Title: Seismic Interpretation, Prospects and Structural Analysis, Great South Basin

Operator: OMV New Zealand Ltd

Author: OMV New Zealand Ltd, Francesca Ghisetti

Date: 2010

This report has been compiled from material submitted to the New Zealand Government under legislation or voluntarily by exploration companies. An acknowledgement of this work in the following bibliographic format would be appreciated:

Author/Operator; Date; Title; Ministry of Economic Development New Zealand Unpublished Petroleum Report PR4173
Structural Analysis of the Great South Basin

Author: Francesca Ghisetti

PEP 50119, 50120, 50121

Work Programme Obligation 9.a.v.

Operator: OMV New Zealand Limited

Date: 2010-June

NZ-EXP-10-003-C Regional Structural Analysis of the GSB.pdf
Great South Basin, New Zealand
PEPs 50119, 50120, 50121

Structural Analysis of the Great South Basin

NZ-EXP-10-003-C

Francesca C. Ghisetti, April 2010
Final Report prepared for OMV (Contracts EXP09-011 and NZ 2010-007)

<table>
<thead>
<tr>
<th>Rev No</th>
<th>Date</th>
<th>Revision</th>
<th>Originator</th>
<th>Checked</th>
<th>Approved</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>13/04/2010</td>
<td>Draft for Review</td>
<td>FG</td>
<td>RAC</td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>27/04/2010</td>
<td>Technical revisions &amp; clarifications</td>
<td>FG</td>
<td>RAC</td>
<td>RAC</td>
</tr>
<tr>
<td>C</td>
<td>20/05/2010</td>
<td>Formatting, captions</td>
<td>EB</td>
<td>TA</td>
<td>TA</td>
</tr>
</tbody>
</table>

Confidential

Copyright © 2010, OMV New Zealand Limited, Wellington, New Zealand

This document is the property of the PEPs 50119, 120 & 121 Joint Venture. Circulation is restricted to Joint Venture partners and their designated associates, contractors and consultants. It shall not be copied or used for any other purpose other than which it is supplied, without the expressed written authority of the PEPs 50119, 120 & 121 Joint Venture Operator. Except where provided for purposes of contractual requirements, the PEPs 50119, 120 & 121 Joint Venture disclaims any responsibility or liability for any use or misuse of the document by any person and makes no warranty as to the accuracy or suitability of the information to any third party. Any misuse of the document is redressable by the PEPs 50119, 120 & 121 Joint Venture.
INDEX

1. Introduction .............................................................................................................. 7
2. Regional Tectonic setting of the Great South Basin ............................................. 10
   2.1 Basement Terranes ............................................................................................... 11
   2.2 Late Cretaceous Rifting ....................................................................................... 15
   2.3 Growth and Establishment of the Convergent Plate Boundary Across New Zealand ......................................................................................................................... 18
3. Structural geometry onshore and offshore .......................................................... 21
   3.1 Onshore Structural Map ....................................................................................... 22
   3.2 Offshore Structural Map ...................................................................................... 25
   3.3 Structural Domains ............................................................................................. 27
4. Structural assemblages in the seismic lines used for the regional transects .......... 30
   4.1 Syn-Rift Structural Assemblages ........................................................................ 31
   4.2 Post-Rift Structural Assemblages ....................................................................... 32
   4.3 Syn-Orogenic Structural Assemblages ............................................................... 34
5. Regional transects .................................................................................................. 38
   5.1 Transect T01 (Plate 4) ........................................................................................ 39
   5.2 Transect T02 (Plate 5) ........................................................................................ 42
   5.3 Transect T03 (Plate 6) ........................................................................................ 44
   5.4 Transect T04 (Plate 7) ........................................................................................ 46
   5.5 Transect T05 (Plate 8) ........................................................................................ 48
6. Restoration and balancing of the regional transects ............................................. 50
   6.1 Goals and Methods ............................................................................................ 50
   6.2 Restored Regional Transects .............................................................................. 53
   6.3 Length and Area Balance of the Regional Transects ........................................... 62
7. SYNTHESIS ............................................................................................................. 65
   7.1 Deformation Sequence ......................................................................................... 65
   7.2 Deformation Mechanisms ................................................................................... 67
   7.3 Crustal Stretching ................................................................................................. 79
   7.4 Crustal shortening ............................................................................................... 85
   7.5 Uplift .................................................................................................................. 87
8. CONCLUSIONS AND Recommendations ......................................................... 90
9. ACKNOWLEDGEMENTS ..................................................................................... 95
10. REFERENCES ......................................................................................................... 96
EXECUTIVE SUMMARY

This report presents a regional structural analysis of the Great South Basin (GSB), commissioned by OMV for the areas of PEP 50119, 50120, 50121, and adjacent onshore regions.

Goals of this study are to:

1. Reconstruct the geometry, time sequence, and mechanisms of deformation imposed on the GSB and adjacent onshore regions since Late Cretaceous rifting.
2. Provide a regional structural synthesis through construction, restoration and balancing of regional geological transects that link offshore and onshore geology.
3. Quantify the horizontal (stretching and shortening) and vertical (uplift, subsidence, tilting) components of deformation in the study area.

The new analyses carried out for this study incorporate relevant published literature and geological cartography (see References). The offshore analysis is based on public domain data and on the confidential database provided by OMV for this study, with - in particular —100 seismic lines across the GSB, structural and isopachs maps of key horizons, chrono-stratigraphic definition of the seismic sequences and their interpretation for a number of the depth-converted seismic lines.

The regional structural setting both onshore and offshore is summarised in Plate 1 and Tables 1, 2, 3 that show the trend and structural characteristics of the major fault sets, within adjacent, but distinct tectonic domains. Plate 1 shows:

- Good correlation between the onshore and offshore tectonic domains.
- The imprint of the inherited Terrane architecture (with WNW-ESE Terrane boundaries).
- Dissection of the upper crust by sets of closely-spaced N-S to NE-SW and ENE-WSW faults formed during Late Cretaceous rifting, but variably reactivated in compression since the late Miocene.
- Relationships between faulting and syn-tectonic sedimentation (see Plate 2 for the chrono-stratigraphy).

The structural assemblages within the study area are based on OMV’s interpretation of TWT seismic data, tied to sparse well control, and subsequently converted to depth. Enclosure 1 contains selected examples taken from the same seismic lines used for constructing the regional geological transects. Principal features are:

- The syn-rift faults are arranged in conjugate sets; many faults dip at low angle (≤ 30º). Vertical throws are of the order of 1-2 km on average. Activity was intense in the early syn-rift phases.
- The post-rift structural setting is dominated by the gravitational detachment of the Mass Transport Complex (MTC), and by the detached flexure of the Eastern Imbricate Belt (EIB), the last one localised at the transition between thick post-rift sequences in the subsident basin and thin sequences above the uplifted eastern margin.
- The syn-orogenic structural setting is dominated by compressionally reactivated normal faults, predominantly within a preferential belt close to the coastline.
- The coastline coincides with the hinge of a compressionally elevated crustal region. Structural elevation is accommodated by rigid translation and rotation along the inherited faults. Some evidence exists of compressional (transpressive) reactivation along the Terrane boundaries. Another structure that deforms the syn-orogenic sequences is the NE-SW Toroa Anticline, characterised by gentle, low amplitude and long wavelength folding.
Five regional transects (T01 to T05, with orientation NW-SE) summarise the geologic setting of the GSB ([Plates 4-8], see [Plate 1] for their traces) down to depths of 11 km. These transects are based on the onshore geology (Institute of Geological and Nuclear Sciences Qmaps), and use the sequence stratigraphy of OMV for the definition of the sedimentary sequences ([Plates 2 and 3]). The offshore geology is derived from the interpretation of suitable, depth-converted seismic lines (see [Plates 4-8] for the identification and the location of the seismic lines).

The five geological transects T01-T05 have been restored and balanced (using LithoTect™ Software) by progressively removing superposed deformations in steps that follow the inverse chronological order of their formation (from the youngest to the oldest). Deformation mechanisms used for the reconstructions are rigid body rotation and translation (for basement units), vertical shear and fault-parallel shear (for the syn-rift and syn-orogenic units) and layer-parallel shear (for the post-rift units). The results are shown in [Plates 9-13] and [Plate 14]. Restoration and balancing quantify the syn-rift stretching (low to moderate, with $\beta$ factor in the range 1.1-1.4) and the syn-orogenic shortening (low, from 0.98 to 0.84).

Major points are:

- The geometry of the GSB is dominated by the imprint of the Late Cretaceous rifting. Basement faults are crustal structures with a long history of activity. Faults at the western margin are inferred to be listric. The geometry of the eastern margin is not constrained by the available data.

- Post-rift subsidence was accommodated by differential tilting (with layer-parallel slip and pervasive jointing) and accompanied by the gravitational instabilities of the detached Mass Transport Complex and Eastern Imbricate Belt.

- Late Miocene to Present shortening is dominated by the compressionally reactivated inherited normal faults with folding of cover sequences and by the Toroa anticline. Terrane boundaries were possibly reactivated with transpressive left-lateral shear during this stage. Basin modelling (OMV confidential data) indicates that the long-wavelength folding of the Toroa anticline was initiated by the vertical load imposed by the mass of the Plio-Quaternary delta prograding at the western margin of the basin. Thus, compressional inversion – if active - was late and of minor effect.

The analyses performed in this study provide a regional reconstruction of the geometry, chronology and sequence of deformation ([Plate 15] and [Figure 6]) from the Late Cretaceous (112 Ma) to the Present; show that the proposed interpretations are coherent and kinematically admissible, and validate the mapping of the seismic horizons and sequences performed by OMV.

However, further insights into the structural setting of the GSB are needed to: (1) reduce existing uncertainties; (2) validate the proposed deformation mechanisms (through forward modelling); (3) explore the role of gravity tectonics; (4) define the reactivation history of Terrane boundaries; and, (5) understand the relationships between deformation and redistribution of fluids within the basin.
LIST OF FIGURES

Figure 1  Regional tectonic setting of New Zealand in the Pacific region. The continental masses (with shelf bounded by the 2000 m isobath) are in grey. AU: Australian plate, PAC: Pacific plate, ANT: Antarctic plate. GSB: area of the Great South Basin. Modified after Sutherland (1999). ...................................................... 11

Figure 2  Simplified map of Terranes in the SE South Island. Terrane boundaries are traced in the offshore of the GSB following the interpretive map of OMV, and the data from exploration wells. ................................................................. 13

Figure 3  Movement trajectories of a body detached by mechanisms of gravity sliding, gravity spreading, or mixed mode between the two end members. The change in shape of the detached body is shown in light colour. In red is the basal detachment. Redrawn and modified after Rowan et al. (2004). ....................... 72

Figure 4  Progressive deformation of a syn-rift basin by selective compressional rectivation. Stages b and c are recognised in the offshore of the GSB. Stage d is recognised in the Otago province onshore. Redrawn and modified after Sibson and Ghisetti (2010). ................................................................. 74

Figure 5  Simplified structural map showing the spatial relationships between the belts of localised compressional inversion, the Toroa Anticline and the Terrane boundaries. Also shown is the position of the most pronounced flexure of the EIB in relation to the east flank of the Toroa Anticline. See text for details. ................................................................................................................ 75

Figure 6  Sequence of progressive deformation along an ideal transverse transect across the GSB and adjacent onshore regions. See text for a discussion. .......................... 78

Figure 7  Continental extension model of McKenzie (1978) and definition of the stretch factor $\beta$. Note that under the assumption of pure shear, if stretching in the rift occurs by a factor $\beta$, then thinning of the crust is $1/\beta$. ........................................ 79

Figure 8  Comparison of estimates of $\beta$, based on extension from crustal faults along selected transects (Transects T01-T05 and selected seismic lines in Cook et al., 1999), and based on tectonic subsidence from wells (contour lines reproduced from Cook et al., 1999). ........................................................................................................ 81

Figure 9  Transect T03 superposed on the crustal interpretation of the line AWI2003001 by Grobys et al. (2009). The crystalline crust of the GSB is thinned to 13 km (Tf). If the original thickness is $\sim$ 22 km (To, eastern Campbell Plateau), $\beta \sim 1.7$, but for original values up to 45 km of the Gondwana orogenic crust $\beta$ may be as high as 3................................................................. 84
**LIST OF PLATES**

**Plate 1** – Structural map of the Great South Basin and conterminous onshore regions. See Tables 2 and 3 for the geometric and structural definition of the faults identified with progressive number.

**Plate 2** – Chrono-stratigraphic Table. Reproduced from OMV. See Table 1 for additional information on rifting.

**Plate 3** – Legend for the regional transects (Plates 4-8), the restored transects (Plates 9-13) and the balanced transects (Plate 14).

**Plate 4** – Geological transect T01. Location in Plate 1. Legend in Plate 3.

**Plate 5** – Geological transect T02. Location in Plate 1. Legend in Plate 3.

**Plate 6** – Geological Transect T03. Location in Plate 1. Legend in Plate 3.

**Plate 7** – Geological Transect T04. Location in Plate 1. Legend in Plate 3.

**Plate 8** – Geological Transect T05. Location in Plate 1. Legend in Plate 3.

**Plate 9** – Restored geological transect T01. Legend in Plate 3.

**Plate 10** - Restored geological transect T02. Legend in Plate 3.

**Plate 11** – Restored geological transect T03. Legend in Plate 3.

**Plate 12** – Restored geological transect T04. Legend in Plate 3.

**Plate 13** – Restored geological transect T05. Legend in Plate 3.

**Plate 14** – Balanced geological transects T01-T05. Legend in Plate 3.

**Plate 15** – Chronology of deformation events in the Great South Basin.
**Table 1** - Chronology and style of Cretaceous rifting episodes from literature. See References for the quoted sources.

**Table 2** – Geometric and structural characteristics of the major onshore faults mapped in Plate 1. The columns define: (1) fault name and location in the GNS Qmaps; (2) minimum fault length in km; (3) quadrant of dip direction; (4) dip angle (estimated whenever possible from maps and published data); (5) quadrant of down-faulted block; (6) maximum throw (estimated whenever possible from maps and published data); (7) type of movement (estimated whenever possible from maps and published data); (8) evidence for Quaternary activity; (9) syn-rift sediments preserved in the fault hanging wall; (10) indexed fault number in Plate 1; (11) additional structural information (from maps and published data).

**Table 3** - Geometric and structural characteristics of the major offshore faults that offset the top of basement, mapped in Plate 1. The columns define: (1) fault number (as in Table 1) and location in the offshore structural domains; (2) orientation (see Table 1 for the definition of orientation classes); (3) quadrant of dip direction; (4) dip angle (when measurable); (5) estimated average vertical throw of the top basement (in km); (6) evidence for compressional inversion.

**Enclosure 1** – Structural assemblages in the seismic lines used for the construction of the regional transects T01-T05 (Plates 4-8).
1. Introduction

The Great South Basin (GSB) occupies a vast crustal domain (~ 85,000 km\(^2\)) from the shortened margin of the SE South Island to the differentially foundered, offshore region, resting on normal to attenuated continental crust.

From the Late Cretaceous rifting to the Paleogene post-rift sagging and the Neogene shortening, the GSB experienced a sequence of superposed tectonic processes, each one characterised by tectonic assemblages with distinct geometry, crustal penetration and kinematics, as well as by contrasting finite strain (extension versus shortening), in response to the changes of regional stress field controlled by relative motion of the Australia, Pacific and Antarctic plates.

A regional analysis of the geometry and time sequence of the structural assemblages in the GSB, tied to the geology of the adjacent margin emergent in the South Island of New Zealand, and incorporating the newly acquired set of high quality offshore seismic data is fundamental for assessing the vertical and horizontal mobility of the crustal domain that hosts the sedimentary sequences of the basin; the structural segmentation and along-strike variations in the GSB controlled by major fault systems, and the impact of superposed deformation on the thermal evolution of the basin, fluid circulation, and the petroleum system.

Under the contracts EXP09-011 and NZ 2010-007 OMV commissioned F. C. Ghisetti (TerraGeoLogica) to carry out a regional structural study of the GSB.

Defined goals of this study included:

- Assessing the structural architecture of the basement in the GSB and adjacent emergent areas, with emphasis on the imprint of the inherited Terrane fabric, and the structural setting in the rifted region.
• Defining and classifying the geometry and kinematics of structural assemblages formed in the sequence of superposed tectonic events that affected the GSB and adjacent regions since the rifting stages, with definition of the major faults, characterisation of the structures that deform the cover sequence, identification of domains with contrasting structural style.

• Providing a regional synthesis of the structural geometry and evolution of the GSB, through construction, sequential restoration and balancing of regional geological transects.

• Quantifying the horizontal (stretching and shortening) and vertical (uplift, subsidence and tilting) components of deformation across the study area.

This Final Report and accompanying Figures, Plates, Tables and Enclosure describe the methods of the study, the principal results, the interpretations based on available data, and the uncertainties that still need to be addressed for better understanding the structural evolution of the GSB. It utilises a database of public domain data (see References) together with confidential stratigraphic and seismic data acquired and elaborated by OMV. OMV also supplied the depth conversion and the horizon interpretation for all the seismic lines incorporated in the regional transects.

The new structural analyses described in this report include relevant bibliographic data on the study area (see References), and – for the onshore part of the study – available cartographic data. In consideration of the regional scale of the study, geological cartography uses the geo-referenced Qmap series (scale 1:250,000), published by the Institute of Geological and Nuclear Sciences.

Definition of the structural assemblages is based on the analysis and interpretation of ~100 seismic lines, selected from the whole database of OMV as the most representative for reconstructing the structural style and the chronology of superposed deformation within the GSB. The lines belong to surveys 71_h_a; 72_h_a; 72_h_b; 72_h_e; 72_h_f; 83_h; 83_u; AWI2003; CB_82; DUN_06, OMV_08 and SIGHT.
The geological interpretations of the seismic data, the construction of cross sections, the sequential restoration and balancing of the regional geological transects have all been performed using the mapping and interpretive tools of “LithoTect™” software (Geo-Logic systems, LLC, Boulder, Co, USA, http://www.geologicystems.com).

The statements, opinions, and conclusions provided in this report are given in good faith, and in the belief that all the efforts have been made to incorporate all the relevant data supplied by OMV and/or available in the public domain.

The writer believes that the review and conclusions provided in this report give a sound interpretation of the existing data, but the acquisition of new data in the region of study and in the adjacent areas and further examinations may result in the refinement and/or modification of the interpretations reached so far.
2. Regional Tectonic setting of the Great South Basin

The two major islands of New Zealand are the emergent tip of a large submerged continental mass (~2x10^6 km^2), composed of plateaux and troughs and crossed by the subduction margins of the Hikurangi and Puysegur trenches, linked by the right-lateral Alpine Fault transform boundary. The edge of the continental shelf lies close to the 2000 m isobath. The area of sedimentation and subsidence of the Great South Basin (GSB) is located within the Campbell Plateau, at the south-eastern margin of the South Island (Figure 1).

The present-day crustal setting of the south-west Pacific region around New Zealand (Figure 1) bears the imprint of a complete plate tectonic cycle with: (1) Cambrian to Early Cretaceous convergence, subduction and accretion at the Pacific-facing Gondwana margin; (2) Late Cretaceous rifting, with separation of the Australia, Pacific and Antarctica plates along a complex system of ridges, and dispersal of the New Zealand micro-continent within the Pacific plate; and, (3) establishment in the last 45 Ma of a new active plate margin between the Australia and Pacific plates that controls the emergence of the landmass of the North and South Islands of New Zealand and the present active oblique convergence at rates of ~4 cm/year, with associated seismicity and volcanism. South of New Zealand, spreading at the Pacific-Antarctic ridge is still active, with rates of ~6 cm/year.

The long-term deformation history of New Zealand and surrounding regions has been analysed in a large number of regional geological studies, based on field mapping, stratigraphic correlations, metamorphic petrology, geochemistry, and geo-chronological dating. Some papers are the basis of this summary review (e.g. Cook et al., 1999; Sutherland, 1999; Sutherland et al., 2000; Mortimer et al., 2002; Eagles et al., 2004; Laird and Bradshaw, 2004; Mortimer, 2004; Wandres and Bradshaw, 2005; Tulloch et al., 2006; Kula et al., 2007; LeMasurier and Landis, 1996; Sutherland et al., 2010), and the reader is referred to these papers for additional references that are not specifically quoted in this report.
Figure 1  Regional tectonic setting of New Zealand in the Pacific region. The continental masses (with shelf bounded by the 2000 m isobath) are in grey. AU: Australian plate, PAC: Pacific plate, ANT: Antarctic plate. GSB: area of the Great South Basin. Modified after Sutherland (1999).

2.1 Basement Terranes

The earliest phases of convergence at the Gondwana margin are documented by a suite of Middle Cambrian to late Early Cretaceous “Terranes” (far-travelled volcano-sedimentary and igneous units with metamorphic overprints, juxtaposed along transpressive shear zones) that make up the basement foundations of New Zealand.
These Terranes juxtapose paleotectonic realms belonging to subduction zones, ridges, magmatic arc systems and intervening basins.

The identified Terranes (Mortimer 2004) are affected by oroclinal bending and dissection resulting from the superposed Neogene deformations with strike-slip displacement of ~460 km along the Alpine Fault. Plate tectonic restorations have been used to re-assemble the crustal configuration of the Gondwana margin before the onset of rifting and sea-floor spreading (Sutherland, 1999; Wandres and Bradshaw, 2005; Tulloch et al. 2006).

The configuration at 90 Ma (see Figure 1 of Tulloch et al. 2006) is of particular significance for understanding the present distribution of basement rocks in the South Island, the subsequent evolution of the crust of the Campbell Plateau (that hosts the GSB), and the heterogeneous lithospheric strength inherited from the earliest deformation stages. Terranes can be subdivided between Eastern and Western Province, separated by the >800 km long belt of the Median Batholith. The Median Batholith consists of Devonian to Cretaceous plutonic intrusions (375-105 Ma), with a main phase of Mesozoic magmatism between 170-105 Ma. Rocks belonging to this phase are distributed into a younger (130-105 Ma) western belt (Mortimer et al., 1999; Tulloch and Kimbrough, 2003; Allibone and Tulloch, 2004) and an older (130-170 Ma), eastern belt (see Figure 2 of Tulloch et al., 2006).

The plutons of the two belts differ in their geochemistry, indicating processes of late partial melting of basaltic continental lithosphere generated during the earlier stages of subduction-related arc magmatism.

The western belt has been interpreted as the locus of crust thick enough to isostatically support an Andean-style mountain range along the Cretaceous continental margin (“Cordillera Zealandia” of Tulloch et al., 2006). This mountain range was a likely source of localised sediment supply to the rift-basins formed at a later stage.

The Terranes that are of interest for the study of the GSB (Figure 2) are– from south to north - the Median Batholith (outcropping in Stewart Island), and the Brook Street,
Murihiku, Dun Mountain-Maitai, Caples and Rakaia Terranes, all belonging to the Eastern Province.

Figure 2 Simplified map of Terranes in the SE South Island. Terrane boundaries are traced in the offshore of the GSB following the interpretive map of OMV, and the data from exploration wells. Redrawn and modified after Mortimer et al. (2002).

The crustal tectonic setting resulting from the juxtaposition of Terranes is revealed by a recently acquired deep crustal seismic reflection profile (SESI line), described by Mortimer et al. (2002).
The general characteristics of the Terranes are summarised below.

The **Brook Street Terrane** consists of dominant basaltic volcanic and volcaniclastic rocks of Permian age (≤14 km thick), belonging to an intra-oceanic calc-alkaline island arc-basin complex.

The Brook Street Terrane is overthrust by the Murihiku Terrane along the **Letham Thrust Fault**.

The **Murihiku Terrane** consists of prevailing marine Late Permian- Late Jurassic volcaniclastic sandstones, mudstones and tuffs, in a sequence ≤13 km thick, deposited in a fore-arc (or possibly back-arc) basin, adjacent to an active volcanic island arc at the Gondwana margin, and metamorphosed to zeolite facies (Turnbull and Allibone, 2003). The marine facies evolve to proximal deltaic and non marine (with coal) at the top of the sequence.

The **Hillfoot Fault** is the boundary with the Dun Mountain-Maitai Terrane.

The **Dun Mountain-Maitai Terrane** comprises the Dun Mountain ophiolite assemblage (Coombs et al., 1976), a belt characterised by a strong magnetic anomaly that is the remnant slice of Early Permian oceanic crust, overlain by a sedimentary cover of ~6 km of Late Permian to Middle Triassic volcaniclastic sediments metamorphosed to zeolite and prehnite-pumpellyte facies (Maitai).

The **Livingstone Fault** is the boundary with the Caples Terrane.

The **Caples Terrane** comprises a 5-7 km thick Late Permian to Triassic andesitic to felsic volcaniclastic sequence with weakly metamorphosed sandstones and mudstones, originally deposited in submarine fans discharging towards a trench adjacent to an island arc (Turnbull, 1979a, 1979b; Roser et al., 1993).

The **Rakaia Terrane** is mainly composed by Permian to Late Triassic weakly
metamorphosed turbidites, with a distinct sandstone petrofacies (of rhyodacitic affinity) relative to the adjacent Caples Terrane, inferred to be derived from a continental volcano-plutonic arc involved into a subduction-related accretionary prism close to a continental margin (Mortimer, 2004).

The contact between the Caples and Rakaia Terranes is overprinted by the metamorphic belt of the low-grade to greenschist facies Otago Schists (Mortimer, 1993), that expose an exhumed portion (350º-410º C at 30-35 km depth) of the Jurassic-Cretaceous Rakaia-Caples accretionary prism, with a large-scale antiformal culmination sub-parallel to the WNW-ESE trend of the Terrane boundaries.

The offshore extension of these Terranes and relative boundaries into the GSB is based on the tectonic trends, sparse borehole information in the eight offshore wells (cf. Cook et al., 1999), the seismic facies of the basement in the reflection profiles, the tracing of the magnetic anomaly connected with the Dun Mountain ophiolite belt (cf. Sutherland, 1999), and the change in structural style as observed at top basement level. A general interpretation is provided in Figure 2.

2.2. Late Cretaceous Rifting

A rapid transition from convergence to extension, crustal thinning and formation of rifted basins is documented in New Zealand starting at ~112 Ma. The change in tectonic regime is marked by a diachronous regional unconformity ranging in age from late early to late Albian (Bradshaw, 1989; Laird, 1993). In the central Raukumara Peninsula and in the northeastern South Island (Marlborough region) the unconformity is overlain by sediments dated as mid-Albian (~101-102 Ma, see Laird and Bradshaw, 2004). The plate tectonic scenarios advanced for explaining the switch in tectonic regimes include ridge subduction (Bradshaw, 1989); subducted slab capture (Luyendyk, 1995), and mantle plume activity (Weaver et al., 1994; Sutherland et al., 2010).

Crustal extension was accommodated by low-angle ductile shear zones (with N-S to NE-SW stretching lineation relative to the present north), active in both the Eastern and
Western Province (Tulloch and Kimbrough, 1989; Forster and Lister, 2004; Gray and Foster, 2004; Laird and Bradshaw, 2004). Progressive exhumation and uplift in the Eastern Province is constrained by the age (~93-99 Ma) of the Tucker Hill pseudotachylytes (Barker, 2005), associated with activity of a normal shear zone at seismogenic depths (10-15 km). In the Western Province, large magnitude crustal extension prior to sea-floor spreading in the Tasman Sea is documented for the Paparoa metamorphic core complex (Tulloch and Kimbrough, 1989). Stretching lineations indicates a NNE transport direction (relative to present North), with rapid exhumation of mid-crustal rocks between 95-88 Ma derived from thermochronology (Spell et al., 2000). Forster and Lister (2003) have proposed that also the antiformal dome of the Otago Schists is inherited from an exhumed metamorphic core complex ~112-109 Ma old.

Extension resulted in erosional dissection of the previously shortened and thickened crust that was emergent above sea level, with formation of fault-bounded basins filled with terrestrial clastic deposits (and inter-layered tuffs) derived from adjacent mountain ranges. Facies range from fan deltas to river flood plains, debris flows and lacustrine (Laird and Bradshaw, 2004).

A mid-Albian (Urutawan) unconformity marks the base of coarse terrestrial deposits, with the first appearance of schist fragments in the Kyeburn Formation in central Otago (Bishop, 1974a). Correlative sequences are recorded at Shag Point, on the eastern coast. The mid-Albian unconformity and the sedimentary input record the propagation and growth of brittle faults breaking through the surface, and controlling morphology and sedimentation in the syn-rift basins. The sequences preserved at Kyeburn (Bishop, 1974a) are bounded by two sets of sub-orthogonal normal faults, trending NW-SE and NE-SW. Silicic tuffs in the Kyeburn Formation and Ignimbrites at Shag Point have been recently dated as ~ 112 Ma old (Tulloch et al., 2009); thus older than previously documented by Adams and Raine (1988).

Alkaline intraplate igneous activity at 99-95 Ma followed the early rifting stages (Laird and Bradshaw, 2004). The products have a geochemistry compatible with upwelling of a mantle plume that can be located between the present Marie Byrd Land in Antarctica
and the position of the future New Zealand (see also Sutherland et al., 2010).

At ~87 Ma (late Coniacian) the continental mass of New Zealand was still attached to east Gondwana and was an emergent, low-relief region, with sedimentation continuing in the area of the GSB (Laird and Bradshaw, 2004). Onset of sea floor spreading with drift of the New Zealand microcontinent away from Gondwana started at ~85 Ma in the Tasman Sea (Hayes and Ringis, 1973; Gaina et al., 1998). The oldest sea floor between the SE margin of the Campbell Plateau and West Antarctica is 83-79 Ma old (Sutherland, 1999).

A renewal of extensional activity from 89 to 82 Ma is documented for the Sisters Shear Zone (Tulloch et al., 2006; Kula et al, 2007) that bounds the eastern margin of Stewart Island relative to the GSB, and has an orientation oblique to the faults formed at earlier stages in the rifting. The Sisters Shear Zone is a continental, low-angle extensional detachment, with inferred displacement of 15-22 km, and 7-10 km of footwall exhumation (Kula et al., 2007; 2009). The last documented movements on the Sisters Shear Zone (83 Ma) are coeval with the transition from continental extension to formation of oceanic lithosphere at the Pacific-Antarctic spreading ridge.

Along the West Coast of the South Island, a later episode of rifting driven by spreading in the Tasman Sea occurred from 80 to 55 Ma (Laird, 1993; Beggs et al., 2008). Faults of this second episode are oriented NE-SW and NNE-SSW, i.e. sub-orthogonal to the WNW-ESE faults of the previous stage.

According to Cook et al. (1999) there is no evidence for a change in extension direction during a second rift episode in the GSB. New faults may have not been formed during this stage, but it is possible that inherited, suitably oriented faults parallel to the Terrane boundaries were reactivated in extension. Onshore, the NW-SE Waihemo and Waitaki Fault systems have an orientation compatible with reactivation during this stage.

Convective upwelling of the mantle and lithospheric necking during sea-floor spreading are associated with uplift, emergence and erosion, leading to breakup unconformities, followed by marine transgression recorded in the sedimentary sequences (Braun and Beaumont, 1989; Allen and Allen, 2005).
In New Zealand, breakup unconformities at 87-85 Ma (Laird and Bradshaw, 2004) are correlated with the progressive development of a low-relief erosion surface (Waipoumanu Erosion Surface, WES, Le Masurier and Landis, 1996), now exhumed in the Otago region. The surface is a composite feature, produced by fluvial erosion, followed by marine transgression and wave-cut planation. A similar low-relief surface can be traced in West Antarctica (LeMasurier and Landis, 1996). Both surfaces trace progressive post breakup relaxation and moderate subsidence during waning of extensional deformation. In the GSB, a low relief, diachronous erosive surface associated with marine planation has been mapped by OMV, and testifies to the transition from segmented rift basins to post break-up regional flexural warping, subsidence and marine transgression.

In Otago, WES cuts upper Paleozoic-Mesozoic schists and greywackes (with the youngest rocks 100-85 Ma old), probably emergent - prior to break-up time - in a rugged landscape with tectonic relief $\geq 1.5$ km along fault-controlled scarps (LeMasurier and Landis, 1996). Above WES rest fluvial and swamp deposits and shallow marine strata. These deposits are Late Cretaceous in age near the coast (~75 Ma) and become progressively younger inland (from Paleocene to Eocene and Oligocene), thus tracing the diachronous onlap of marine sediments above the wave-cut planar unconformity during continuous subsidence of the New Zealand continent (LeMasurier and Landis, 1996).

A summary of the rifting tectonic events is provided in Table 1.

### 2.3 Growth and Establishment of the Convergent Plate Boundary Across New Zealand.

South of the SW tip of the South Island in Fiordland, an inactive rift margin that propagated from the southeast Indian Ridge into southern New Zealand at $\sim 45$ Ma can be traced SW for $>500$ km (Resolution Ridge). Seafloor spreading began in the Emerald Basin and Solander Trough, and the associated extensional regime affected the
New Zealand region (Lamarche et al., 1997; King, 2000). By 40 Ma, spreading through the continental crust of the South Island was established, and an intra-continental rift system propagated northward along the Moonlight Fault Zone (Norris et al., 1978). The propagation trajectory of this boundary marks the precursory development of the new plate boundary across New Zealand, as suggested by its close spatial association with the position of the Alpine Fault that overprints and incorporates the earlier extensional features (Sutherland et al., 2000). The pole of rotation between the Australia and Pacific Plates in the interval 40-20 Ma was very close to the South Island, resulting in low components of motion.

The eastern passive margin of this rift zone defines the western margin of the Campbell Plateau. Deformation in the distant regions of the GSB during this event is poorly understood, but was likely of limited extent. The major impact of the plate boundary reorganisation was probably on the drainage systems and characters of sedimentation, because of the diversion of the clastic sediment supply into the Solander Trough (Cook et al., 1999). Along the western margin of the GSB, localised, short wavelength folding of the Eocene sequence above basement faults is contemporaneous with formation of the Solander Trough (see Section 4.2).

During the Oligocene-Miocene the progressive migration to the south-east of the rotation pole was responsible for the gradual change to dextral strike-slip and convergence along the Australia-Pacific plate boundary, with growth of the transform system of the Alpine Fault since 25 Ma. Increasing amounts of convergence characterise the Late-Miocene Quaternary plate movements, and were accommodated by reverse components along the plate boundary (with ~70 km of shortening across the Alpine Fault in the last 10 Ma) as well as in a wide belt of normal faults reactivated in compression on both sides of the Alpine Fault (Ghisetti and Sibson, 2006; Norris et al., 1990). Plate motion calculations indicate 600 km of dextral displacement since the early Miocene (Sutherland, 1995). The Alpine Fault is the major structure within a deformation zone 150-350 km wide (Sutherland et al., 2000). The Alpine Fault has accommodated >50% of this displacement since its inception, and accounts for ~70% of the current plate motion, with the present displacement vector at ~30° to the fault in the central South Island (Norris et al., 1990; Walcott, 1998). The development of the
modern plate boundary is associated with crustal thickening and uplift in the convergent belt, with rise of the Southern Alps, and increased sediment supply in the adjacent basins.

The accommodation of convergence in a wide belt away from the Alpine Fault is clearly demonstrated in the Otago region, where steep reverse faults displace the basement, cause repetition of metamorphic zones, and control recent folding and distribution of Quaternary basins. There is clear evidence that the faults are in many cases reactivated normal faults that belong to two sets oriented N-S to NE-SW and NW-SE (see next section).
3. Structural geometry onshore and offshore

The regional structural setting of the South Canterbury, Otago and Southland regions in the South Island has been documented by several decades of field research, and is largely synthesised in the geo-referenced Qmap Series at scale 1:250,000 (Institute of Geological and Nuclear Sciences Ltd). The onshore areas adjacent to the Great South Basin (GSB) are covered by the geological maps Murihiku (Turnbull and Allibone, 2003); Dunedin (Bishop and Turnbull, 1996); Wakatipu (Turnbull, 2000); Waitaki (Forsyth, 2002), and Aoraki (Cox and Barrell, 2007).

Plate 1 is a summary map of the onshore public domain data and of the major offshore faults (compiled from OMV data). Plate 1 provides an overview of the dominant fault sets that control the geometry of the basement, with particular emphasis on the mechanical segmentation associated with Terrane boundaries, and the structural imprint consequent on the Late Cretaceous rifting. Plate 1 and the accompanying Table 2 and Table 3 define: (1) the geometry and kinematics of the principal faults; (2) the cross-cutting relationships between fault sets; and (3) the structural domains characterised by distinct setting.

In Plate 1 orientation of onshore faults is differentiated with different colour for strike classes of 20° (from N000° to N 180°). Faults are identified in terms of their kinematics (normal faults, reverse faults, strike-slip faults, compressionally reactivated normal faults), and dip angle (whenever possible). Active faults are distinguished with a thicker line. Terrane boundary faults, ductile shear zones, axial traces of major anticlines and synclines and the axial trace of the Otago Schists antiform are also indicated. The onshore faults are indexed with progressive numbers that are the same as in Table 2, where additional information is provided (average fault trend, minimum fault length in km, dip direction, average dip angle, quadrant of the down-faulted block, maximum vertical throw in metres, fault kinematics, evidence for Quaternary activity, preserved syn-tectonic sedimentary sequences in the hanging wall).

Plate 1 shows only the major faults that offset the top basement offshore. Faults are subdivided in directional classes of 20° (from N000° to N 180°), but some classes are
not represented (101°-120°, 121°-140° and 161°-180°). Dip direction is shown by ticks. Inferred Terrane boundaries, morphological scarps parallel to Terrane boundaries, and axial traces of folds in the Murihiku Terrane are also indicated. Faults are indexed with progressive numbers, and additional information is provided in Table 3 (strike class, dip direction, dip angle, vertical throw of top basement, evidence for compressional inversion).

Plate 1 also shows the domain of most intense exhumation of the Otago Schists onshore, and its relationships with ductile shear zones and Terrane boundaries, which control the distribution of syn-rift, hanging wall clastic deposits preserved in scattered outcrops. The area of exhumation is also the area crosscut by closely-spaced systems of compressionally reactivated normal faults (several of which are still active in the present tectonic regime) and is the domain affected by the largest uplift relative to the subsiding domain presently below sea level. The area of deepest foundering of top basement offshore is also indicated.

3.1 Onshore Structural Map

Dominant Structural Sets

Onshore, the predominant fault sets are oriented NE-SW (directional classes 021°-040° and 041°-060°). Numerous faults belonging to these classes extend for length ≤70 km (in few cases ≥100 km), though individual fault segments are generally 10-15 km long. Faults are spaced ~20 km on average. N-S trends (000°-020° and 161°-180°) are less frequent, but present north of the Waitaki Fault.

A second set of NW-SE faults (directional classes 121°-140° and 141°-160°) is frequent, especially along the segmented systems of the Waitaki (#32), Waihemo (# 21) and Tuapeka (# 7 and 8) Faults, that extend regionally for length ≥100 km.

E-W trends are generally poorly developed and limited to scattered, short fault segments.
Structural segmentation along Terrane boundaries is relevant, especially considering the mechanical and strength contrast between Terranes with distinct lithology and inherited structures. In particular, the Dun Mountain-Maitai Terrane, bounded by the Livingstone Fault to the north (with active fault traces), and the Hillfoot Fault to the south is a belt of mechanical weakness between the Caples and Murihiku Terranes. Most faults of the Otago region do not extend south across this belt.

The Rakaia-Caples Terrane boundary (overprinted by the Otago Schists metamorphism) runs sub-parallel to the foliation in the Otago Schists and was probably reactivated as a low-angle extensional detachment during Late Cretaceous rifting. This boundary is crosscut and offset by all the major NE-SW faults of the Otago region. The active Waitaki and Blue Lake Faults (# 32 and # 30, respectively) are located at the northern boundary between the Otago Schists and the non-metamorphosed greywackes of the Rakaia Terrane.

Structural Inheritance of the Late Cretaceous Rifting

Shortening of the onshore regions is the consequence of the Miocene-present transpressive movements between the Australia and Pacific plates. However, the inherited fabric of the Late Cretaceous rifting is still well preserved, mostly because the compressional stress field has caused preferential reactivation of the earlier fault systems, rather than creation of new fault sets.

The activity of both NE-SW and NW-SE normal faults during the Late Cretaceous is well documented by clastic sequences of Cretaceous age (Kyeburn and Horse Range Formations, Blue Spur Conglomerates, Henley Breccia) preserved in the hanging wall of the Waihemo (# 21), Tuapeka (# 7 and 8), and Titri (# 1) Faults (Bishop, 1974a; b; Litchfield, 2001; Els et al., 2003). Hanging wall clastic sequences are also documented for the Ostler Fault (# 36); however, these sequences are Neogene in age, and there is no proof of Cretaceous sediments ever having been deposited so far inland (Ghisetti et al., 2007).

Notably, the areas of preserved Late Cretaceous syn-rift sediments are distributed at the
margins of the antiform dome of the exhumed Otago Schists, bounded by NW-SE and NE-SW faults. Both the Tuapeka and Waihemo Faults are sub-parallel to ductile shear zones associated with extensional metamorphic core complexes. This setting suggests that the brittle, upper crustal faults that controlled the distribution of Late Cretaceous sedimentary basins have their crustal roots in the deep-seated shear zones that are now exhumed as extensional detachments in the “Otago Core Complex” (Forster and Lister, 2003).

Though there are no syn-rift sediments preserved in the hanging walls of the NE-SW faults that crosscut the strongly exhumed Otago Schists, the penetration of these faults in the crystalline crust together with their orientation suggest that they were indeed formed during the Late Cretaceous rifting. The same consideration applies to the Akatore Fault (# 2) that runs parallel to the Titri Fault along the coast, south of Dunedin.

Geometry and Kinematics of the Major Onshore Faults

Most onshore NE-SW faults in the exhumed Otago region dip 50º-60º NW. Close to the coastline the two principal faults are the Titri (# 1) and Akatore (# 2) Faults, both dipping 60º SE.

Dip angles cannot be established for many onshore faults. The available maps at scale 1:250,000 simplify field relationships and are not ideal to derive fault dip from topographic intersections. Also, dip angles measured at surface are not necessarily representative of dip angles at depth. The Tuapeka and Waihemo Faults dip away from the exhumed core of the Otago Schist antiform, the Tuapeka Fault (# 8) has segments dipping 30º SW.

The NE-SW normal faults were reactivated with reverse slip in the Miocene-Present stress field (Norris et al., 1990). There is also evidence (Forsyth, 2001; Upton et al., 2009) of reactivation in the same stress regime for faults sub-parallel to the Terrane boundaries (e.g. the Livingstone and Waitaki Faults).
Strike-slip offset is documented for the Waitaki (#32), Dansey Pass (#23), Awakino (#25), and Moonlight (#38) Faults, as well as for the Livingstone Fault (see Table 2).

Compressional inversion of inherited normal faults (indicated as INF in Table 2) is documented for the Titri (#1), Waihemo (#21 and #30), Moonlight (#38), Ahuriri (#37) and Tuapeka (#8 and #42) Faults, as well as for the Otematata (#28) and Wharekuri (#26) Faults, that crosscut the Waitaki Fault. Inversion of the Ostler Fault (#36) is suggested by Ghisetti et al. (2007).

Cross-Cutting Relationships

Deformation is strongly compartmentalised within regions characterised by distinct lithology, mechanical strength, and fabric, inherited from the lithostratigraphy and tectonic setting of the Terranes.

A clear example is the dominance of the NE-SW faults within the area of the Caples Terrane and Otago Schists, bounded by the Waihemo Fault to the north and by the Livingstone Fault to the south. South of the Dun Mountain-Maitai belt, the onshore structural setting is dominated by the Southland Syncline in the Murihiku Terrane. However, along the coastline, the active Settlement Fault (#9) and other sub-parallel faults suggest that the structural trend of the Akatore and Titri Faults is also present south of the Dun Mountain-Maitai belt. The N-S and NW-SE active faults of the Tekapo, Timaru and Canterbury regions do not propagate south across the active Waitaki Fault.

Overall, the NE-SW and NW-SE fault segments display ambiguous and non unique crosscutting relationships, probably a consequence of their coeval activity during the syn-rift stages and the later compressional stages.

3.2 Offshore Structural Map
Plate 1 shows the largest and most continuous fault systems that control significant vertical throw of the top basement, and is based on the map of the top basement interpreted by OMV using a large data set of seismic lines.

For a number of faults it is possible to quantify the orientation, dip direction, dip angle, and the amount of vertical throw (Table 3). Additional analysis of selected seismic lines makes it possible to identify the Neogene compressional inversion of normal faults originated during Late Cretaceous rifting, based on the partial or total annihilation of the original normal throw in the top basement, as indicated by different symbols in Plate 1.

Plate 1 also displays the most relevant cross-cutting relationships between faults parallel to the Terrane boundaries and the syn-rift faults, and shows the correlations between onshore and offshore domains.

The offshore map shows the strong segmentation imparted by the Terrane boundaries, especially along the belt of the Dun Mountain-Maitai Terrane. The syn-rift faults are characterised by distinct en échelon geometry, with short, overlapping segments that define fault systems continuous for tens of kilometers, and distributed in wide belts of localised deformation bounding syn-rift depocentres.

The separation of faults into orientation classes shows two principal distributions: the first one comprises faults oriented 021º-060º; the second one faults oriented 061º-100º. The second group has the orientation of the Sisters Shear zone (# O16-O19). However, the largest vertical throw of the top basement (shaded in light blue in Plate 1) is localised in a rhomboidal graben, bounded by both NE-SW and WNW-ESE faults, the latter sub-parallel to the Terrane boundaries. This setting is consistent with activity of sets of sub-orthogonal faults during rifting, as already discussed for the onshore regions (see also Table 1).

The inherited normal faults that underwent compressional reactivation are prevalingly distributed along the western margin of the Great South Basins, and are also documented farther north at the western margin of the Canterbury Basin (OMV data).
In the study area the belt of compressional reactivation is displaced farther offshore south of the Tara 1-Toroa 1 wells, at the Brook Street-Median Batholith Terrane boundary (faults # O21 and # O32). However, a number of compressionally inverted faults are scattered across the basin, and some (e.g. faults # O36 and # O37) are developed at the margins of the Kawau 1 structural high.

As already discussed for the onshore faults, there is evidence of compressional reactivation of WNW-ESE faults along the Terrane boundaries, especially along the Dun Mountain-Maitai-Murihiku and Brook Street-Median Batholith boundaries. Compressional inversion is also documented for the Waihemo Fault system, west of Galleon 1.

### 3.3 Structural Domains

All the features described above allow the distinction of adjacent tectonic domains, each one possessing characteristic structural fabric.

The domains distinguished in Plate 1 are:

**Domain A**

It corresponds with the areas of the Brook Street and Murihiku Terranes, and is bounded by the Hillfoot Fault to the north. NE-SW crosscutting faults are seemingly lacking onshore, but some structures may be buried underneath the extensive alluvial covers. The overall paucity of faults in this domain possibly reflects the strong mechanical anisotropy imposed by the WNW-ESE folds of the Southland Synclise, with shear largely accommodated by bedding-parallel slip. Offshore, a number of basement ridges are correlated with sets of short wavelength folds, truncated by faults trending 041°-060° and 061°-080°.

**Domain B**

It is bounded by the Livingstone Fault to the south and by the Waihemo Fault to the
north. It corresponds with the areas of the Caples Terrane and Otago Schists, deformed onshore by an antiformal dome with N-S to WNW-ESE axial trace, that plunges eastward close to the coastline. Exhumation of crustal rocks in a large-scale extensional core complex (Forster and Lister, 2003) is supported by the limb-parallel dip of the low-angle shear zones that bound the uplifted dome of the Otago Schists, and by the already mentioned preservation of Late Cretaceous syn-rift sediments at the margins of the dome. One of the bounding faults is the Tuapeka Fault (# 7) whose offshore continuity is suggested by morphological scarps in the top basement. However, the fault appears to die against the Dun Mountain-Maitai belt. Domain B is characterised by closely spaced sets of NE-SW faults, formed during Late Cretaceous rifting and reactivated in compression. The vast majority of onshore faults dip NW, whereas most offshore faults dip SE. The hinge between NW-dipping and SE-dipping faults runs along the coastline, bounded by the SE dipping Titri and Akatore Faults onshore (# 1 and # 2), and by the Takapu Fault offshore (# O1). The closely-spaced set of NE-SW faults in domain B is segmented at the intersection with the Dun Mountain-Maitai belt, and south of it faults belonging to the same directional class are more widely spaced and more discontinuous.

Domain C

It coincides with the narrow belt bounded by the Waihemo Fault to the south and by the Waitaki Fault to the north, with strong overprinting of the NW-SE brittle fabrics imparted by these two regional faults. A complex kinematics with strike-slip and reverse components is documented for the Waitaki Fault, inferred to be active in the present stress field.

Domain D

It is defined north of the Waitaki Fault. It extends west as far as the Ostler Fault (# 36). It is characterised by compressional reactivation of inherited, high-angle, N-S normal faults, and by the propagation of range-front thrust faults, all W-dipping. Dominant tectonic trends are N-S and NE-SW, but NNW-SSE faults are regionally represented by the active Waimate-Hunter thrust faults (# 34, 48 and 49), continuous over a length >100 km. Most faults within this domain are active, and the discontinuous traces mapped in the alluvial cover are indicative of buried active faults in the basement (e.g.
Irishman Creek and Tekapo Fault, # 43 and 44, and Orari-Montalto thrust, # 46).

**Domain E**

It is defined only offshore, south of domain A and south of the Murihiku-Median Batholith Terrane boundary. The structural fabric is characterised by prevailing ENE-WSW faults (Sisters Shear Zone), with vertical throw of top basement ≥3 km close to Stewart Island. Seemingly, the ENE-WSW faults truncate the faults of the NNE-SSW system, consistently with activity of the Sisters Shear Zone in the final stages of rifting. However, the ENE-WSW faults do not crosscut the Hillfoot Fault, and they are not present in Domain B; or in any of the onshore domains. Closely spaced sets of 041º-060º faults dipping both south-east and north-west are mapped around the Pakaha 1 and Pukaki 1 wells, and do not characterise the whole area of Domain E; however this difference may be only apparent and caused by paucity of data.
4. **Structural assemblages in the seismic lines used for the regional transects**

*Plate 2* provides the sequence stratigraphy nomenclature used by OMV in relation to the New Zealand and international time scale, and in relation to the litho-stratigraphy defined by previous studies. The chronological sequence of the regional tectonic and igneous events is also provided.

The stratigraphic sequences distinguished in *Plate 2* are grouped in super sequences that were deposited during episodes of: (1) syn-rift extension (112-82 Ma, with Syn-Rift Stage 1: 112-104 Ma, Syn-Rift Stage 2: 104-89 Ma, and Syn-Rift Stage 3: 89-82 Ma); (2) post-rift subsidence (82-19 Ma, with Post-Rift Stage 1: 82-61 Ma, Post-Rift Stage 2: 61-36 Ma, and Post-Rift Stage 3: 36-19 Ma); and, (3) syn-orogenic shortening (19-0 Ma).

The chronology of these stratigraphic sequences largely reflects the age of processes consequent on plate tectonic dynamics; however, propagation of convergent deformation towards the distal regions of the GSB was diachronous (Norris et al., 1990), and postdates the onset of the convergent plate margin at 25-20 Ma (King, 2000; Sutherland et al., 2000; Schellart et al., 2006).

The regional tectonic events that have shaped the geometry and stratigraphic evolution of the Great South Basin (GSB) through time are characterised by structural assemblages that formed in response to the regional stress field, the intensity of deformation, and the mechanical response of the involved rock units.

In order to identify the geometry of the structures formed in the GSB and the sequence of superposed deformation, a large number of seismic lines were analysed for this study. Particular care has been devoted to: (1) reconstruct the structural evolution of the GSB through time; (2) understand the control exerted by tectonic events on sedimentation, and, (3) identify plausible mechanisms of deformation. All these
elements are necessary for the correct interpretation and construction of the regional geological transects (Paragraph 5) and for their restoration (Paragraph 6).

Only a few, selected examples are presented in this report (Enclosure 1), and are briefly described below.

4.1 Syn-Rift Structural Assemblages

The syn-rift structures are the most prominent structures of the GSB, in terms of lateral extent, amount of displacement, and control on sedimentation. The geometry of the major offshore normal faults is shown in Plate 1 (see also Table 3), and significant examples of the extensional structural style are provided in Enclosure 1 (examples 1-7).

Major features shown by these examples are:

- Cross-cutting of the basement by systems of conjugate normal faults dipping NW and SE (examples 1, 2, 4, 5).

- Presence of fault-bounded blocks displaying one dominant sets of similarly dipping normal faults, compatible with mechanisms of domino rotation (examples 2, 3, 7).

- Predominant dips \( \leq 30° \) for the majority of normal faults.

- Presence of closely spaced faults (1-2 km apart), but faults with largest vertical throws are spaced 10-20 km apart (examples 1, 4).

- Seemingly, the largest components of vertical displacement were accumulated during the earliest syn-rift stages, during deposition of Sequences K20 and K30 (examples 1, 2, 4, 5, 6).
• Most faults are onlapped by top K50, but some faults propagate across the post-rift sequence, as in the region of the Toroa Anticline (example 4).

• Vertical throws along the largest normal faults are of the order of 1-2 km, but in some cases individual faults have throws \( \leq 4-6 \) km (example 4).

• Rotation of hanging wall panels is compatible with listric geometry for the largest normal faults (examples 2, 3, 5, 6). On some seismic lines there is evidence for listric geometry of major faults, but this evidence is limited to modern, long-offset seismic acquired in optimal weather conditions (OMV data). In general, seismic data do not image sufficiently deep to confirm the presence of intra-basement, sub-horizontal detachments.

• Syn-rift sequences K20 and K30 are seemingly thick, but the poor resolution of the seismic lines in the deepest syn-rift units generally precludes accurate mapping.

• The seismic lines chosen for the regional transects are parallel and/or oblique to Terrane boundaries and do not provide a clear image of their role during the rifting. The analysis of additional, more suitably oriented seismic lines shows that the boundaries were indeed reactivated, but also suggests that new faults were formed parallel or slightly oblique to the inherited basement boundaries. Strike-slip components parallel to the boundaries cannot be proven or excluded.

4.2 Post-Rift Structural Assemblages

The geometry of the GSB during the long time period of post-rift sagging (82 to 19 Ma) is characterised by dominant differential subsidence and compaction, associated with tilting and draping above inherited structural highs.

Some characteristic settings of the post-rift units are shown in Enclosure 1 (example 8).
Mild, synsedimentary deformation affects the late Paleocene to Eocene sequences, as mostly documented by disharmonic, short-wavelength folding of top T10, that does not propagate into the overlying sequences (short arrows in examples 8 and 11). This structural style may well be characteristic of the organic-rich shales of the Tartan Formation (cf. Cook et al., 1993), and associated with differential subsidence and localised overpressuring above structural highs and lows of the basement.

The vast majority of normal faults active during the syn-rift stages did not continue their activity, with the notable exception of the faults at the western margin of the GSB. Along this margin, especially in the region crossed by the Tara 1 well, there is evidence of mild shortening within the Eocene sequence, testified by short wavelength folding above basement faults, with development of a localised structural high that controlled syn-depositional thickness variations in the post-rift sequence (structure marked by arrow in example 9). Similar settings characterise the Kawau 1 region (OMV data). Note that the offset in the basement faults (cutting up-section through top K70) remains normal.

Example 8 shows two major regional structures that affect the upper post-rift sequences:

- **Mass Transport Complex (MTC)**
- **Eastern Imbricate Belt (EIB)**

The **Mass Transport Complex (MTC)** is a regional scale gravitational instability. The basal surface of the MTC starts in the uppermost Sequence T70 (at the western shelf margin), and propagates down-section within T70 (examples 10 and 11), down to the boundary with T60. 

*MTC* is generally not recognised on the elevated eastern margin of the basin, where the post-rift sequences are thin and condensed.

Gravitational instability is associated with components of sliding from the elevated western margin into the central basin, but south to north components of motion must be significant as well. In fact, the removal of portions of Sequence T70 away from the
plane of the sections is necessary to create the sedimentary gaps filled by Sequences T80 and T90 in the hanging wall of the MTC (examples 10 and 11). The largest components of sliding are contemporaneous with deposition of Sequence T80. Sequences T90 and T100 onlap the body of the MTC along the western margin of the basin.

The *Eastern Imbricate Belt* (*EIB*) consists of a flexure localised at the transition between the thick post-rift sequences and the condensed sequences of the eastern structural highs (example 8).

Deformation consists in the eastward monoclinal tilting of Sequence T70, together with the basal surface of the MTC (examples 12, 14). The flexure defines an asymmetric anticlinal culmination; the flexed limb is onlapped by sediments of Sequence T80, that fill the small sedimentary basin created above the flexed limb (examples 12, 13, 14).

The *EIB* is detached within T70, with flexural slip accommodated by layer-parallel or oblique faults detected in a few seismic lines.

The *EIB* was formed after initial sliding of the MTC, but over the same time interval constrained by the onlap of Sequence T80 above both structures. In some seismic lines the MTC appears to have been further remobilised at a later stage, and truncates the frontal fold of the *EIB*.

### 4.3 Syn-Orogenic Structural Assemblages

Compressional deformation within the GSB is of relatively young age (post early Miocene), and the resulting shortening is mild.

Structures associated with these late deformations are inherited normal faults inverted in compression and large-scale folds. These structures have an orientation compatible with their formation in the late Miocene-present stress field, and are superposed onto
the structures that originated at an earlier stage by tilting and gravitational detachment.

**Normal Faults Inverted in Compression**

Reactivation of inherited normal faults by reversal of their displacement in a compressional regime (positive inversion) is mostly evident along the western margin of the Canterbury and Great South basins, in a belt parallel to the coastline. However, there are examples of compressional inversion (but not in the lines used for the regional transect) of other normal faults in the central region of the GSB that have been inverted in compression, especially in the regions of the Rakiura 1-Kawau 1 structural highs. Late stage compressional inversion appears also to affect basement faults beneath the EIB, as shown in example 13.

The inverted normal faults trend N-S to NNE-SSW; thus they belong to the same system that was reactivated in the Otago region (Domain B, **Plate 1**). The majority of onshore faults dip west, but along the coast and offshore the faults generally dip 50°-60° E and SE.

A typical feature of reactivated, inherited normal faults is the closure and reversal of the normal offset in the basement, coupled with reverse faulting and localised, short wavelength folding of the syn-rift sequences originally deposited in the hanging wall of the normal faults (see also Figure 4). The sequences deposited during active shortening onlap the hanging wall folds, and fill the sedimentary basins formed by progressive foundering of the fault footwall during accumulation of reverse displacements.

Some of these features are shown by the faults at the western margin of the GSB (faults marked by arrows in examples 15, 16 17), but in some cases the offset in the basement remains normal, indicating that the amount of compressional inversion is small, and unable to obliterate or reverse the previously acquired normal displacement. In other cases (Takapu 1 Fault in example 15) there is no evidence for a syn-rift basin in the fault hanging wall. It cannot be excluded that some faults that display reverse offset are new structures that propagated in the late Miocene-present stress field. Eocene sediments are folded and truncated at seabed, consistent with very recent movement.
identified on the Akatore and Titri Faults (Litchfield, 2001). The Takapu 1 well also has an anomalously high bottom hole temperature and geothermal gradient (Funnell and Allis, 1996), consistent with thermal disequilibrium arising from very recent uplift and erosion.

Compressional deformation is post T90, and in some cases the sea bottom appears to be deformed as well. However, the west margin of the GSB close to the coastline is poorly imaged by the seismic lines, and in most cases the structures are masked by strong seismic multiples caused by the shallow water depth.

*Toroa Anticline*

Large-scale folding of the Paleogene sequence occurs along the Toroa Anticline (example 18).

The axial trace of the Toroa Anticline extends for a length of ~170 km, with trend 030°, and plunge to both the N and S. At its northern and southern plunge, the axial trace of the fold terminates against Terrane boundaries (Hillfoot Fault to the north and boundary between the Median Batholith and the Western Province to the south, see Plate 1).

The fold is broad and symmetric, with a gently arched, circular hinge (example 18). Longitudinal lines across the fold reveal similar characteristics. Folding is defined by the structural elevation of the post-rift sequences; the base of the MTC and the sea bottom are folded concordantly. Fold amplitude decreases below top T10. The fold is an isolated feature, and there are no paired synclines.

The fold is located above the area of maximum syn-rift stretching and foundering of the basement. Syn-rift faults beneath the core of the fold appear to have propagated up-section by growth of poorly interconnected fault segments. However, the components of displacement in the basement, syn-rift and post rift sequences are all normal (example 18).
5. Regional transects

The geometry and sequence of deformation events in the GSB within the areas of PEP 50119, PEP 50120 and PEP 50121 are summarised by five regional transects transverse to the structural trends, that tie the offshore and onshore geology and incorporate data from selected seismic lines and exploration wells in the offshore and geological maps in the onshore.

The location of these transects is provided in Plate 1; the legend is in Plate 3. Transects T01 to T05 are shown in Plates 4, 5, 6, 7, 8 respectively.

Note that in Plates 4-8 the geological transects are displayed with a vertical exaggeration (VE) of 8:1, in order to emphasise subtle geometric features arising from mild deformation. The diagram reproduced in each transect facilitates the conversion from apparent to real angles of dip consequent on VE=8:1.

The onshore geology is based on the geo-referenced data base of the geological maps (scale 1:250,000) Wakatipu (Turnbull, 2000), Dunedin (Bishop and Turnbull, 1996), and Murihiku (Turnbull and Allibone, 2003). The topographic profile is plotted from the DEM provided by OMV.

The offshore geology is based on a set of suitable seismic lines, depth-converted by OMV, using the velocity model established for the GSB.

OMV provided the interpretation and mapping of the seismic stratigraphy horizons in the depth converted seismic lines that were used for the geological transects. The key stratigraphic horizons chosen to illustrate the geological setting and the progressive evolution of the GSB are (from top to bottom): top T90 (19 Ma), top T80 (27 Ma), top T70 (36 Ma), top T60 (46 Ma), top T10 (57 Ma), top K100 (65 Ma), top K70 (77 Ma), top K50 (82 Ma), top K40 (89 Ma), top K20 (only for transect T04), top K00 (top basement). Refer to Plate 2 for the chrono-stratigraphy of these horizons. The regional
transects also include the basal surface of the *Mass Transport Complex (MTC)*, a gravity slide detached within the T70 Sequence.

All the available data have been re-interpreted and assembled within a structurally homogeneous and tectonically consistent framework, and incorporated into the geologic interpretation of the regional transects.

All transects are interpreted to depth of 11 km, but seismic data generally provide constraints to only 6 km below sea level, whereas onshore geology provides constraints for elevations ≤2 km above sea level. Thus, structural setting of the deep basement is largely unconstrained.

### 5.1 Transect T01 ([Plate 4](#))

T01 extends with average orientation 127° for a total length of ~380 km. It extends from the Moonlight Fault (2146031.1 E 5560012.0 N, all reference points in NZTM2000) to the coastline at Taieri River mouth (2294142.7 E 5458497.0 N). The onshore portion of this transect crosses the Caples Terrane across the Wakatipu (Turnbull, 2000), Murihiku (Turnbull and Allibone, 2003) and Dunedin (Bishop and Turnbull, 1996) Qmaps. The south-eastern end-point of T01 is offshore at 2454028.4 E 5346477.1 N. Offshore interpretation is tied to the Takapu-1 well (2311585.8 E 5446708.1 N), and based on seismic lines 70_h_b and DUN_06_03_R08.

The whole transect is within the Caples Terrane (undifferentiated), with low-grade metamorphic units (dominant volcaniclastic sandstones) outcropping onshore. The complex regional folding of the units is sketched by the foliation formlines in outcrop. The seismic lines do not provide enough resolution for sketching the foliation geometry within the offshore basement units. The transition to the Otago Schists is likely to occur at depths ≥10 km, but it is not represented.

The exhumed and uplifted units of the Caples Terrane are crosscut by the set of NE-SW
oriented reverse faults of Domain B (Plate 1). Some of these faults are ostensibly inherited from the Late Cretaceous rifting and inverted in compression since the Late Miocene, with evidence for activity in the present stress field (e.g. Nevis Fault, # 39 in Table 1). The crustal penetration and the geometry of the faults at depth are unconstrained. It is conceivable that the longest fault systems extend at ~10 km depth, but there is no evidence for the presence and/or position of an intra-basement detachment.

Strong components of post Miocene uplift have resulted in the deep dissection of the sedimentary units deposited above the basement. However, the presence of inliers of shallow water Oligocene limestones (Bobs Cove limestones, cf. Turnbull et al., 1975) in the footwall of the Moonlight Fault (at the western end of the transect), and the preservation of the fault-dissected Waipounamu Erosional Surface (WES) all along the transect (from the Dunedin coastline to the shores of Lake Wakatipu) provide two near sea-level sub-horizontal markers that constrain the transgression of the Oligocene marine sediments (see paragraph 6).

Close to the coast T01 crosses the active Titri Fault that dips steeply (~65°) to the east (Litchfield, 2001). The Titri Fault is a clear example of a Late Cretaceous normal fault with original normal offsets of 2500-3000 m that was inverted in compression since the late Miocene. Its early activity as a normal fault is documented by deposition of the Henley Breccia in the fault hanging wall. The Henley Breccia is a non marine fanglomerate of early Late Cretaceous age (Raukumara Series), and has been correlated with the offshore Sequence K40 in this transect. Along the coastline the transect crosses the active Akatore Fault, that is parallel to the Titri Fault, but has no sediments indicative of its Late Cretaceous formation.

The shortened, tilted and uplifted western margin of the GSB is well defined in the panels between the Titri, Akatore and Takapu I Faults, all displaying net reverse offset. Along the western GSB margin crossed by this transect, the extremely thin (or absent) syn-rift sequences overlying the basement are onlapped and transgressed by the T00 Sequence (Teurian, early Paleocene) and by overlying Eocene- Oligocene marine sequences (T20 to T80). These deposits, truncated by the set of NE-SW reverse faults,
are preserved in the coastal Dunedin region and crop out in the footwall of the Titri Fault. This setting shows that the western margin of the Late Cretaceous rift was affected by subsidence and marine transgression during the post-rift stages, and again tilted and uplifted during the stages of Miocene to present syn-orogenic shortening. The abundant clastic supply consequent on hinterland uplift is documented by the Miocene to Plio-Quaternary delta, prograding over the continental shelf.

The syn-rift basins along transect T01 are bounded by sets of conjugate normal faults, with throws \( \leq 2 \) km. The faults are represented as maintaining the same dip angles (ranging from \( \sim 30^\circ \) to \( 50^\circ \)) down to depth of 11 km, but there are no constraints for this interpretation.

The deepest syn-rift basins (at depth of \( \sim 6 \) km) are filled by Sequences K30-K50; the same sequences are substantially thinner (or absent) above the intervening structural highs. The presence of the older Sequence K20 above the basement (Syn-Rift Stage 1) is likely, but unconstrained by the seismic data. Between km 260-320 volcanic units are present in the central rift area.

The largest fault throws controlling thick syn-sedimentary depocentres are recorded by the earliest syn rift phases (1? and 2), whereas Sequence K50 (Syn-Rift Stage 3) drapes the structural highs, and generally displays less marked thickness changes within the fault-bounded depocentres relative to the underlying sequences.

The infilling of the syn-rift basins become thinner at the eastern end of transect T01 (km 340-360), but the eastern margin of the rift is not represented in this transect because of the lack of appropriate seismic lines.

Post-rift Sequences K60 to T60 are thicker between km 220-320, and thin laterally towards the western and eastern margins of the basin. The most marked thickness change is at \( \sim \)km 320, where the package of Sequences T20-T60 pinches out and closes laterally.

The post-rift Sequence T70 is crossed by the basal surface of the MTC that propagates
within the upper part of this sequence from the shelf margin to ~km 270, and cuts down-section to propagate along the T60-T70 boundary from km 270 to 340, where it eventually dies out. The \textit{MTC} is not mapped at the south-eastern end of T01.

The gravitational movements of the \textit{MTC} are likely responsible for the transport out from the section of packages of Sequence T70 that is strongly channeled at km 270 and 300, with the channels filled by the overlying Sequence T80. At km 270 Sequence T80 appears to be affected by the gravitational instability as well.

The uppermost post-rift package is strongly dissected by sets of closely spaced fractures and/or low-displacement faults that are concentrated in the Sequences T70 and T80.

Note also the presence of an isolated volcanic cone within the Plio-Pleistocene sequence at km 340. The feeders are not imaged in the seismic line.

\textbf{5.2 Transect T02 (Plate 5)}

T02 extends with average orientation 128° for a total length of ~440 km. It extends from the Letham Thrust (2112209.7 E 5491637.4 N, all reference points in NZTM2000) to the coastline near Long Point (2248846.4 E 5399568.5 N). The onshore portion of this transect is entirely within the Murihiku Qmap (Turnbull and Allibone, 2003) and is comprised within Domain A (Plate 1). The transect crosses the Brook Street-Murihiku Terrane boundary along the N-S trending Letham Thrust, and extends in the Murihiku Terrane. The boundary between the Murihiku and Dun Mountain-Maitai Terranes is offshore at ~km 300. The south-eastern end-point is in the offshore at 2473780.8 E 5242536.5 N. T02 intersects T03 at 2231860.4 E 5411010.2 N. Offshore interpretation is based on the seismic lines 71_h_a_28r and DUN_06_010_R08.

The Letham Thrust at the north-western margin of the Transect is interpreted as the
major fault boundary (dipping ~20° SE) that superposes the Murihiku Terrane above the Brook Street Terrane from the surface to depth ≤10 km. However, it is conceivable that the geometry is much more complex, and that the boundary fault flattens to shallow dips at depth ≥10 km (see Mortimer et al., 2002).

The several formations distinguished within the Murihiku Terrane (cf. Turnbull and Allibone, 2003) are not differentiated along this transect, and the simplified geometry of the units is sketched through the bedding formlines. T02 runs sub-parallel to the axial trace of the Southland Syncline; thus it does not show the geometry of this large-scale structure. The superposition of the Dun Mountain-Maitai Terrane above the Murihiku Terrane occurs along the NE-dipping Hillfoot Fault onshore (Plate 1). Offshore, this boundary is marked by a strong morphological and elevation contrast in the top basement, and appears to have been reactivated during the syn-rift extensional episodes. However, the precise location of the Terrane boundary is not clearly identifiable along the DUN_06_010_R08 seismic line. The internal geometry of the Murihiku and Dun Mountain-Maitai Terranes in the offshore is not constrained as well.

Most faults cutting through the Murihiku units onshore are discontinuous, and represented as rooted at depth ≤4-6 km, with the exception of the fault systems close to the coastline, likely to be part of the Titri-Akatore Fault system. These faults, together with the sub-parallel offshore faults (at km 180-200 along T02) control the uplifted and inverted west margin margin of the GSB; their geometry is similar to that already described for T01. Compressional inversion is clear for the faults at km 200, with almost complete closure of the normal throw in the basement, and folding of the post rift sequences and of the basal surface of the MTC. The syn-rift sequences are lacking along the inverted west margin, the post-rift Sequences T00-T10, T20-T60 and T70 pinch-out and close in the proximity of the present coastline, whereas Sequence T80 is preserved in down-faulted onshore panels (km 80-100), where it overlies Eocene terrestrial coal measures, and is overlain by terrestrial Miocene to Plio-Quaternary deposits.

The architecture of the rifted basement is characterised by dominant E-dipping faults in the Murihiku Terrane, and by sets of conjugate faults in the Dun Mountain-Maitai
Terrane. The deepest syn-rift depocentres are bounded by faults with $\leq 4$ km of vertical displacement and are separated by horsts that maintained their role of structural highs during the whole rifting stages. As for transect T01, the presence of the earliest syn-rift Sequence K20 is unproven but likely, and the largest normal offsets were reached during deposition of Sequences K30 and K40. The thickness of the sequences infilling the deepest depocentres is of the order of 2-4 km, but the top of basement in the Dun Mountain-Maitai Terrane is 2-4 km higher than in the Murihiku Terrane.

As already discussed for T01, there is no clear evidence for intra-basement detachments or listric geometry of the syn-rift faults that are projected with average dips of $\sim 30^\circ$ down to depth of 11 km. This transect does not extend to the eastern shoulder of the rift, because of lack of data.

The post-rift sequences are thicker from km 200 to km 320 and become thinner at the south-eastern end of this transect. Sequences T20-T60 do not close laterally as in T01. A sill is mapped within T60 at $\sim$km 230.

The basal surface of the $MTC$ is traced within Sequence T70 from km 200 to 260, and cuts down-section at the boundary of Sequences T70-T60 from km 260 to 380. To the SE there is no evidence for gravitational sliding. In contrast with T01, there is no obvious removal of portions of Sequence T70 out from the section, whereas a deep channel filled by Sequence T90 erodes Sequence T80 at km 260-270.

As already noted for the previous transect, Sequences T70 and T80 are crossed by sets of closely-spaced fractures and/or low displacement normal faults.

**5.3 Transect T03 (Plate 6)**

T03 extends with average orientation $140^\circ$ for a total length of $\sim 540$ km. It extends from 2151553.5 E 5493407.4 N (all reference points in NZTM2000) to the coastline near Tahakopa Bay at 2242581.3 E 5400229.5 N. The onshore portion of this transect is
entirely within the Murihiku Qmap (Turnbull and Allibone, 2003) and within Domain A ([Plate 1](#)). At its north-western end this transect intersects with sub-parallel trajectory the Livingstone Fault (Caples-Dun Mountain-Maitai Terrane Boundary). At km 40 the transect cuts obliquely through the Hillfoot Fault (Dun Mountain-Maitai-Murihiku Terrane boundary), and remains in Murihiku Terrane for its remaining length. The south-eastern end-point is in the offshore at 2503342.2 E 5113115.6 N. Offshore interpretation is based on the reprocesses seismic line AWI2003_0013, originally acquired by the A. Wegener Institute and GNS Science (Grobys et al., 2009).

The internal geometry of the Dun Mountain-Maitai and Murihiku Terranes is extremely simplified, and shown only through bedding formlines. There is no differentiation of the units. The geometry of the Southland Syncline is better depicted in this transect because of its oblique orientation (rather than sub-parallel, as in T02) relative to the fold axial trace.

Onshore there are no constraints for the geometry and extent of the faults at depth. The Hillfoot Fault is hypothetically depicted as sub-vertical down to 11 km depth. A set of NE-SW oriented, closely spaced faults controls the west margin of the GSB along this transect, but – in contrast with the previous transects T01 and T02 - there are no markers indicative of compressional inversion for the faults close to the coastline (km 110-130). However, compressional inversion is suggested for faults farther offshore (km 180-200), consistent with the net reverse offset in the basement and minor folding of the cover sequences. Sequences T00-T10, T20-T60 and T70 close laterally above the Murihiku basement close to the coastline. Outcrops of shallow-water Oligocene marine deposits (Sequence T80) are preserved onshore where they overlie the Murihiku and Dun Mountain-Maitai units and are overlain by Miocene to Plio-Quaternary terrestrial deposits.

The geometry of the rifted basin along this transect is characterised by deep depocentres filled by 2-4 km of Sequences K30 and K40 (K20 unproven, but likely), bounded by sets of conjugate normal faults with vertical throws $\leq 4$ km. The top basement is down to depth of $\sim 8$ km in the central basin (km 240-260) and becomes progressively shallower to the SE. From km 440 to 540 the geometry of the Murihiku
Terrane is dominated by structural highs with intervening narrow grabens. The top basement is shallow (~2 km below sea level), and no syn-rift sequences were deposited above the structural highs.

As already observed for T01 and T02, the post-rift sequences are extremely thick to the NW (from km 200 to 360) and become substantially thinner to the SE. Thinning is associated with the closure of Sequences T20-T60 at ~km 360.

The transition from thick to thin post-rift units is also marked by the development of the fold of the *Eastern Imbricate Belt* (EIB) within Sequence T70. The fold is defined by a monoclinal flexure at ~km 310, onlapped by Sequence T80. Sequence T80 is thicker at the front of the EIB, consistent with its deposition during formation of the frontal flexure.

The basal surface of the *MTC* is mapped with continuity from the near offshore to ~km 310 in the upper part of Sequence T70. The base of the *MTC* is folded by the EIB, and propagates down-section at the boundary between Sequences T70 and T60 between km 330-360, where it dies out together with Sequences T20-T60.

There is a substantial thinning of Sequence T70 between km 310-360, possibly indicative of gravitational transport out from the section.

### 5.4 Transect T04 ([Plate 7](#))

It extends with average orientation 128° for a total length of ~334 km. This transect is entirely offshore and extends from 2187930.6 E 5345213.8 N to 2459019.3 E 5149909.8 N (all reference points in NZTM2000). This transect is tied to the projected wells Tara-1 (2219886.8 E 5314585.7 N) and Toroa-1 (2244597.1 E 5301452.0 N). Interpretation is based on seismic lines OMV_08_081 and 72_h_f_14x_R96.

At ~km 60 this transect intersects the boundary between the Median Batholith and the
Murihiku Terrane. Though the Murihiku Terrane is characterised by a rather distinct seismic facies, the actual position of the boundary remains uncertain, especially because Late Cretaceous normal faults appear to dissect and displace the original Terrane boundary.

The syn-rift geometry is characterised by faults with large vertical throws that bound narrow horsts relative to wide basins (e.g. faults at ~km 40 with inferred throws ≤8 km); average vertical throws of the largest faults are of the order of 2-4 km. Faults are in conjugate sets, but some syn-rift basins have the geometry of half grabens (e.g. basins at km 20-40, and km 120-160).

In contrast with the previous transects, the quality of the seismic data in the lowermost syn-rift allows the mapping of Sequence K20 in the western-central part of this transect. This sequence could also be present (but unproven) in the deep depocentre at km 200-210. The largest components of vertical throw are in the lowest syn-rift. The top basement is down to 8-10 km depth beneath the Tara 1 and Toroa 1 wells and becomes progressively shallower (~3 km) towards the south-eastern margin of the transect.

Some of the syn-rift normal faults propagate up-section in the post-rift sequences, as e.g. the faults at km 0, the faults beneath Tara 1 and Toroa 1 and the faults at km 160 and km 200. This setting and the associated folding of the post-rift sequences (see also example 9 in Enclosure 1) suggest some components of compressional reactivation. However, the faults maintain normal offset in the basement, in the syn-rift sequences, and in Sequences K60-K70 as well.

The post-rift sequences are thick in the region between Tara 1 and Toroa 1, with an abrupt change in thickness at ~km 180, where the structure of the EIB is developed. Sequences T20-T60 are much thinner, but continuous to the end of this transect.

The EIB is defined by the monoclinal flexure of Sequence T70, the last onlapped by Sequence T80. As already observed for T03, the largest thickness of Sequence T80 occurs at the frontal flexure of the EIB.
Transect T04 crosses the Toroa Anticline; the structural culmination is defined by the folding of Sequences T20-T60, T70, and T80. The sea bed is gently folded as well. Both the Toroa Anticline and the folds of the $EIB$ are detached within Sequence T10.

The Toroa Anticline and the $EIB$ deform the basal surface of the $MTC$. The basal surface propagates within Sequence T70 for its whole length, from km 40 to km 230. There is no obvious transport of portions of Sequence T70 away from this section.

As already observed for the other transects, sets of closely spaced fractures and/or low displacement faults crosscut Sequences T70 and T80.

### 5.5 Transect T05 (Plate 8)

T05 extends with average orientation 129°, for a total length of ~280 km. This transect crosses Stewart Island from 2106757.1 E 5339724.5 N to 2123809.3 E 5327556.5 N (all reference points in NZTM2000), and extends to the offshore point 2342734.9 E 5171729.8 N. The transect deviates to azimuth 133° between points 2203829.4 E 5274421.4 N and 2238000.7 E 5245522.1 N. Offshore interpretation is based on seismic lines 83unz_06_5, OMV08_086 and 72_h_b_6x. The wells Pakaha-1 (2252927.8 E 5212223.8 N) and Pukaki-1 (2298804.6 E 5179457.6 N) are approximately 18 km away from the line, and are not projected onto it.

The basement along the whole transect is the Median Batholith. This transect crosses the Sisters Shear Zone in the immediate offshore of Stewart Island. However, there are no seismic data to constrain the geometry of the fault and of its hanging wall, and interpretation is based on the outcrops in the Sisters Islets (see also Tulloch et al., 2006; Kula et al., 2009) and from extrapolation to the NW of the horizons mapped in line 83unz_06_5. The conglomerates in the Sisters Islets are of unknown age, and they are tentatively attributed to Sequences K30-K40 in the interpretation of T05. However, isopachs in the syn-rift (OMV data) suggest that activity of the Sisters Shear Zone...
continued during deposition of K50. The interpretation proposed in T05 limits the syn-rift vertical throw of the Sisters Shear Zone to ~4 km, but – as already stated – the depth of the basement in the fault hanging wall is unknown.

The syn-rift architecture of the basement along this transect is characterised by sets of conjugate faults with vertical throws ≤ 4 km, with discontinuous syn-rift basins separated by horsts with thin or absent cover of Sequences K30-K40 and K50. Some structural highs persisted throughout the entire rifting stages (e.g. horsts at km 40-80 and km 200-240). The basement is emergent or very shallow at the NW end of this transect, and is down-faulted to maximum depth of ~5 km at km 160-200. Unfortunately T05 does not depict the eastern margin of the rift, because of the lack of appropriate seismic data.

The post-rift sequences are thin at the NW end of this transect, and tilted to the SE. Note that they are truncated against the faults of the Sisters Shear Zone system. Thickening of the post-rift sequences is well marked for T20-60 and T70, with abrupt decrease in thickness at km 240, coincident with the development of the EIB. As already noted before, Sequence T80 thickens in front of the EIB and onlaps the tilted flank of the flexure.

The basal surface of the MTC is traced in the uppermost Sequence T70 from km 20 to km 240, and cuts down-section from km 240 to the end of the transect, where it is rooted along top T10.
6. Restoration and balancing of the regional transects

6.1 Goals and Methods

Techniques of restoration and balancing are used to validate geometry and sequence of superposed deformation in 2D cross sections based on field data and/or seismic data and wells.

A brief review of the restoration and balancing techniques used for this work, and a discussion of their limits of applications is provided in the following. Additional information can be found in Groshong (1999) and Tearpock and Bischke (2003), and in the references quoted therein.

Restorations are performed for both compressional (e.g. Dahlstrom, 1969; Hossack, 1979) and extensional (e.g. Gibbs, 1983; White et al., 1986; Rowan and Kliegfield, 1989; Schultz-Ela, 1992; Erickson et al., 2000) tectonic settings, but a necessary condition is that the totality of deformation (or at least the largest components) are comprised in the plane of the section. Thus, cross sections should be drawn parallel to the transport direction, as directly measured or – more often - assumed to be perpendicular to the strike of the regional structures. Components of lateral or oblique movements away from (or into) the plane section (e.g. in the case of strike-slip tectonics or oblique reactivation of pre-existing faults) may be accounted for if known, but they are generally unknown for most seismic interpretations, and may result in unaccountable deformations and imperfect restorations.

Restoration starts from the present deformed state and progressively removes superposed deformation in incremental steps that follow an inverse chronological order (i.e. from the youngest to the oldest). This procedure accounts for the reconstructed sequence of deformation phases, defines the structural elements active during each tectonic phase, and aims at restoring sedimentary units in their undeformed state, accounting for eventual syn-depositional thickness variations. Transformations remove the strain resulting from shortening (and/or extension) according to geometric rules that
characterise the observed or interpreted deformation mechanisms. There are no constraints relative to mechanical rheology and stresses.

The successful reconstruction of a realistic pre-deformation sedimentary multilayer (with acceptable loose line geometries) and the fit - without gaps or overlaps - between restored portions of the sections (usually bounded by major faults) provide a test of the geometric coherence and admissibility of the interpretation, but do not guarantee that the geological reconstruction is correct and/or unique.

For a restoration to be performed it is necessary to define the appropriate kinematic model governing formation and development of each structure. Commonly used mechanical models include: (1) rigid-body rotation and translation; (2) flexural slip (simple shear parallel to bedding); (3) fault-parallel slip; (4) vertical simple shear; and, (5) oblique simple shear (see Groshong, 1999).

Different mechanisms can be applied for the same cross section, and they eventually produce different restored geometries. This means that restoration in itself cannot constrain the real deformation path that has created the observed structures, and that the undeformed state can be reached with non unique solutions, all geometrically valid. Thus, in absence of additional data, the kinematic path during the geological history remains unknown.

All the limitations discussed above must be taken into consideration, but the methodology remains a valuable tool for testing the accuracy and consistency of geological reconstructions, for reducing geometric uncertainties in interpreted seismic lines, and for reconstructing the structural and synsedimentary history of deformation during migration and entrapment of hydrocarbons.

For these reasons, there are a few computer programs that facilitate the iterative (and time consuming) procedures of balancing and restoration. Computer-assisted analyses help in correcting balancing problems, evaluating the merits of alternative interpretations, and comparing the consequences of different deformation mechanisms during the geological history.
In this study all the geological interpretations and restorations have made use of the software LithoTect™ by Geo-Logic Systems (www.geologicsystems.com).

Incremental restoration has been applied to the regional transects T01 to T05 (Plates 4-8), but only to those portions where sedimentary layers can be restored to a pre-deformation datum. This is generally impossible for the onshore areas, where only the basement is exposed. The only exception applies to Sequence T80, deposited during the stages of Oligocene maximum marine highstand, and preserved in scattered outcrops onshore. The base of the marine transgression can also be traced thanks to the preserved Waipounamu Erosional Surface (WES) in T01. WES is time transgressive, but its farthest advancement inland was reached during the Oligocene marine transgression (LeMasurier and Landis, 1996).

Restoration of transects T01-T05 assumes a horizontal depositional datum for the restored stratigraphic boundaries, because of lack of exhaustive data relative to synsedimentary, paleobathymetric variations along the regional transects. This assumption is simplistic, but at the scale of the adopted reconstruction, paleobathymetric variations of hundred of metres (which is the likely order of magnitude) convert into variations of the datum of the order of millimeters, and thus have a minimal impact on the results.

Note also that though the western margin of the Late Cretaceous rift is included in the regional transects, the eastern margin is not, because none of the available seismic lines used for the geological interpretations extends to that margin. This means that the amount of overall extension cannot be estimated, especially if extension is not homogeneously distributed and concentrated at the rift margins.

The kinematic models used for the restorations include: (1) rigid-body rotation and translation for the basement rock units; (2) vertical shear and fault-parallel shear for removal of extensional deformation; and, (3) flexural shear for removal of folding.

Restoration has been performed in incremental steps, restoring and retro-fitting all the
regions separated by faults interpreted as active during each deformation stage. Minor misfits were corrected and accommodated during this process. No large misfit occurred during restoration.

6.2 Restored Regional Transects

Plates 9-13 reproduce the present geometry of the regional transects T01-T05 (Plates 4-8) and show the progressive retrodeformation steps (refer to Plates 2 and 3 for the chrono-stratigraphic legend): Note that Sequences T90-T110 have been backstripped and not restored. The faults considered to be active in each of the steps are represented in bold (in red if active with reverse mechanisms). In order to emphasise the geometry resulting from deformation, the sections in all restoration steps are shown with vertical exaggeration VE= 8:1. However, restoration uses the real angles of dip, and there is no distortion consequent on vertical exaggeration.

The restoration steps are the following:

Step 1 – Restoration to the horizontal of top T80 and of Sequence T80 (27-36 Ma) – Post-Rift Stage 3.

Step 2 – Restoration to the horizontal of top T70 and of Sequence T70 (36-46 Ma) – Post-Rift Stage 2.

Step 3 – Restoration to the horizontal of top T60 and of Sequences T60-T20 (46-57 Ma) – Post-Rift Stage 2.

Step 4 – Restoration to the horizontal of top T10 and of Sequences T10-T00 (57-65 Ma) – Post-Rift Stage 2.

Step 5 – Restoration to the horizontal of top K100 and of Sequences K100-K80 (65-77 Ma) - Post-Rift Stage 1.
Step 6 – Restoration to the horizontal of top K70 and of Sequences K70-K60 (77-82 Ma) – Post-Rift Stage 1.

Step 7 – Restoration to the horizontal of top K50 and of Sequence K50 (82-89 Ma) – Syn-Rift Stage 3.

Step 8 – Restoration to the horizontal of top K40 and of Sequences K40-K20 (89-112 Ma). Syn-Rift Stage 2.

Step 9 – Restoration of top basement to pre-rift setting, and evaluation of the stretching factor $\beta$, measured as the ratio between the final length within the sampled rifted region (L_f, measured at the end of the Syn-Rift Stage 3) and the initial length before extension (L_o), measured in the restored basement.

Some general remarks on the restored transects (Plates 9-13) are provided in the following.

*Restoration of Transect T01* (Plate 9)

Transect T01 is the only transect for which it is possible to reconstruct the geometry of the restored top T80 far inland, by: (1) reconstructing Sequence T80 onshore in the hanging wall of the Akatore and Titri Faults; (2) restoring the Waipounamu Erosional Surface (WES) to the horizontal (based on the evidence that WES is a time-transgressive wave-cut, near sea-level erosional surface), and (3) using the outcrops in the footwall of the Moonlight Fault.

The reconstruction of the stratigraphic sequences above the present erosional surface is illustrated in Step 1a, a step that is shown only for this transect to clarify the way reconstruction has been carried out. The restoration of the reconstructed Sequence T80 is shown in Step 1b. Restoration of top T80 to the horizontal is achieved by removing the offsets and hanging wall folds along the reverse faults of the Otago compressional province. The comparison between Steps 1a and 1b illustrates the amount regional tilting at the GSB western margin imposed by compressional reactivation, but also
shows that a component of tilting was acquired prior to Step 1b.

As already discussed earlier on, Sequence T70 in Transect T01 is incomplete, with erosional channels that were filled by sequence T80, and presumably formed by transport and re-sedimentation during gravitational sliding of the MTC. The lack of areas of T70 (volumes in 3D) indicates their removal away from the plane of the section, a condition that makes restoration impossible. Regional data and reconstructions from OMV suggest components of transport northwards. In order to achieve restoration, top T70 has been reconstructed by adding the missing top of the sequence (top T70), and reconstructing its virtual geometry parallel to top T60. The resulting restoration is shown in Step 2, and the added missing portions of Sequence T70 are shown in pink. The reconstruction shows that the MTC has removed a large volume of Sequence T70 from the central region of the basin.

The restoration steps of the post-rift sagging and drift stages (Steps 2-6) show that the largest components of subsidence occurred in the central rifted domain relative to the eastern and western margins. Relevant westward tilting of the eastern margin occurred during Steps 6 to 2 and pronounced tilting of the west margin during Steps 3 to 2. Note that rift-bounding normal faults at the western margin are considered active during Step 6.

The restoration of the syn-rift sequences shows the coeval activity of conjugate systems of normal faults, with possible migration of fault activity towards the western margin from Steps 7 to 6, and consequent widening of the stretched basin during deposition of the Sequences K60-K70.

The restoration of the pre-rift morphology of the basement has been obtained by rigid-body rotation and translation of the fault-bounded blocks. The reconstructed geometry shows the trajectory of propagation of faults that will become active during the rifting, and indicates the presence of two classes of dip ($\leq 30^\circ$ and $\leq 60^\circ$). The fault dips correspond with the real dips only if faults are crossed at $90^\circ$ by the transect (otherwise they are lower). This is generally true for the faults crossed by T01. The faults belonging to the class with lower dips ($\leq 30^\circ$) appear to be active in the early syn-rift
stages (Steps 8 and 7) and concentrate in the central area of rifting, whereas the faults with more typical angles of dip (~60°) appear to concentrate at the western margin of the rift, and in the area presently onshore. This setting suggests that the earliest normal fault were rotated during rifting and crustal stretching, and implies possible domino-style rotations above an unmapped intra-basement detachment.

The reconstructed topography in the top basement shows differences in elevation of the order of 1 km between highs and lows, with systematic increase in elevation from NW to SE. However, the reconstruction is incomplete, because the eastern shoulder of the rift is only partially included in T01. Note also that the faults are restored with the same angle of dip for ~5 km of crustal penetration, and that no data exist for inferring their listric geometry at depth. Rotations in the hanging wall of a spoon-shaped fault plane may lead to changes in the proposed reconstruction.

The stretching factor measured for T01 is $\beta=1.2$, indicative of low to moderate stretching. Note that stretching is clearly heterogeneous, and that the stretching factor is calculated over an initial distance $L_0$ of 93.2 km in the central part of the rifted basin.

**Restoration of Transect T02 (Plate 10)**

Restoration of Sequence T80 in T02 uses the onshore outcrops at km 80-100, and extrapolates the continuity of the marine sequence from the margin of the basin to the NW. Removal of reverse offsets in Step 1 does not alter significantly the geometry of the tilted west margin, because shortening is low.

The restoration of an erosional channel in Sequence T80 (km 270 in T02) causes a geometric problem, with apparent folding of the entire sedimentary pile in Step 1 (arrow 1). In order to avoid this inconsistency, the missing portion of top T80 (removed from the channel) should be added to the section before restoration. In fact, this is the procedure used in Step 2 for reconstructing the missing portions of T70 (in pink) above the basal surface of the MTC. However, given the small scale of the resulting geometric problem, no correction has been applied. Note – as already discussed for T01 – that a
large area (volume in 3D) of Sequence T70 has been transported away from the section, and that the zone with the most intense gravitational transport is laterally bounded by two structural highs.

The restoration steps relative to the post-rift sagging and drift (Steps 2 to 6) show a large component of westward tilting in Steps 2 and 3, associated with thickening of Sequences T60-T20 and T10-T00 in the central part of the basin, and against faults at the western margin that are considered to be active during these stages. This geometry is suggestive of rotational components, possibly associated with listric curvature of the bounding faults at depth ≥5 km (faults marked by the asterisk *). Thickness variations also suggest the possible reactivation (with strike slip components?) of the Murihiku-Dun Mountain-Maitai Terrane boundary crossed by this transect.

The restoration of the syn-rift stages (Steps 7-8) shows the coeval activity of conjugate fault systems in the rifted domain, with rotational components testified by the thickening of the sequences in the hanging wall of faults at the western margin of the rift (marked by the asterisk *). The rotational components can also be attributed to domino-style rotation of the whole set of E-dipping faults, but there is no evidence for an intra-basement detachment. As already noted for T01, there is some evidence that the rifted domains widened to the west from Step 7 to Step 8.

Restoration of the pre-rift top basement in Step 9 shows an elevation difference of ~1 km between the Murihiku Terrane to the NW and the Dun Mountain-Maitai Terrane to the SE.

The calculated stretching factor $\beta = 1.2$, indicative of low-moderate stretching of the region of original length $L_0=180$ km. Restoration has been achieved by accommodating the rotational components along a putative intra-basement detachment 6-7 km deep.

**Restoration of Transect T03 (Plate 11)**

Restoration of Sequence T80 in T03 uses the onshore outcrops at km 20-60 and shows – as in the previous transects – the transgression of the Oligocene marine deposits
above the tilted western shoulder of the GSB. Restoration of top T80 to the horizontal in Step 1 does not entirely remove the reverse offset on the faults at km 180-220; however, only a small component is left, and it is probably consequent on the way the faults have been interpreted in the seismic line.

Transect T03 is characterised by pronounced folding of Sequence T70 in the EIB. The restoration of top T80 in Step 1 clearly shows that the monoclinal flexure of the EIB was already formed during sedimentation of Sequence T80.

Restoration of top T70 in Step 2 reconstructs the area of Sequence T70 (volume in 3D) affected by gravitational sliding and transported out from the section along basal surface of the MTC. This area is smaller than in T01 and T02. The position of the EIB relative to the MTC is also marked.

Significant westward tilting of the eastern margin of the GSB is reconstructed for Steps 2 and 3. The boundary fault at the western margin of the basin (marked with the asterisk * and with the arrow 8 in Step 4, arrow 9 in step 5 and arrow 10 in step 6) was probably active during the early post-rift stages, with associated anticlinal folding in its hanging wall. This setting is indicative of an early stage of localised compression along this margin.

Restoration of the syn-rift sequences in Steps 7 and 8 shows an initial foundering of the central rift area controlled by sets of conjugate normal faults, and the time migration of fault activity towards the western margin. The eastern margin was characterised by discontinuous syn-rift deposits, with elevated horsts (with condensed or absent cover sequences) separating narrow intervening basins. The depositional geometry of Sequence K50 is compatible with syn-sedimentary listric faulting at the western margin (on the same fault system active in the post-rift, indicated by arrows 10 in Step 6).

The pre-rift geometry of the Murihiku basement reconstructed in Step 9 shows a central depressed region bounded by marginal highs with elevation differences \(\leq 2\) km. This setting, and the observation that most faults in the central rifted area dip \(\leq 20^\circ\), whereas the faults at the western margin dip \(\sim 60-70^\circ\), suggest that crustal stretching was
accommodated by different generations of faults, with some faults likely to be active before deposition of Sequences K20-K40. This implies that some components of stretching are not accounted for in the restored section, and that the computed stretching factor $\beta=1.38$ is a minimum estimate.

Restoration of Transect T04 (Plate 12)

Transect T04 has no onshore connections, and therefore it does not include the western shoulder of the GSB, nor it does extend to the eastern margin.

Restoration of top T80 in Step 1 removes the fold of the Toroa Anticline and shows the thickening of Sequence T80 in front of the EIB. This reconstruction shows that there is a pronounced eastward tilting at the western end of this transect, not relatable to the basement geometry.

In contrast with the previous transects, there is no missing portion of Sequence 70 removed by gravitational sliding of the MTC, and the basal surface of the MTC is restored in Step 2. The EIB is restored to the horizontal in Step 2 as well.

Restoration of the post-rift sequences in Steps 3 to 6 shows a marked difference between the north-western, strongly subsiding basin, with basinward tilted margins, and the south-eastern region, characterised by very thin sequences.

Note that restoration of top T70 (Step 2) and of top T60 (Step 3) does not remove the folding associated with the fault beneath the Tara 1 well, suggesting that localised compression occurred at an early stage in this position (see also the geometry shown at the west margin of T03 in Steps 3 and 4). However, as already noted, the basement fault maintains normal offset.

In contrast with the other transects, Sequence K20 (earliest syn rift) has been mapped along the north-western part of T04. Thus, the syn-rift restoration includes an additional step (Step 8b). K20 is only mapped between km 0-200, and thus its restoration shows
an eastern uplifted area in Murihiku Terrane, with elevation ≤2 km above the subsiding rift basin. However, this reconstruction can change if some of the sequences attributed to K30-K40 in the southeastern part of the basin actually include Sequence K20 as well. This seems likely, especially considering the marked difference in thickness of Sequences K30-K40 between the north-western and south-eastern parts of this transect.

The rifted domain is segmented by a central horst that separates a north-western basin affected by strong tectonic subsidence from a south-eastern domain filled by a thinner syn-rift sequence. This structural high appears to control also the subsequent development of the basin during the post-rift stages, and its position coincides with the frontal imbrication of the *EIB*.

The reconstruction of the pre-rifted basement in Step 9 shows a systematic increase in elevation to the south-east that is consequent on the lack of Sequence K20, but may also result from the lack of W-dipping faults at the south-eastern margin of the transect, that does not extend to include the eastern shoulder of the rift.

The measured stretching factor is low, with $\beta=1.16$, but – as already discussed – this transect does not include the margins of the rift, where the largest components of displacement are likely to occur.

*Restoration of Transect T05 (Plate 13)*

This transect does not show compressional inversion at the north-western margin, controlled by the faults of the Sisters Shear Zone. Restoration of top T80 to the horizontal preserves the fold of the *EIB* that controls localised thickening of an otherwise thin Sequence T80.

Both the *EIB* and the *MTC* are restored in Step 2. As for T04, the *MTC* has not removed portions of Sequence T70 away from the section.

Steps 2 and 3 show a strongly asymmetric basin with a gentle, eastward tilted north-
western margin and a steeper, westward tilted, south-eastern margin (arrows 2 and 4). Restoration Steps 4 and 5 show a more uniform distribution of the post-rift sequences along this transect, whereas Step 6 bears a strong imprint from the syn-rift stages, with discontinuous Sequences K60-K70, that are extremely thin or absent above the structural highs.

The syn-rift Steps 7 and 8 are characterised by discontinuous basins separated by horsts, bounded by systems of conjugate normal faults. The rotation of the syn-rift sequences in the hanging wall of the Sisters Shear Zone (marked with asterisk *) is consistent with the listric curvature of this fault at depth. Note that the rotation of the sequences is reconstructed for Steps 8 to 4, suggesting that activity of the fault continued after the syn-rift stages. Note also that the elevated footwall of the Sisters Shear Zone acted as a boundary for all the sequences.

The restoration of the basement in the pre-rift stages (Step 9) gives a low stretching factor $\beta=1.1$. Note that the eastern shoulder of the rift is not included in this transect.
6.3 Length and Area Balance of the Regional Transects

The restoration steps described before provide a plausible, geometrically coherent, and kinematically admissible reconstruction of the structural and sedimentary evolution of the GSB, and are a test of the reliability of the interpretation of the seismic lines and of the connections between offshore and onshore geology.

A further test is provided by the length and area balance of the horizons and sequences in the regional transects T01-T05. Poor balancing can be caused by unrealistic structures, errors in the seismic interpretations, and/or incorrect kinematic models.

Balancing assumes that the deformed state and restored cross sections maintain constant area. The original concept of balanced cross sections was based on the principle of conservation of bed length and area, i.e. no change in thickness (Bally et al., 1966; Dahlstrom, 1969). If there is no deformation in the direction perpendicular to the plane of the section, an area-balanced section maintains constant volume after deformation.

However, the assumption of constant bed length implies that flexural slip (i.e. slip parallel to bedding) is the dominant deformation mechanism. Other deformation mechanisms (e.g. fault-parallel simple shear, or simple shear along vertical or oblique slip lines) do not maintain bed length and thickness, but preserve area. These mechanisms are generally used in the restoration of extensional structures. Note that extension of rifted domains lowers the reference beds below the regional datum, promoting sedimentation and syn-sedimentary thickening in the depocentres located in the hanging wall of active normal faults. These mechanisms cause a net increase in the area of the cross sections (volume increase in 3D).

Many studies have emphasised that slip on major faults and large-scale folds only partially account for the total strain (Fischer and Coward, 1982), and that a substantial amount of strain (30-60% according to Kautz and Sclater, 1988) is actually
accommodated by sub-resolution small-scale structures and by internal ductile deformation that is generally not considered by regional reconstructions and in the interpretation of seismic reflection data.

The software LithoTect™ used for the analysis of the regional transects T01-T05 performs a “template” balance, that restores a cross section to the horizontal starting from the final deformed state, by measuring the length of the horizons and the area of the stratigraphic sequences bounded by the horizons. The undeformed template is constructed by fitting back to the horizontal all the areas separated by faults. The thickness assigned to the balanced units is computed from the area/length ratio.

The template restoration is useful to highlight interpretive problems (though it does not reveal the source of the problem), and shows how the sequences in the multilayer are length and/or area balanced relative to each other. Note that lack of balancing is a problem only if: (1) the initial stratigraphic sequences have a layer-cake depositional geometry; and, (2) all deformation occurred after deposition of the sequences. In the case of syn-sedimentary extension or shortening, or in the case of growth faulting, length and area of superposed sequences are not expected to balance, and differences in length are actually a measure of the amount of shortening or extension.

Note also that balancing of cross sections should be performed only for sections that extend laterally to comprise undeformed margins (i.e. foreland to foreland sections for fold-and-thrust belts and sections that include the shoulders of the rift in extensional settings).

Plate 14 shows the template balance for transects T01-T05. The legend is the same as in Plate 3. Sequences from top T80 to the top basement are balanced, and the relative measurements of length of horizons (in km) and area of sequences (in km²) are given in Plate 14. Vertical exaggeration is 8:1.

Given that the onshore portion of Transects T01, T02 and T03 is dominated by exhumed and uplifted basement with only minor outcrops of preserved sedimentary sequences, only the offshore portion has been balanced. Note also that where
Sequences T10 and T20 are separated by the *MTC*, this boundary has been used as the stratigraphic base of T20, in order to avoid a significant discrepancy between the length of the boundary and the area of the Sequences T20-T60.

In all transects the balance is generally good, and shows: (1) the increase in length of the syn-rift sequences (K20-K40 and K50); (2) the elongation of the post-rift horizons connected with widening and deepening of the sagging basin; and, (3) the moderate syn-orogenic shortening.

Note that, with the only exception of T04, the post-rift units are longer than the top of balanced basement. This reflects the draping and sagging of units above a rigidly faulted basement.

Widening of the rifted area is shown by the lengthening of the syn-rift sequences in all transects. A problem arises with T04, where Sequence K20 is only mapped in the north-western half of this transect. This incomplete interpretation results in Sequence K20 being shorter than Sequences K30-K40 and K50. This geometry is probably unrealistic.

Sequences T20-T60 are well balanced in all transects, with the exception of transect T03. This discrepancy results from the fact that in T03 Sequences T20-T60 disappear at km 360 and are not mapped along the south-eastern part of this transect.

Shortening is computed as the ratio between the length of top T80 and the length of top T10. Sequences T90 and T100 have not been used (and are not included in the balance) because their geometry is strongly influenced by the prograding delta. T10 is chosen because its geometry is not affected by the *MTC*. Shortening in all transects is very low (from 0.98 to 0.84), and is not measured in T02.

Stretching is computed as the ratio between the length of top K50 and the length of top K40, and thus gives the stretching during the syn-rift stages. and provide an additional, independent measure of the stretching factor $\beta$ measured in Plates 9-13. It increases from 1.02 in T04 to 1.2 in T01, 1.21 in T03, 1.24 in T02, and 1.46 in T05. These values
are slightly different but comparable with those obtained in Plates 9-13. However, as
the top K40 is intra syn-rift, the estimates of $\beta$ must be considered conservative. Note
that in transect T04 the alternative measure of $\beta$ as the ratio between top K50 and top
K20 gives a value of $\beta=2.0$ that is clearly unrealistic because – as discussed before –
the short length of Sequence K20 is caused by its lack of interpretation in the south-
eastern half of the seismic line.

The other discrepancy is the relatively larger value of $\beta$ in T05 (1.46) relative to the
measurement obtained in Plate 13 ($\beta = 1.1$). This discrepancy results from the very
discontinuous sedimentary area of Sequences K20-K40 relative to Sequence K50.

7. SYNTHESIS

7.1 Deformation Sequence

The deformation sequence reconstructed from the seismic lines and validated through
the construction and restoration of the regional transects is summarised in Plate 15.

The chronology of the identified regional structures is based on the deformation of the
stratigraphic horizons used for the seismic interpretation and on the relative order of
superposition. In Plate 15 the deformations of the GSB are located in time relative to
the regional tectonic and igneous events in the greater New Zealand region.

The major points highlighted in this study (and summarised in Plate 15) are:

- Formation of the GSB rift was initiated by intense normal faulting on sets of brittle
  N-S to NE-SW faults, with tectonic subsidence in the basins bounded by conjugate
  fault sets, or localised in the hanging walls of faults presumed to be listric at depth $\geq 10$
  km. Crustal stretching and exhumation of the rift shoulders are testified by the activity
  of the ENE-WSW Sisters Shear Zone in the latest rifting stages (89-83 Ma). It is
  unclear whether the obliquity of the Sisters Shear Zone relative to the early syn-rift
  faults results from a change in the trajectory of the extensional stress or from rotation of
  the exhumed crustal blocks.
• Rifting in the GSB was completed by 82 Ma, but extensional reactivation of the inherited Terrane boundary faults cannot be excluded during the interval 85-55 Ma.

• The long post-rift period of sedimentation from 82 to 19 Ma was not characterised by recognisable tectonic deformation in the GSB, though important regional tectonic events occurred in adjacent regions, with opening of the Tasman Sea (85-55 Ma), and rifting in the Solander Basin (45-20 Ma). Extensive interpretation by OMV of seismic lines and data from Tara 1 and Kawau 1 wells, together with the restoration of transects T03 and T04 (Plates 11 and 12), provide some evidence for synsedimentary pulses of shortening, testified by folding of T10, and by short wavelength folding above basement faults that maintain normal offset. Some of these deformations may actually reflect local accommodation during subsidence and compaction of water rich sediments undergoing rapid burial. It is also possible that the western margin of the Campbell Plateau was affected by the far field stress generated in the adjacent regions, but there are no simple explanations for why some individual faults underwent short term, transient pulses of compressional reactivation.

• During the post-rift interval, vertical movements were principally controlled by differential thermal subsidence that affected the whole rifted region and its margins. Differential subsidence was accommodated by tilting. In the post-rift stages 1 and 2 strong components of westward tilting are localised along the eastern margin of the GSB, at the transition between thick post-rift sequences deposited in the subsiding basin and condensed sequences on the eastern shoulder of the rift.

• Slow subsidence and marine transgression along the western margin of the rift resulted in the diachronous westward propagation of the Waipounamu Erosional Surface.

• The post-rift stage 3 is characterised by a change in structural style, with regional scale gravitational movements and layer-parallel detachments along the Mass Transport Complex (MTC). The detached body comprises sediments of Sequence T70 and is onlapped by sediments of Sequences T80 and T90. The onlap of sediments of different
age above the sliding mass indicates that movement occurred in jerks of progressive
detachments, rooted along a common basal surface. Composite, diachronous
movements of mass transport complexes are described in the literature (Moscardelli and
Wood, 2008; Butler and McCaffrey, 2010).

- The detached flexure of the *Eastern Imbricate Belt (EIB)* was formed in the late
  post-rift 3 stage. The time development of the *EIB* is short, and constrained by the
  observations that the fold deforms the base of the *MTC*, and is onlapped by top T80.
  Thus, formation of the *EIB* chronologically overlaps progressive sliding of the *MTC*
  and the eastward tilting of the western margin of the GSB.

- Shortening of the GSB occurred in a relatively short time interval from 19 Ma to
  Present. Shortening is principally testified by moderate compressional inversion of the
  inherited syn-rift normal faults (oriented N-S to NE-SW). Compressional inversion is
  concentrated at the western margin of the GSB along the Otago coastline. However,
  there is evidence of localised compressional inversion for other faults sparsely
  distributed in the GSB (notably, the faults of the Rakiura 1 and Kawau 1 structural
  highs). Shortening is also accompanied by formation of new reverse faults and
  transpressive reactivation of the Terrane boundaries.

- Growth of the Toroa Anticline occurred during the time interval of syn-orogenic
  compressional deformation, and the fold amplified after deposition of Sequences T100-
  T110.

- Sets of fractures and low-displacement faults are present in many seismic lines,
  especially in Sequences T70-T90 and the carbonate-rich units

### 7.2 Deformation Mechanisms

The structural style of the GSB is characterised by split-level deformation, influenced
by contrasts in mechanical competence both vertically and laterally within the basin.
Vertical contrasts are controlled by the difference in mechanical rigidity between the basement (with inherited fabric from Gondwana accretion, and metamorphic overprints) and the low-competence cover sequence, characterised by alternating permeable and impermeable horizons that underwent subsidence and differential compaction during the long time interval of post-rift sagging.

The horizontal contrasts are controlled by the inherited Terrane boundaries and/or the reactivation of the zones of crustal weakness separating the accreted Terranes. These differences are reflected by the distinct structural setting of tectonic domains delimited by the inherited Terrane boundaries (Plate 1). Components of strike-slip along the Terrane boundaries are likely, but have not been analysed in detail for this study.

Deformation mechanisms in the GSB result from the mechanical response of the deformed units to the imposed magnitude and orientation of regional stress fields, but structural interference is caused by the inherited structural fabrics and layer-parallel detachments (locally promoted by overpressuring).

**Syn-Rift Deformation Mechanisms**

Syn-rift extension in the basement is only partially imaged in the seismic lines, and resolution is generally poor below 6 km depth. Deformation over the analysed depth range is predominantly brittle, and the tectonic style is characterised by block faulting, with rigid body translation and rotation. The resolution of the seismic lines is insufficient to reveal small-scale structures likely to be present in the belt of long-lived faults with large displacement.

Onshore outcrops (e.g. the Sisters Shear Zone, see Kula et al., 2007; 2009) indicate that the brittle fabric of large-displacement faults in the upper crust overprints mylonitic belts formed along ductile shear zones at depth ≥10 km. Progressive exhumation of basement rocks along seismically active faults is also marked by pseudotachylytes, as described for Tucker Hill (Barker, 2005, see Table 1), and by belts of gouges and cataclastic fault rocks. However, the thickest fracture meshes are generally associated
with low-displacement faults, and their formation predates the propagation of a late, through-going fault (Sibson, 2000). The grain size, lithology and cementation of the gouge and cataclasites, and the foliation eventually imposed by shearing control the permeability contrast between the fault zone and the surrounding host rocks. No data for the syn-rift fault permeability are available for this study.

Syn-rift deformation phases in the GSB are identified through the sedimentary filling of the subsiding basins bounded by normal faults. The geometry of the syn-rift units is indicative of progressive down-faulting, with the largest components of extension acquired in the earliest syn-rift stages (and, possibly, earlier than deposition of the oldest sediments).

Progressive faulting in the syn-rift is structurally documented by the presence of different generations of normal faults, with varying dip angles, and with different amounts of total displacement.

The majority of the offshore faults within the area of maximum foundering of top basement possess low dips (20º-30º), much lower than the dip (~60º) expected for formation of new normal faults in an extensional regime. However, the onshore syn-rift faults show higher dips (50º-70º). This geometry may reflect multiple causes, like: (1) extensional domino rotation of normal faults from original angles of 50º-60º to 20º-30º in the region that underwent the greatest amount of crustal stretching; (2) steepening by late compressional rotation of originally low-angle onshore syn-rift faults in the onshore area; (3) extensional reactivation of low-angle thrust faults inherited from accretion along the Paleozoic Gondwana margin.

There is evidence on some seismic lines of listric curvature of the syn-rift normal faults, but seismic data do not image sufficiently deep to confirm the presence of intra-basement, sub-horizontal detachments. The observed rotation of the hanging wall blocks, especially for the faults along the western margin of the GSB, is also consistent with listric geometry of the normal faults at depth ≥10 km.

The geometry of the whole rifted zone remains undefined, because no seismic data are
available to identify the structural geometry of the eastern margin of the GSB. Thus, it is impossible to discriminate between the alternative possibilities of: (1) an asymmetric half graben with localised low-angle extensional shear zones along the faulted western margin (e.g. simple shear models of Wernicke, 1981; Lister et al., 1986), or, (2) a symmetric graben bounded by conjugate sets of normal faults of comparable displacement at both margins (e.g. pure shear model of McKenzie, 1978).

The two models carry different implications for the geometry of the detachment between the upper brittle crust and the lower ductile crust, for the position of the area of maximum crustal thinning in relation to the fault zones, and for the geometry of hot mantle upwelling in the rifted region, thus implying different thermal evolution during the rift and post-rift stages.

Post-Rift Deformation Mechanisms

Differential movements caused by thermal subsidence, high sedimentation rates (possibly contributing to overpressuring), and differential compaction dictate the deformation style of the cover sequences in the post-rift. Competence contrasts and fluid expulsion from units undergoing rapid burial and compaction may in fact explain some of the post-rift deformation, like the localised folding affecting T10.

The geometry of a strongly subsiding stretched area relative to the rift shoulders controls the relevant change in thickness of the post-rift sequences and the marked tilting at the rigid shoulders of the rift that maintained the character of rigid buttresses during deposition of Sequences T10-T70. However, it is unclear to what extent these mechanisms may have triggered the localised deformation within the Eocene sequence observed at the western margin of the rift, with sparse, compressional inversion of individual faults that cannot be easily fitted with models of post-rift deformation.

Differential subsidence, compaction and tilting are likely to be accommodated by a combination of layer-parallel slip and pervasive jointing (Price and Cosgrove, 1990). Most seismic lines reveal intense, high-angle jointing in Sequences T70-T90. These
structures are likely to develop in large numbers at scales not resolvable by the seismic lines, and may play an important role in controlling lateral and vertical fluid migration from overpressured compartments.

Layer-parallel and layer oblique detachment of poorly consolidated sediments during the late post-rift stages is documented by the gravitational sliding of the Mass Transport Complex (MTC). Movements were likely triggered by tilting at the rift shoulders, but restoration of the regional transects and regional geological data suggest that the principal trajectory of movement was from south to north, and related to the regional tilting of the GSB towards the Bounty Trough. Sliding was probably diachronous and connected with multiple detachments that sole into a major basal surface. The role played by currents in redistributing poorly consolidated sediments has been emphasised for the Oligocene-Miocene sediment drifts in the Canterbury basin (Fulthorpe and Carter, 1991), and relates to the initiation of the Antarctic Circumpolar Current. This effect may well play a role in the GSB, but the seismic evidence suggests that the primary mechanism of emplacement of the MTC is gravitational instability.

The other relevant structure that affects the late post-rift sequences is the Eastern Imbricate Belt (EIB). This is a peculiar structure, because: (1) it is not a complete fold, but rather a monoclinal flexure; (2) it is detached within T70; (3) it was formed in a rather short time period immediately after the onset of the MTC and during its detachment; (4) it is localised at the transition between the thicker post-rift units and the thin, condensed sequences overlying the eastern structural highs.
All these features suggest that gravity plays a role in the formation of the EIB, and that a genetic link possibly exists between the MTC and EIB. In particular, the two structures may result from a combination of gravity gliding (i.e. down-slope movement by translation of a mass above a weak detachment surface) and gravity spreading (i.e. vertical collapse and lateral spreading of a mass internally deformed under its own weight, cf. Schultz-Ela, 2001 and Rowan et al., 2004). A combination of gravity gliding and gravity spreading (Figure 3) is generally associated with disequilibrium compaction and overpressuring induced by rapidly prograding deltas (cf. Ings and Beaumont, 2010), as eventually could be the case along the western margin of the GSB.

A better understanding of these mechanisms and of their potential role in the generation of the MTC and EIB requires a 3D reconstruction of the detachment surfaces and of the sliding bodies in relation to sedimentary input and paleoslopes within the basin.

**Syn-Orogenic Deformation Mechanisms**
Selective, compressional inversion of inherited normal faults is well documented in the Otago region and along the western margin of the GSB. Characteristic features of mechanisms of compressional inversion (cf. Williams et al., 1989) are: (1) coupled deformation between basement and cover sequences; (2) localised anticlinal folding of short wavelength in the fault hanging wall; and, (3) “harpoon-head” geometry of the thick syn-rift and post-rift sequences onlapped by syn-orogenic deposits (Figure 4). Such deformation involves a combination of rigid translation and rotation in the basement and layer-parallel flexural slip and bending in the cover sequence.

Compressional inversion of inherited syn-rift normal faults is accommodated by drape folding of the detached cover for low to moderate amounts of shortening. Reversal of movement by increasing shortening is indicated by minimal to null vertical displacement in the basement and in the syn-rift sequences. In contrast, the post-rift and syn-orogenic sequences are offset with reverse displacement, following propagation of new fault segments across the post-rift and syn-orogenic sequences (cf. Sibson and Ghisetti, 2010).

Reactivation of steep reverse faults during compressional inversion likely depends on the local accumulation of fluid overpressures, with the fault system then serving as an escape pathway for episodic expulsion of overpressured fluids (Sibson, 1995).

The structural geometry of inverted faults at the margin of the GSB indicates only mild to moderate inversion, in agreement with the low shortening measured from the balanced cross sections (Plate 14). The inherited normal faults reactivated in compressional inversion are both rift-bounding faults (predominantly NNE-SSW oriented), and transverse faults (WNW-ESE oriented), controlled by the Terrane boundary architecture.
Figure 4  Progressive deformation of a syn-rift basin by selective compressional rectivation. Stages b and c are recognised in the offshore of the GSB. Stage d is recognised in the Otago province onshore. Redrawn and modified after Sibson and Ghisetti (2010).

The coast-parallel fabric of NNE-SSW localised compressional inversion in the Otago offshore is abruptly interrupted by the transverse structures of the Waihemo Fault to the north and by fault systems localised along the Dun Mountain-Maitai belt to the south. The simplified structural map of Figure 5 relates Miocene to present transpressive shearing at the Terrane boundaries with the left step of the belts of compressional inversion from the coastal region of Otago to the west margin of the Toroa Anticline and to the offshore regions close to Rakiura 1 and Kawau 1. Late crosscutting of NE-SW structures by strike-slip faults is likely to have affected fluid migration within the basin.
Note, however, that the structural map of Figure 5 is preliminary, and its validation requires further testing. In particular, it is desirable to define more precisely the geometry and timing of faulting at the Terrane boundaries, based on the analysis of suitably oriented seismic lines and on the constructions of additional regional geological transects transverse to the boundaries.

The structural map of Figure 5 shows the position of the Toroa Anticline at the eastern,
outer margin of the belts of compressional inversion, the geometric confinement of the faults in between Terrane boundaries, and the position of the most pronounced monoclinal flexure of the EIB at the eastern margin of the Toroa Anticline.

Although the Toroa Anticline was formed over the same time interval of compressional inversion at the GSB margin, the following features suggest that compressional inversion was not the prime mechanism of formation for this structure:

(1) Eversion of a large scale symmetrical fold of the dimensions of the Toroa Anticline requires inversion of an initial graben of comparable width, by compressional reactivation along its bounding faults. In the case of the Toroa Anticline compressional inversion is only present at the western margin of the basin. The anticline is not localised above a particular fault system, and its wavelength seems not likely to be controlled by faults that bound the deepest depocentre of the syn-rift basin.

(2) Net fault offsets in the basement, the syn-rift, and the post-rift sequences remain normal. There is no eversion of the syn-rift sequence. The horizon above which the fold grows in amplitude is top T10 in the post-rift sequence. These relationships show that compressional inversion of inherited basement faults is unlikely to have initiated buckling of the post-rift sequences.

Thus, gentle folding at such a large scale cannot be ascribed to localised reverse faulting in the basement.

Modelling of the evolution of the GSB performed by independent studies (OMV data) suggest that the Toroa Anticline grew in response to bending caused by the differential load imposed by the prograding Plio-Quaternary delta. If this is the case, the structure is related to differential flexure of the western limb, and was not formed in response to regional shortening.
However, it is worth noting that the axial trace of the fold does indeed lie parallel to the structural trends that characterise compressional inversion. It cannot be excluded that components of shortening in the Paleogene and Neogene sequences are caused by transpressive reactivation of WNW-ESE faults, especially considering that the lateral extent of the Toroa Anticline is confined by Terrane boundaries. In fact, some seismic lines parallel to the coastline south of Takapu 1 cross the offshore continuation of the Tuapeka Fault and the Hillfoot Fault (see Plate 1), and show a belt of transpressive (sinistral?) shearing, characterised by both reverse and normal movements on high-angle faults, and by folding of the Miocene to Plio-Pleistocene sequences. The faults are localised along the belt of the Dun Mountain-Maitai Terrane, but they crosscut the earlier basement structures localised at the boundary with the Caples Terrane to the north and with the Murihiku Terrane to the south.

Thus, the possibility that components of transpression at the Terrane boundaries coupled with compressional inversion along the western margin of the GSB amplified an initial instability arising from differential vertical load cannot be excluded.

A synthesis of the deformation phases in the GSB is provided in Figure 6, showing the chronology and character of the principal deformation events along an ideal transverse transect that encapsulates the features of the regional geological transects T01-T05.
Figure 6 Sequence of progressive deformation along an ideal transverse transect across the GSB and adjacent onshore regions. See text for a discussion.
7.3 Crustal Stretching

The effect of rifting on continental lithosphere is its net stretching, with shallowing of the Moho, high surface heat flow, volcanism, isostatic adjustments with subsidence in the rifted basin, and elevated topography at the rift shoulders (see Allen and Allen, 2005 for an extensive review and references).

Models of continental extension have evolved from the uniform, pure shear stretching model of McKenzie (1978) that assumed simple boundary conditions, with instantaneous, uniform stretching with depth, no magmatic activity, vertical heat conduction and Airy isostasy. According to the McKenzie (1978) model, syn-rift tectonic subsidence imposed by normal faulting depends on the ratio between the initial thickness of the crust and lithosphere, and on the amount of extension measured by the stretch factor $\beta$ (Figure 7). Post-rift subsidence decreases exponentially with time, and depends only on $\beta$. Conductive heat flow reaches 1/e of the original value after a time period of ~ 50 Ma for standard lithosphere.

![Figure 7 Continental extension model of McKenzie (1978) and definition of the stretch factor $\beta$. Note that under the assumption of pure shear, if stretching in the rift occurs by a factor $\beta$, then thinning of the crust is $1/\beta$.](image)

Subsequent variations and refinements of the McKenzie model (e.g. Kusznir et al., 1991; Watts, 2001) introduce more complex boundary conditions (e.g. depth-dependent stretching, magmatic activity, radiogenic heat sources, flexural isostatic support).
Extension by simple shear (Wernicke, 1981) has been used to model asymmetric rift margins, exhumation of core complexes, and the lateral shift between the upper crustal domain of extension and the zone of upwelled asthenosphere. Other models (e.g. Kusznir et al., 1991) assume simple shear in the crust and pure shear in the mantle lithosphere.

In all models heat flow and subsidence are related to the stretch factor $\beta$; thus, independent estimates of these parameters are essential for constraining the evolution of rifted basins through time.

Estimates of $\beta$ are generally based on the measurement of variations in crustal thickness using gravity and seismic data, and/or on deep seismic reflection and refraction profiles. Other procedures reconstruct the amount of crustal extension based on the offset along upper crustal faults (through section restoration and balancing) or reconstruct the tectonic subsidence of the basin by backstripping the sedimentary layers and computing the combined weight of sediments and paleowater depth (e.g. Vangen and Faleide, 2008).

In this study $\beta$ has been estimated from the restored regional transects T01-T05 as the ratio between the length of an extended region at the end of rifting (i.e. before thermal, post rift subsidence) relative to its equivalent, pre-rift length (see Plate 9 to Plate 13). These values thus provide an estimate of the component of tectonic stretching due to brittle displacement on regional faults, and do not account for small-scale deformation, and/or ductile stretching at depth.

As already discussed, the values of $\beta$ obtained for transects T01-T05 depend on the homogeneity of fault distribution along the sampled length of the regional transects, on the inclusion of all the relevant faults in the analysed seismic line, on the regional extent of the transects relative to the margins of the rift, and on the age of the horizons used for calibrating the offset. For all these reasons, the obtained values likely provide minimum estimates of tectonic stretching in the GSB.
Figure 8  Comparison of estimates of $\beta$, based on extension from crustal faults along selected transects (Transects T01-T05 and selected seismic lines in Cook et al., 1999), and based on tectonic subsidence
It is also important to note that estimates of $\beta$ will be affected by: (1) the orientation of the transect relative to the stretching direction; and, (2) the (unproven) assumption that the fault slip vector is perpendicular to the fault strike, i.e. faults move with pure dip-slip mechanisms. There are no systematic data on these constraints, with the only exclusion of the measurement of $\sim 160^\circ$ for the stretching lineation of the Sisters Shear Zone (Kula et al., 2009), that was active in the late rifting stages and is oblique to the majority of faults in the GSB (Plate 1). It is interesting to note that the transect with the highest measured value of $\beta$ is transect T03, that is not parallel to the other transects and has an average trend of 140°, closer to the stretching direction of the Sisters Shear Zone than all the other transects. The measurements made for transects that are oblique to the stretching direction will systematically underestimate the $\beta$ values of an amount that increases with the angle of obliquity.

The $\beta$ values obtained by balancing transects T01-T05 (Plate 14) are not dissimilar from those measured from the restored transect (Plates 9-13). However, the estimates are based on elongation of syn-rift markers, and thus possibly underestimate the total stretching. In addition, problems of inhomogeneous distribution of the syn-rift sequence cause unrealistic values for transects T04 and T05, and are thus omitted from this discussion.

Figure 8 shows a comparison between the $\beta$ values obtained in this study and those estimated by Cook et al. (1999), based on extension from normal faults along seismic lines and tectonic subsidence curves from the GSB wells. There is a rather good agreement between the data sets, both indicating low to moderate stretching ($1.1 \leq \beta \leq 1.4$), without a clear gradient in the amount of extension associated with faulting across the GSB. However, as already pointed out, and as also indicated by Cook et al. (1999) these values are likely to be conservative and to underestimate the total extension of the GSB.

The analysis of tectonic subsidence from wells (Cook et al., 1999; Sutherland et al.,
2010) provides slightly higher values, with $1.1 \leq \beta \leq 1.5$, and show that the largest values (Figure 8) coincide with the area of maximum foundering of the basement (see Plate 1) and syn-rift sedimentary infilling.

Unfortunately, there are no published data on the crustal thickness in the GSB, and thus it is impossible to estimate the value of $\beta$ from the depth of the Moho in the rifted region relative to its margins. The only comparison can be done relative to the regional transect T03, interpreted from the reprocessed seismic line AWI2003001 (Grobys et al., 2009). The refraction and wide-angle reflection profile and the accompanying magnetic and gravity data collected along this line have been assembled by Grobys et al. (2009) in an interpretive crustal transect that is reproduced in Figure 9, underneath the simplified structural outline of the geological transect T03 (Plate 6).

The interpretation extends to depths of 36 km; it shows the necking of the Moho beneath the GSB, and the transition to the thickened crust of the western margin and to the thinned to normal continental crust in the eastern Campbell Plateau. Beneath the eastern margin of the GSB, a high velocity crustal body 14-22 km deep is interpreted by Grobys et al. (2009) as a layer of underplated crust.

The crystalline crust of the GSB is thinned to 13 km along this transect. The pre-rift crustal thickness is unknown, and could vary from 35-40 km (thickened Gondwana orogenic crust, similar to the present thickness of cratonic crust in Australia) to an average value of 30 km characteristic of undeformed continental crust. At the eastern margin of T03, the crust of the Campbell Plateau is ~ 22 km thick (Figure 9), but whether this thickness is the original pre-rift thickness or has rather been reduced by stretching remains speculative.
Figure 9  Transect T03 superposed on the crustal interpretation of the line AWI2003001 by Grobys et al. (2009). The crystalline crust of the GSB is thinned to 13 km (Tf). If the original thickness is ~22 km (To, eastern Campbell Plateau), $\beta$ ~ 1.7, but for original values up to 45 km of the Gondwana orogenic crust $\beta$ may be as high as 3.

$\beta$ values estimated from crustal thinning therefore range from 1.7 to 3, depending on the thickness of the pre-rift crust, and are significantly higher than the values estimated from crustal faulting along the same transect ($\beta$ =1.38, see Plate 11 and Figure 8).

For many extensional margins stretching measured from the heaves on faults mapped on seismic lines is actually insufficient to explain the amount of crustal thinning modelled from crustal experiments and gravity, or that required to explain the tectonic subsidence (cf. Reston, 2009).

Some authors (e.g. Marrett and Allmendinger, 1992) have suggested that up to 50% of stretching is actually taken up by distributed deformation, with small-scale faulting not accounted for by the resolution of the seismic lines.
Other lines of evidence based on the modifications to the McKenzie (1978) model indicate that stretching is depth-dependent and larger in the lower crust relative to the upper crust where the brittle faults accumulate their offset.

A third possibility is that mapping of syn-rift faults in the seismic lines systematically underestimates or neglects the early rift events, especially in the case of poly-phased dominoing of early faults, that rotate to lower dip angles and become obscured by onlapping sediments and new, cross-cutting faults (see Figure 13 of Reston 2009).

All these factors could play a role in explaining the stretching discrepancy observed for the GSB, but additional data on crustal stretching and on upper crustal extension are needed to fully analyse this problem and quantify the discrepancy in the study region.

### 7.4 Crustal shortening

The tectonic regime that has dominated the evolution of New Zealand since the middle-late Miocene is imposed by the opposed polarity of subduction between the Australia and Pacific plates at the Hikurangi and Puysegur trenches, accommodated by the transpressive right-lateral Alpine Fault and the Marlborough transform system (Figure 1).

Today, the far-field plate vector has an estimated orientation ENE-WSW, with relative motion of ~ 37 mm/yr (DeMets et al., 1994). This component is resolved in ~ 35.5 mm/yr of right lateral shear along the Alpine Fault boundary and ~ 10 mm/yr of NW-SE convergence across it in the central South Island (Beavan et al., 2002). The Marlborough faults have a cumulative slip rate of the order of 38 mm/yr, and thus entirely accommodate the plate motion.

Localised deformation along the plate boundary has resulted in ~ 600 km of right-lateral shear over the last 20 Ma (cf. Sutherland, 1999), with crustal thickening and uplift of the Southern Alps leading to a dramatic change in the source-sink depositional systems of the South Island (Norris et al, 1990).
Exhumation rates are of the order of 6-9 mm/yr in the central Southern Alps, decreasing to ≤1 mm/yr in areas well east of the Main Divide (Upton et al., 2009).

From a regional compilation of 362 focal mechanisms of Mw≥5.7 shallow earthquakes, Bird and Liu, (2002) infer maximum horizontal stress in the South Island to be generally oriented NW-SE.

Most structures lie sub-orthogonal to the imposed stress (Berryman, 1979; Pettinga and Wise, 1994), but in some regions (especially Canterbury and Otago, see Plate 1) systems of N-S to NE-SW and NW-SE oriented faults are arguably all active in the present tectonic regime. Contemporaneous reactivation of sub-orthogonal faults can be explained by mechanical weakness of some inherited structures, by a stress regime where the maximum and intermediate horizontal stresses have similar magnitude, and can easily switch during the stress cycle associated with peaks of seismic activity (Upton et al., 2009), or by an obliquely oriented stress field.

In the Canterbury and Otago regions, deformation extends east from the Southern Alps to the coast and beyond, reverse and oblique strike-slip faults active in a belt 00 km wide (Norris et al., 1990; Berryman et al., 1992; Ghisetti et al., 2007). However, crustal thickening in these regions is not significant. Available estimates for the Moho depth are of the order of 25-32 km along the east coast of the South Island. The southward increase in crustal thickness across the Waitaki-Waihemo Fault belt (see Upton et al., 2009) is probably inherited from the architecture of the Gondwana accretion.

Onshore, in the Otago region, reverse fault activity is accommodated along a set of closely-spaced faults, resulting in elevated strain rates (Liu and Bird, 2002). The coastline coincides with the hinge of a compressionally elevated crustal region, with structural elevation accommodated by reverse faulting. The structural data discussed for the GSB indicate that the outermost belt of N-S to NE-SW compressional convergence extends offshore. However, the inherited gradients in crustal thickness and rheology at the Terrane boundaries likely play an additional role in the basement deformation (Figure 5).
Overall, the structural trends and mechanisms of the compressional structures are consistent with a regional WNW-ESE orientation of maximum horizontal compression, and also with the available breakout data reinterpreted for the Galleon 1 well (R. Crookbain, personal communication).

The boundary between compressionally deformed and undeformed regions of the GSB is transitional, and evidence for compressional inversion of isolated offshore faults suggests that many faults can be reactivated in the regional stress field. However, the conditions (geometry, mechanical weakness, fluid pressure) driving selective reactivation of individual faults remain unsolved.

Balancing of the regional transects shows that reverse faulting and folding have not resulted in significant amounts of contraction in the offshore of the GSB (estimated values ranging between 0.84 to 0.98 for transects T01-T05, Plate 14).

However, it is important to note that strike-slip components are not quantified, and that shortening measured in provinces of compressional inversion is generally low until new, suitably-oriented reverse faults propagate across the high-angle inherited fault systems (Ghisetti and Sibson, 2007; Sibson and Ghisetti, 2010).

7.5 Uplift

Significant uplift is associated with all the deformation events that have affected the GSB since the Late Cretaceous rifting (see Figure 6 and Plate 15).

During rifting and associated crustal thinning flexural compensation was induced by tectonic unloading along extensional detachments, resulting in subsidence of the sedimentary basins and uplift of the rift shoulders. The models for shoulder uplift depend on the assumptions of listric vs. planar geometry of the bounding faults, on the partitioning of simple vs. pure shear in the upper and lower crust, on the flexural rigidity of the lithosphere, and on regional vs. local isostatic compensation (Kusznir et
As already discussed, there is evidence on some seismic lines of listric geometry, but seismic data do not image sufficiently deep to confirm the presence of intra-basement, sub-horizontal detachments. The regional transects (Plates 4-8) and the restoration of the syn-rift stages (Plates 9-13) show that extension on unequally spaced faults (with inferred domino geometry) was associated with elevated fault block crests separating discrete sub-basins within the early syn-rift stages and with stable, and persistently elevated shoulders during the whole rift episode. Data from the Sisters Shear Zone at the western boundary of the basin (Kula et al. 2007; 2009, see also Table 1) indicate exhumation and uplift at the end of rifting, compatible with lithospheric isostatic flexure following stretching and thermal disequilibrium.

Uplift during rifting may also result from underplated melts at the base of the crust. The density of the underplated material is less than the density of the replaced mantle, resulting in net components of uplift that are ~1/10 of the thickness of the underplated material (see Allan and Allan, 2005).

The crustal transect across the GSB analysed by Grobys et al. (2009) includes a high-velocity, 6 km thick body in the central-eastern Campbell Plateau, interpreted as underplated magmatic rocks (see also Figure 9). Further crustal transects are needed, however, to constrain the presence and lateral extent of underplating within the rifted region of the GSB, and to understand if the lower crust high-velocity body is indeed related to the late Cretaceous rifting or rather is inherited from the previous accretionary history at the Gondwana margin. Given the present state of knowledge, the possible role of underplating on the uplift history of the GSB is largely unconstrained.

Transient vertical instabilities in the syn-rift and post-rift stages may also be related to dynamic mantle support. In the North Sea, anomalous trends of post-rift uplift followed by rapid subsidence histories that depart from the McKenzie (1978) model have been attributed to Icelandic mantle plumes, i.e. to effects from a source region located ~900 km away from the subsiding basin (Nadin and Kusznir, 1995). A similar interpretation by Sutherland et al. (2010), involves a Late Cretaceous-Cenozoic West Antarctica
mantle plume (triggered by assimilation of the Gondwana subduction slab) to explain the anomalous isostatic rebound in Marie Byrd Land, and the rapid Paleogene subsidence in the GSB (70-40 Ma), with accommodation space 0.5 to 0.9 km higher than that expected from the measured $\beta$ stretching factor (assuming the McKenzie model).

In the post rift and syn-orogenic stages, important components of uplift are likely related to volcanism. In fact, the new seismic data available for the GSB show a complex distribution (in both space and time) of volcanic edifices, adding to the regional picture based on onshore data.

The youngest volcanic activity onshore is concentrated along the east coast of the South Island from Dunedin to Banks Peninsula. (~10 Ma in the Dunedin volcano and ~5.8 Ma in the Banks Peninsula volcano). More recent activity is recorded onshore near Timaru (2.5 Ma) and in the offshore GSB.

According to the flexural model of Godfrey et al. (2001), uplift along the SE coast of New Zealand at ~11 Ma was driven by a hot, buoyant load in the mantle beneath the Dunedin volcano. The modelled uplift profile (with 5 km of lithospheric elastic thickness) fits the flexed and uplifted geometry of the Oligocene marker horizon that is traced along the SIGHT seismic lines offshore from Dunedin.

High percentages of mantle helium, and high heat flow values measured today in the Dunedin region are consistent with ongoing active melting and basaltic underplating at the western margin of the GSB. It has been argued that volcanic activity is suppressed by the present compressional stress field.

Data presented in this study suggest that the recent components of uplift are predominantly controlled by ongoing compressional inversion along the margin of the South Island, and consequent progressive thickening of the crust. Their effect extends beyond the region affected by the Neogene volcanic centres, and cannot be solely explained by the mechanisms invoked by Godfrey et al. (2001).
Progressive restoration of the regional transects (see in particular T01, Plate 9) suggests that removal of compressional inversion at the west margin of the GSB is enough to restore back the Oligocene marker (top T80, restoration Step 1b) and the Waipounamu Erosional Surface to the horizontal. Restoration does not allow for paleobathymetric variations of the Oligocene limestones but these are supposed to be low, of the order of ~ 200 m for the facies outcropping onshore.

Note, however, that this report does not specifically analyse the history of volcanism in the study area, and the reader is referred to the recent review by Timm et al. (2010). Further analyses are needed to address the complex volcano-tectonic relationships in the GSB, and the impact of volcanic activity on vertical movements, stress field, heat flow and fluid migration.

8. CONCLUSIONS AND Recommendations

The study presented in this report provides a regional synthesis of the structural setting of the Great South Basin (GSB), in relation to surrounding regions (Plate 1).

The construction, progressive restoration and balancing of five regional transects transverse to the structural trends within the areas of PEP 50119, 50120 and 50121 (Plates 4-13) provide a consistent interpretation of the geometry, chronological sequence, and mechanisms of superposed deformation that affected the GSB and adjacent regions in the time interval from the Late Cretaceous (112 Ma) to the present.

This report focuses on the first-order structural assemblages resulting from the regional tectonics, and shows that there is a distinct change in the structural geometry of the GSB from north to south and from west to east, resulting from the inherited configuration of the basement Terranes, the geometry of the rifted basin, and the superposed late Neogene shortening.

Section restoration (Plates 9-13) and balancing (Plate 14) show that the interpretations
of the horizons in the seismic lines are geometrically coherent, and lead to a kinematic reconstruction that can be related to regional deformation events and large scale plate tectonic dynamics (Plate 15 and Figure 6). The connection between onshore and offshore geology is meaningful, though the stratigraphic horizons cannot be extrapolated for long distances inland and large portions of denuded basement cannot be restored.

However, it is worth remembering a basic paradigm of structural restoration and balancing to the effect that, while geometric coherence and restorability are necessary, they do not by themselves guarantee correctness of the interpretation.

Restoration of the transects validates the general interpretation of seismic horizons performed by OMV, but also emphasises some inconsistencies that are not necessarily caused by interpretive inaccuracy, but may reflect the combined effects of a non-layer-cake stratigraphy with: (1) changes in thickness of sequences; (2) gravitational movements out from the plane of the section; (3) changes in the offset of markers along individual faults; (4) accommodation of deformation by structures that are not mapped or visible in the seismic lines because of their scale, orientation or dip; (5) components of strike slip movements along the Terrane boundaries.

Overall the geometry of the GSB is dominated by the imprint of the Late Cretaceous rifting phases. Stretching of the basin calculated from the restored sections is moderate to low, with $\beta$ factor in the range 1.1 to 1.4. The basement faults active during rifting are regional crustal structures with a long history of propagation and activity, which culminated in selective compressional reactivation during Neogene shortening (and, possibly, even at an earlier stage in the Eocene).

The faults at the western margin of the GSB are inferred to be listric, with rotation into flat detachments at depth $>10$ km. The overall geometry of the GSB western margin is compatible with the growth of an asymmetric half graben (formed by crustal simple shear), but lack of detailed structural data for the eastern margin leaves open the alternative possibility of a symmetric graben (formed by crustal pure shear).
Post-rift sagging was accompanied by strong subsidence in the central part of the basin, localised uplift at the previously established rift shoulders, and tilting. Vertical instability along the basin’s margins likely triggered gravitational soft-sediment deformation, and detachments.

The change to regional compressional stress regime documented for the whole of New Zealand since the early-mid Miocene is reflected in the GSB by the compressional reactivation of the early normal faults, with this style of deformation prevalingly localised along the western margin of the basin, but also affecting a sparse number of faults in other regions of the basin.

Shortening was accomplished by rigid rotation and tilting along sets of sub-parallel faults, with folding in the low competence sequences. Where measured, the amount of shortening is low ($\leq 0.81$), consistent with the outer position of the GSB relative to the plate boundary and the relatively young age of the movements. In most cases compressional inversion does not remove the original normal offset of the basement along individual faults, indicating that the component of shortening was much less than the original component of extension.

Folding of the detached sedimentary sequence in a more external position occurs along the large-scale domal bulge of the Toroa Anticline, but the structure likely results from bending under the vertical load imposed by progradation of the Plio-Quaternary delta along the western margin of the GSB. However, further amplification of the early fold instability by late compressional inversion cannot be excluded.

The geometry of some seismic lines and the regional tectonic setting are consistent with components of deformation accommodated at the Terrane boundaries, by localised faulting with strike-slip components. Strike-slip movements likely occurred during the rifting stages (transfer zones accommodating differential extension) and during the late shortening events (left-lateral transpression), but there are no definitive data supporting this interpretation.

Further insights into the structural evolution of the GSB are needed in order to fully
understand: (1) mechanisms of deformation and interactions between applied stresses, vertical movements related to differential compaction and subsidence in the thick post-rift sequences; (2) geometric and mechanical constraints imposed by the configuration of the rifted region, basement architecture, and clastic sedimentary input from the thickened and uplifted crust along the Neogene transform boundary; (3) feedback interactions between structural development, fluid migration and overpressuring.

In addition, there has been no attempt in this study to evaluate the role of syn-rift igneous activity and of the Neogene volcanism in the Dunedin region, and the interactions between volcanism and tectonic setting in the GSB.

The interpretations and conclusions of this study provide the structural input for assessment of (1) distribution and segmentation of source rocks in the rift basin; (2) location, and geometry of hydrocarbon traps; (3) vertical and horizontal components of deformation during the evolution of the petroleum system; (4) timing of trap formation in relation to hydrocarbon migration; (5) role of detachments and gravitational sliding affecting the integrity of seals; (6) effects of late deformation episodes on the integrity of the trap-seal system.

Some of the key observations are as follows:

- The structural development of the basin occurred in three stages, characterized by different orientation and magnitude of the stress field, with:
  - Syn-rift (112-82 Ma), dominated by horizontal extension and vertical foundering.
  - Post-rift (82-19 Ma), dominated by differential subsidence, compaction, tilting and gravitational instabilities.
  - Syn-orogenic (19-0 Ma), dominated by horizontal shortening and vertical uplift.

- Fluid paths are affected by the geometry and structural characteristics of faults active in the regional stress field, with extension favouring upward migration of fluids, and shortening favouring fluid entrapment.
• The syn-rift stretching factor $\beta$ constrained by this study is in the range 1.1 to 1.4, suggestive of moderate extension. These values provide the minimum estimate of stretching of the GSB, and can be used as an input for basin modelling.

• The regional transects show a range of structural traps developed during the syn-rift and syn-orogenic phases of the GSB development. The most common structures are fault-bounded tilted blocks (syn-rift) and inversion anticlines (syn-orogenic). Other settings within the post-rift units (e.g. drape anticlines, remobilized sediments in the $MTC$) may also affect the trap-reservoir system.

• The Syn-rift traps are considered to be more favourable from a charge timing perspective, the peak charge being modelled (OMV confidential data) as significantly preceding the syn-orogenic phase.

• Re-activation of syn-rift faults during the post-rift and syn-orogenic phases, coupled with systematic high-angle fracturing and propagation of new faults is expected to impact on fluid redistribution within the GSB, and may have controlled secondary hydrocarbon migration.

This study provides a consistent structural reconstruction of the GSB, but additional analyses will help in reducing existing uncertainties. Suggested further investigations include:

• A refinement and/or improved definition of paleotopography along the regional transects during subsequent evolution stages, calibrated with sea-level variations.

• Decompaction of the sedimentary sequences during progressive retrodeformation.

• Forward modelling of structures of problematic interpretation, as e.g., the Eastern
Imbricate Belt and the Toroa Anticline, aimed at testing the likelihood of alternative deformation mechanisms.

- Reconstruction of the 3D geometry of the Eastern Imbricate Belt and possible relationships with the Mass Transport Complex.

- Forward modelling of the geometry of the syn-rift sequences controlled by normal faults with different geometries (planar vs. listric) and detached at different depth. Modelling of the progressive rotation of syn-rift normal faults with increasing stretching.

- Construction of geological transects suitably oriented for defining deformation along Terrane boundaries.

- Analysis of the volcano-tectonic structures and their impact on vertical and horizontal movements.

- Analysis of the relationships between sequence and mechanisms of deformation and structurally-controlled migration of fluids.

9. ACKNOWLEDGEMENTS

This study has benefited from valuable scientific discussions, insightful inputs and careful reviews by R. Crookbain and T. Allan. Scientific and technical support from the OMV team is also gratefully acknowledged, with particular thanks to R. Constable, S. Langdale and B. Sissons.
10. REFERENCES


Institute of Geological and Nuclear Sciences, 1:250,000 Geological Map 21, 1 sheet and 52 pp. Lower Hutt, New Zealand.


