomaly B bounding fault.

<table>
<thead>
<tr>
<th>Anomaly A</th>
<th>Basaltic flow. Possibly associated with intrusion of Anomaly B.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Anomaly B</td>
<td>Shallow saucer-shaped igneous intrusion (probably basaltic), which has caused overlying deformation during intrusion.</td>
</tr>
<tr>
<td>Anomaly C</td>
<td>Probably an outer-sill from the Anomaly B sill.</td>
</tr>
<tr>
<td>Anomaly D</td>
<td>Shallow saucer-shaped igneous intrusion (probably basaltic), which has caused overlying deformation during intrusion.</td>
</tr>
<tr>
<td>Anomaly E</td>
<td>Upper area – stratigraphically controlled free gas (confined by the erosion surface horizon). Lower area – possibly a shallow sill.</td>
</tr>
<tr>
<td>Anomaly F</td>
<td>Stratigraphically controlled free gas (confined by the erosion surface horizon), migrating updip from the southeast Anomaly B bounding fault.</td>
</tr>
</tbody>
</table>

5.2.2 Anomaly timing/age

Anomalies B, C, D, and the lowest part of Anomaly E coincide with the late Early-Miocene horizon UB6 of Bache et al. (2012). The folds above anomalies B and D deform the UB6 reflection, indicating that folding occurred after its deposition towards the end of the Early-Miocene, and the presence of ~400 ms of concordant folded strata above UB6 suggest that a significant amount of deposition occurred before folding. Overlying reflections onlap on to the upper-most deformed reflection above Anomaly B. This onlap surface, termed UB9 in this study (in keeping with the Bache et al. convention), can be traced throughout most of STRATUS-2D-002 (Figure 5.2). The UB9 surface is not cut by the faults that bound Anomaly B, although it may be the uppermost folded reflection. Tracing of the UB9 horizon places it in...
the uppermost folded reflections above Anomaly D. The crest of the Anomaly D anticline has been eroded, and this erosion surface can be traced to Anomaly B where it becomes coincident with the UB9 surface. Anomaly F is confined to the erosion surface horizon. Reflections below this erosion surface onlap on to the uppermost folded reflections above Anomaly D (probably the UB9 surface). The UB9 surface is the upper-most reflection to be offset by the Anomaly D bounding faults. The location of the UB9 surface above Anomaly B is uncertain, but the crest of the anticline shows some erosion.

The UB9 surface corresponds to the paleoseafloor at the time of intrusion of Anomaly D and the associated folding above it. The erosion surface horizon, which constrains Anomaly F and cuts the crest of the Anomaly D anticline, appears to onlap the folded reflections above Anomaly B, although the onlap signature is uncertain, and it may actually erode the Anomaly B anticline. If the UB9 surface is not folded above Anomaly B, it suggests that intrusion of Anomaly D occurred before that of Anomaly B. However, the relationship of UB9 and the Anomaly B anticline is uncertain; both anomalies may be of a similar age.

Anomaly A is possibly stratigraphically higher than the erosion surface horizon above Anomaly B and very near the modern seafloor, indicating that it is either the same age or younger than Anomalies B and D.

The upper portion of Anomaly E is also confined to this erosion surface, indicating that it is younger than Anomalies B and D and possibly younger than Anomaly A.
Figure 5.2 - Annotated PSTM section of the STRATUS-2D-002 line. Anomalies B and D are indicated with blue letters B and D respectively. The pink horizon represents the UB9 onlap surface discussed in the text. The dark blue horizon represents the UB6 horizon after Bache et al. (2012), and the green horizon represents the UB9 erosion surface.
5.2.3 Significance of anomalous bodies to the Plio-Pleistocene to Recent development of the Reinga Basin and hydrocarbon potential

No direct indicators of gas hydrate were observed in the STRATUS-2D-002 line such as bottom simulating reflections. However, two of the amplitude anomalies studied indicate the presence of shallow gas accumulation. Anomaly F, in particular suggests that there may be active gas migration from deeper in the basin, due to its close association with the southeast bounding fault of the Anomaly B anticline, and apparent updip migration from this area.

The presence of Pliocene/Pleistocene igneous activity in the northwest and Oligocene-Miocene igneous activity in the southeast of the Reinga Basin, and two phases of compression indicate that the Reinga Basin probably has a complex thermal history. The presence of young shallow igneous bodies (e.g., Anomalies B and D) are likely to locally perturb the geothermal regime in the area (including the gas hydrate stability zone) leading to local shallowing of the gas hydrate stability zone.

Saucer shaped sills, with associated forced-folded anticlines, are potential hydrocarbon traps with four-way dip closure (Hansen and Cartwright, 2006). Whilst saucer-shaped sills are inferred at Anomalies B and D, they are very shallow and faulted (i.e., no intact potential seal rocks), and unlikely to have petroleum potential themselves. However, if similar features occur at greater depths in other areas of the basin, they may provide potential hydrocarbon targets if suitable reservoir and seal rocks occur within the folded strata, and they are not disrupted by faults.

5.3 Inversion methods and limitations

All inversion runs showed similar P-wave impedance responses (i.e., consistently higher or lower than background for a given anomaly) for all of the anomalies of interest. However the actual P-wave impedance values for a given anomaly varied widely between runs. Anomaly D for example, showed impedances between runs ranging from ~9000 to 23 000 g/cm³·m/s. There are several possible reasons for these wide ranging, and probably unreliable, inverted impedances, which will be discussed in the following sections including:

1. the lack of seismic-tied well data,
2. different methods of background impedance model generation,
3. high reflection coefficients for most of the anomalies of interest, and
4. the highly 3-D nature of some of the anomalies.
No systematic error analysis is included with the inversion results presented as such error can only be calculated from the misfit between the real data and a synthetic well log with the Hampson-Russell Suite software. Therefore, as only low frequency pseudo-wells were created this caused all error values to be unrealistic, i.e., five ‘reflections’ in synthetic traces from the pseudo-wells compared to 100’s in real data traces. However, the range and variety of inversion parameters and background models used gives an impression for the possible relative error present in these analyses. Whilst the impedance values for all the inversion runs are different, all anomalies show the same response from the background (i.e., always higher or lower than the background impedance), often with similar magnitude (relative to the background impedance value) in the majority of inversions. This suggests that the possible error in these analyses may be low, but the inverted impedances are only relative, and strongly influenced by the starting model.

5.3.1 Absence of seismic-well ties

The lack of nearby, seismically-tied wells with a full suite of wireline logs, including sonic and density logs, poses many limitations to achieving a successful inversion result. In particular, low frequency background impedance trends measured independently from the seismic data cannot be extracted to guide inversion, as well as a lack of measured impedances for any formations in the area which can be used to assign real impedance values to the inversion result. Without well control, even if the correct scaling has been applied to the synthetic data during inversion, the reported impedance values are only relative.

Accurate low-frequency trends are required for full waveform inversion to guide the inversion towards a realistic model, as inversion is non-unique and there are numerous models that could fit the observed data (Virieux and Operto, 2009). These background impedance models strongly control the inversion result; therefore wrong or poorly parameterised models may guide the inversion to local misfit minimum. Without well control to provide low frequency information to the inversion, background impedance models need to be estimated using other methods, such as using processing stacking velocities or tomography based methods.

5.3.2 Background models

Three methods were employed to produce low frequency background impedance models for inverting the STRATUS-2D-002 line (see Chapter 4. Section 4.4). Two models were based
off processing stacking velocities and one was calculated by iterative forward modelling using shot ray-tracing with RAYINVR. Two types of impedance models were produced using the three methods listed above:

1. entire line impedances from interval velocities which allow for lateral variation in the data based off the processing stacking velocities (see Section 4.4.1), and
2. extrapolated pseudo-wells, where pseudo-wells were created from either individual CMP interval velocities or from iterative forward modelling in RAYINVR and extrapolated across the entire STRATUS-2D-002 line constrained by defined reflection horizons in the data (see Sections 4.4.2 and 4.4.3).

P-wave and S-wave impedances were calculated from P-wave velocities using Castagna’s and Gardner’s relationships (Gardner et al., 1974; Castagna et al., 1985). Local modifications to these relationships were not applied, and the lack of independently measured S-wave and density information in the background models tended to cause inverted S-wave impedance and density to very closely follow the P-wave impedance.

All inversion runs show that the background models have a strong control on the inverted impedances, as expected, as they supply the low frequency impedance trends to the inversion. Inversions carried out using the extrapolated pseudo-well method create sharp impedance boundaries in the inverted impedances when a horizon used to guide the original extrapolation is crossed, and limit the impedance variation unrealistically between ‘layers’ of the background model. These pseudo-well impedance models result in unreasonable impedance models at distance from the pseudo-well location.

The background impedance models derived from the stacking velocities for the entire line produced more realistic inversion results, without the strong contrasts that were typical for the other models, whilst still producing similar relative impedance contrasts across the areas of interest. The models produced by stacking velocities for the entire seismic line produced more geologically reasonable inversions.

Future attempts to invert data without well control to provide the low frequency impedance information for the background model would benefit from a more vigorous approach to calculate the background model for the entire line, such as creating a P-wave velocity model by travel-time inversion with RAYINVR using all shots.
5.3.3 Small reflectivity approximation

Anomalies A, B, C, and D exhibit very high amplitudes, indicating that they are associated with large impedance contrasts and therefore large reflection coefficients. Of these anomalies, the only independent estimate of the reflection coefficient available is for Anomaly A, obtained whilst calculating seafloor reflection coefficients (see Section 3.5), which has estimated reflection coefficients of 0.5 to 0.7. The other anomalies of interest have lower, but similar amplitudes to Anomaly A, indicating that they will have similarly high reflection coefficients.

These high reflection coefficients pose a problem for waveform inversion using the Hampson-Russell Suite software. The linearised approximation of Zoeppritz equations, based on a modified version of the Aki and Richards approximation, applied in the Hampson-Russell inversion algorithm assume that the small reflectivity approximation is valid (Hampson et al., 2005). The small reflectivity approximation assumes that the reflectivity coefficient, \( R_i \), can be approximated by \( \frac{Z_i}{2Z_i} \) such that \( R_{pi} \approx \frac{Z_{pi}}{2Z_{pi}} \), where \( i \) is the \( i^{th} \) interface between layer \( i \) and \( i+1 \). However, this approximation is only valid where reflection coefficients are equal to or less than 0.1 (Hampson et al., 2005; Hampson-Russell, 2012). Clearly the high reflection coefficients of Anomalies A to D exceed this assumption. Therefore any inverted impedances from these anomalies are likely to be unreliable, as the response of the inversion to these high reflection coefficients is unknown. Anomalies E and F have much lower amplitudes that the other anomalies, and probably do not exceed the <0.1 reflection coefficient assumption, therefore may have reliable inverted impedances.

5.3.4 3-D properties of anomalous bodies

Anomalies B and D, inferred to be saucer-shaped sills, are inherently three-dimensional (3-D) features, which are likely to have similar dimensions out of the plane of the STRATUS-2D seismic profile (i.e., they have diameters of approximately 3 km). Imaging of such 3-D structures with two-dimensional (2-D) seismic profiles presents several potential problems, both for processing and subsequent inversion. In particular, multi-path reflections (reverberations within the feature) and out-of-plane reflections may violate the assumption that reflections have a hyperbolic travel-time curve with offset (and therefore are not able to be NMO-corrected). 3-D features may not be adequately migrated using 2-D migration methods, especially if there is no dominant dip-direction. 2-D migration of 3-D features may introduce artifacts into the data and affect relative reflection amplitudes (Yilmaz, 2001).
Therefore any inversion of 3-D features in 2-D seismic data may give unrealistic values, because relative amplitudes may not have been retained in processing, and out-of-plane reflections and reverberations may also be present.

5.4 Conclusions

The analysis of the STRATUS-2D-002 profile has enabled a number of conclusions to be drawn about the nature of amplitude anomalies within the Reinga Basin, and the viability of waveform inversion methods with no well control.

1. Six amplitude anomalies were identified on STRATUS-2D-002, most of which appear to be related to the presence of shallow igneous intrusions, Anomalies A, B, C, and D.

2. Reprocessing of the STRATUS-2D-002 line resulted in some improvement in imaging of the near surface, in particular increased reflection continuity beneath Anomaly A and reduction of jittery spatial aliasing noise around Anomaly F.

3. No direct gas hydrate indicators were observed in the Reinga Basin (e.g., BSR). However, Anomaly F is inferred to represent stratigraphically controlled shallow gas.

4. Anomalies B and D represent ~ 100 m-thick saucer-shaped sills, which have formed overlying anticlines during intrusion, which deformed the paleoseafloor.

5. The compressional pop-up identified on AGSO 114-04 by Herzer et al. (1997) is inferred to also represent a saucer-shaped sill with an associated anticline, that is in close proximity to, or potentially the same feature as, Anomaly D.

6. The occurrence of Anomaly F appears to be associated with the termination of the southeast Anomaly B bounding fault, with higher gas saturation inferred near the fault, which decreases in concentration updip, based on observed amplitudes along the anomaly.

7. Waveform inversion using the Hampson-Russell Suite with no well control did not produce reliable impedance inversions for the STRATUS-2D-002 line.

8. The strongest control on these waveform inversion results appears to be the background impedance model; therefore, where no well data is available, vigorous
methods need to be applied to infer the low frequency background impedance; perhaps travel-time tomography for the entire section of interest would be adequate.

9. The small reflectivity approximation used by the Hampson-Russell Suite inversion does not hold for most of the anomalies identified on STRATUS-2D-002, therefore their inverted impedances for these anomalies are unreliable.

5.5 Future work

Improving imaging of the Areas around Anomalies B, D

Anomalies B and D are inherently three-dimensional structures and therefore, conventional processing methods do not result in adequate imaging of the anomalies themselves or below them. Application of more sophisticated processing techniques, such as SRME, inverse-Q gain correction and depth migration may result in improved imaging in and below these features and enable a better understanding of the geology below these features (i.e. evidence of feeder dykes), and the geometry of the basaltic sills causing these high-amplitude reflections.

Improved starting models for inversion with no well control

Most of the anomalies identified on STRATUS-2D-002 exceed the small reflectivity approximation assumed by the inversion algorithm. However, Anomalies E and F may not exceed this limitation. Whilst there is no well data in the basin, improvement of background impedance model estimation is likely to produce more realistic results, and if some higher frequency information can be contained in these models as well as the low frequency information required to guide the inversion, systematic error quantification could be realistically carried out during the inversion analysis step, before inverting the entire section.

Two potential methods could be employed to create less subjective background impedance models; (1) P-wave tomography/travel-time inversion of all shots in the profile, or (2) P-wave pseudo-well logs calculated from stacking velocities from multiple crossing seismic lines, i.e., where the profile of interest intersects with other seismic lines in the area, a velocity model could be derived for each of these locations that fits all of the traces involved, similar to the method employed by Stagpoole (2011) for the Waka Nui-1 well. If all of the Reinga-09
and STRATUS-2D data sets were used this would result in approximately 70 pseudo-well locations across the Reinga Basin that could be used to guide future inversions.
References


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