Seismic interpretation of the eastern Gippsland Basin with application to fault seal analysis in carbon dioxide storage leads

by

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Thesis Submitted in Fulfilment of the Requirements for the Degree of

Doctor of Philosophy

The University of Adelaide

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May 2011
CHAPTER 1—INTRODUCTION

1.1 Study rationale

The Intergovernmental Panel on Climate Change report (IPCC, 2005) provides a comprehensive review of the science of carbon dioxide capture and storage (CCS). The report claims that CCS has the potential to form 15–55% of the cumulative mitigation effort worldwide for carbon dioxide (CO₂) abatement, based on modelling to the year 2100. The storage of natural gas and CO₂ at field scale and CO₂-enhanced oil recovery (EOR) schemes has already proven successful. It is reasonable to expect that CO₂ storage¹ in subsurface reservoirs on a large scale could also be feasible. However, the storage requirements of global CO₂ volumes are massive (IPCC, 2005), with high costs and source-to-sink² matching not always favourable, as is the case in Australia (Bradshaw et al., 2002). Additionally, there are competing uses for subsurface reservoirs, mostly from the petroleum industry but also from groundwater. The IPCC report is a technically and strategically focused document; however, additional layers of geological complexity are to be expected and these have understandably not all been included. Out of proportion effects, on the viability of a CO₂ storage site can, however, occur, when site assessments have not taken into account the full range of geological heterogeneity (Hovorka et al., 2004). The extent, openness and/or interconnectivity of the reservoir system have also recently been shown to be of pivotal importance in regards to the feasibility of CO₂ storage (Economides and Ehlig-Economides, 2009)—and thus, structural studies aimed at assessing fault complexities in relation to CO₂ storage become pertinent. The background science related to CO₂ emissions, CCS, geological trapping mechanisms for subsurface containment of CO₂, CO₂ storage options and assessment strategies are addressed in what follows.

1.1.1 Greenhouse gases and the greenhouse effect

The energy of the sun that reaches the Earth’s surface is re-radiated in the infrared range and subsequently absorbed by atmospheric greenhouse gases. Greenhouse gases include water vapour (H₂O), CO₂, methane (CH₄), nitrous oxide, ozone and chlorofluorocarbons. The term ‘greenhouse effect’ is defined as ‘the heating of the surface of a planet or moon due to the presence of an atmosphere containing gases that absorb and emit infrared radiation’ (www.pacifichydro.com.au/en-us/classroom/climate-change.aspx). For a clear sky, H₂O, CO₂, CH₄ and ozone contribute to the greenhouse effect at levels of 60%, 26%, 6% and 8%, respectively (Kiehl and Trenberth, 1997).

¹ Geological storage of CO₂, as distinct from tank storage, will be referred to as CO₂ storage from this point onwards.
² Source-to-sink matching refers to how proximal stationary emitters of CO₂ (e.g. power stations) are relative to CO₂ storage sites.
Data from the Mauna Loa observatory show that CO₂ concentrations have increased from ~ 330 ppm in 1960 to ~ 375 ppm in 2005 (Hansen, 2005). The present-day CO₂ concentration exceeds the maximum CO₂ concentration ever recorded from 400,000 years of ice core data originating from the Antarctic ice sheet (Hansen, 2005). Ice core data show that CO₂ concentrations have varied from as low as ~ 180 ppm to a pre-industrial level of ~ 270 ppm to the current highest recorded value (~ 390 ppm at the Mauna Loa observatory). The excess CO₂ in the atmosphere from anthropogenic sources is ~ 100+ ppm, and the buildup correlates to elevated atmospheric and surface temperatures. The subsequent increase in energy absorption resulting from increased CO₂ concentrations is a worrying trend that could induce a runaway greenhouse effect. For example, any increase in global temperature could induce the release of CH₄ from the Arctic tundra (Lawrence et al., 2008), as well as induce the release of CH₄ trapped in clathrates below the ocean floor (Archer and Buffett, 2005); thus, compounding the greenhouse effect.

Uncertainty in the environmental consequences resulting from these rising temperatures has prompted worldwide action to mitigate and counteract the further release of CO₂ (IPCC, 2005). Since 1990, the IPCC has produced a number of strategic policy oriented reports, which outline the effects of climate change on human habitation (IPCC, 2005). These reports are produced by scientific assessment, response strategy and impact-assessment groups; ultimately, these lead to in-principle restriction limits and preliminary guidelines for greenhouse gas inventories and climate-change protocols. These preliminary guidelines subsequently resulted in a number of international treaties being voted on and adopted (or not) by the majority of the international community (e.g. UNFCCC, 1992; Kyoto Protocol, 1992; London Protocol, 1996; Copenhagen Protocol, 2010).

The main stationary sources that are responsible for releasing CO₂ (globally) include the power generation industry (10,539 MtCO₂/yr; 79%), the cement production industry (932 MtCO₂/yr; 7%), oil refineries (798 MtCO₂/yr; 6%), iron and steel foundries (646 MtCO₂/yr; 5%), the petrochemical industry (379 MtCO₂/yr; 3%), the natural gas sweetening industry (50 MtCO₂/yr; 0.4%), with all other sources contributing 33 MtCO₂/yr (0.3%; IPCC, 2005). Of the 10,539 MtCO₂/yr released by the power generation industry, coal-generated power contributes 7,984 MtCO₂/yr or 60% of total CO₂ annual emissions. The in-principle restriction limits imposed by the protocols, as well as commencement of carbon credit trading in 2005, have prompted countries like Australia to assess CO₂ storage options at a nationwide level (Cook et al., 2000) and to instigate CCS-related research (Cook, 2006).
1.1.2 Carbon dioxide capture and storage
As defined by the IPCC (2005, p. 3) ‘Carbon dioxide (CO₂) capture and storage (CCS) is a process consisting of the separation of CO₂ from industrial and energy-related sources, transport to a storage location and long-term isolation from the atmosphere’. A further characteristic of CCS is that ‘Capture of CO₂ can be applied to large point-sources. The CO₂ would then be compressed and transported for storage in geological formations, in the ocean, in mineral carbonates⁴, or for use in industrial processes’ (IPCC, 2005). CCS is, however, only one of many mitigation options for the abatement of CO₂ emissions. Other options include conservation and energy efficiency, use of renewable or nuclear energy and coal to natural gas substitution (IPCC, 2005).

Gas reinjection to accompany natural gas storage is well established and, consequently, provides an analogue for demonstrating the potential of CO₂ injection and storage on a large scale. However, CCS does have to consider the source-to-sink emissions match in order for CO₂ storage sites to be practically and economically viable. For example, the source-to-sink emissions match of all stationary emitters in Australia was reviewed, with first pass cost estimates showing significant variations (Bradshaw et al., 2002). A later review of the source-to-sink emissions match was carried out at a worldwide scale, where the known distribution of sedimentary basins was compared against the location of regional emission nodes. Limitations to CCS viability in some regions were obvious (Bradshaw and Dance, 2005).

The economic viability of CCS is reliant on the total cost of both the storage and capture components. For example, the efficiency of coal and gas generating power plants ranges from 31 to 54%, but the increased fuel use when adopting CCS results in CO₂ capture costs increasing by up to 13–27% (IPCC, 2005). The impact of both the increased fuel consumption and capital outlay costs for CO₂ capture naturally affects CO₂ storage assessments, where the latter must be as technically straightforward, risk-averse and low cost as possible. However, regarding CO₂ storage assessments, geology is naturally complex, necessitating the development of efficient and detailed geotechnical workflows.

1.1.3 Trapping mechanisms and containment
Carbon dioxide trapping mechanisms can be grossly divided into three categories: (1) physical trapping that includes stratigraphic and structural trapping; (2) physical trapping that includes

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⁴ ‘Storage of mineral carbonates does not include deep geological carbonation or ocean storage with enhanced carbonate neutralisation. The formation of carbonates naturally occur through mineral reactions in both deep geological carbonation and ocean storage, with the latter only possible above the carbonate compensation depth (~ 3,000 m).’
hydrodynamic trapping; and, (3) geochemical trapping, also referred to as mineral trapping (IPCC, 2005). Mineral trapping generally operates on a longer timeframe (~ 1,000 yr); making it inconsequential when modelling CO₂ fluid flow in the reservoir. Residual CO₂ and solubility trapping mechanisms are considered separately—these provide CO₂ storage capacity along the migration pathway although mechanisms operate over a time span of some 10,000 years. However, the storage potential of residual and solubility trapping within saline aquifers far outweighs the storage potential of structural and stratigraphic traps. In this regard, the maximum solubility of CO₂ in brine, achieved under specific conditions (20 MPa, 40°C, < 10,000 ppm), was shown to translate to an increase in the total storage potential of up to 5.8% (Daniel and Kaldi, 2008).

Containment in CO₂ storage is paramount. Six possible leakage routes relating to CO₂ emanating from subsurface reservoirs were considered in the IPCC report (IPCC, 2005, p35): ‘(1) CO₂ could escape from a trap if its gas pressure exceeded the capillary pressure of the seal; (2) leakage is possible up faults; (3) CO₂ escape is possible through a gap, or thinning of the cap rock, into a higher saline aquifer; (4) the increase in pore pressure as a result of injected CO₂ could cause faults to open and fluid flow pathways to form; (5) escape of CO₂ is possible up poorly cemented wellbores; and (6) CO₂ could dissolve into the formation water, consequently increasing its density and causing it to sink within the reservoir, then to be transported outside of the structural enclosure.’

1.1.4 Carbon dioxide storage options

Carbon dioxide storage options include storage in depleted oil and gas fields, in deep saline aquifers, in enhanced oil (EOR), enhanced gas recovery (EGR) and coal bed methane (ECBM) recovery schemes and in coal beds (IPCC, 2005). Each of these solutions is only economically feasible under certain conditions. For example, the Weyburn oil field project in Canada adopted and demonstrated the technical feasibility of CO₂-based EOR, but was only commercially successful because of the close proximity of a CO₂ source (~ 100 km). Similarly, deep saline formations and depleted oil and gas field storage options are also only feasible under certain economic conditions. Examples of a commercial and pilot project in saline aquifers include the Sleipner Field (Norway) and the Frio projects (U.S.A.), which have planned storage capacities of 20 MtCO₂ and 16 tCO₂, respectively. A successful commercial-scale oil and gas field example is the In-Salah project (Algeria) with 17 MtCO₂ of planned storage capacity. The ECBM option is only in the demonstration phase. Examples of ECBM include the Fenn Big Valley (Canada), Qinshui Basin (China), Yubari (Japan) and Recopol (Poland) projects. Stored tonnages are 200, 150, 200 and 10 tCO₂, respectively. Estimates of the

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4 Hydrodynamic trapping refers to the buoyant flow of CO₂ updip that is restrained by a downdip counter-aquifer flow.
CO₂ global storage capacity for depleted oil and gas fields, ECBM and deep saline formations range from 675 to 900, 3–200 and 1,000–10,000 GtCO₂, respectively. As can be seen, deep saline formations have by far the largest CO₂ storage capacity. However, the estimated storage capacity for depleted oil and gas fields could increase by as much as 25% if undiscovered fields were included (IPCC, 2005).

Geotechnical screening criteria used to assess CO₂ storage in the early 1990s only reflected the first-pass needs of broad desktop studies carried out at the basin/subbasin scale (Bachu, 2000, 2003). For example, regional basin characteristics (tectonic activity, sediment type, geothermal and hydrodynamic regimes), basin resources (hydrocarbons, coal and salt), industry maturity and infrastructure, and societal issues (level of development, economy, environmental concerns, public education and attitudes) were used as basic screening criteria. More geologically focused criteria were incorporated in other assessments (Bradshaw et al., 2002); for example, the study of CO₂ storage in Australian basins included storage capacity, injectivity potential, site details, containment and existing natural resources. The latter set of criteria were superseded by additional criteria, some naturally tailored to the requirements of individual storage site assessments (IPCC, 2005).

In Australia, CO₂-related research was initiated in the GEODISC™ project under the umbrella of the Australian Petroleum Cooperative Research Centre (Bradshaw and Rigg, 2001). In GEODISC™, a CO₂ source-to-sink ranking scheme was used to rank potential CO₂ storage sites and basins, including the Gippsland Basin (Bradshaw et al., 2001b; Bradshaw et al., 2002). A potential storage capacity of 1,600 years was estimated for a storage rate of 100–115 MtCO₂/yr (Bradshaw et al., 2001a); this CO₂ storage-related research has understandably been biased towards basin-scale projects. Currently, a CO₂ storage project that will operate commercially and make use of the saline-aquifer storage option is the Greater Gorgon project (Malek et al., 2004). Also, a depleted gas field in the Otway Basin is being used as a CO₂ storage pilot research project (Cook, 2006). The viability of ECBM and EOR/EGR storage options has yet to be demonstrated in Australia.

1.1.5 Carbon dioxide storage—assessment strategies

Most Australian CO₂ storage site assessments to date have used 2-D seismic data, with a stratigraphic–depositional model being the base geomodel⁵ of choice (Gibson-Poole et al., 2006a, b; Gibson-Poole, 2009). Such assessments have integrated geological modelling of injectivity, containment and storage capacity, economic modelling, numerical flow simulation and, risk and

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⁵ In this study, a base geomodel refers to the starting geological model into which parameters are incorporated.
uncertainty analysis (Figure 1.1). Containment modelling frequently incorporated a geomechanical analysis (including fault reactivation modelling) and, a study of the seal retention capacity and regional hydrodynamic system. Inherent limitations exist when using a sequence stratigraphic–depositional base geomodel as it is constrained by sparse one-dimensional (1-D) well control and interpreted surfaces based on 2-D seismic data. Any subsequent modelling, analysis and simulation are contingent on the base geomodel and populating of reservoir parameters from the loose well and seismic control. Similarly, fault reactivation modelling has relied on fault geometries mostly interpreted from 2-D seismic datasets, as was the case in the Great Australian Bight (Reynolds et al., 2005). Therefore, fault reactivation of small-scale faults, fault arrays and fault zones cannot be modelled comprehensively, even though these often represent areas of probable trap breach; thereby, also undermining the usefulness of 2-D based fault reactivation modelling estimates.

Figure 1.1. Geotechnical workflow for assessing CO₂ storage—sequence stratigraphic–depositional-based approach.
After Gibson-Poole et al. (2006a, b).

An alternative CO₂ storage assessment strategy that would complement the use of sparse well data and 2-D seismic data would be to incorporate seismic attributes extracted from a three-dimensional (3-D) seismic dataset where the geomodel could be populated from a dense grid of data. Use of the seismic attributes would further improve the populating of the sequence stratigraphic–depositional base geomodel, particularly with lateral facies boundaries. For example, spectral decomposition allows the full bandwidth of the 3-D seismic volume to be represented as narrow bandwidth seismic volumes, thus allowing the estimated tuning frequency of reservoir thickness and/or facies and/or fluids to be visualised (Gell et al., 2004)—its utility is demonstrated (Partyka et al., 1999). However, the extent and interconnectivity of the reservoir–aquifer system still could be compromised by the
absence of a 3-D seismic based fault interpretation. An alternative CO\textsubscript{2} storage assessment strategy would be to use a structure-based base geomodel.

A few structure-based base geomodels underpinned by 3-D seismic data and linked to CO\textsubscript{2} storage projects are in use; the Otway Basin (Sharma et al., 2007) and North West Shelf (Malek et al., 2004) are examples. The relevance and impact of faults and the use of a structure-based base geomodel for CO\textsubscript{2} storage studies can be illustrated by making comparisons to petroleum production projects. Faults can cause an unexpected drop in production rate in the later part of a field’s life, as has occurred in many large petroleum fields (Simmons, 2005). The importance of detailed structural interpretations in petroleum production and reservoir engineering projects is well documented outside the Gippsland Basin (Agarwal et al., 1999; Allan and Qing Sun, 2003; Kromah et al., 2005). Also, identification of small-scale fault systems has, for example, led to improved well-plan design and subsequent production improvements in the case of a field in Saudi Arabia (Stenger et al., 2002). Similarly, flow barriers in the form of minor faults have been recognised in large fields of the Gippsland Basin (Hinton et al., 1994; Cousins, 1995; Burnett, 2005). The possible effects of minor faults can be just as significant for CO\textsubscript{2} storage projects. Similarly, the modelled sensitivity of CO\textsubscript{2} injection rates due to the inclusion of faults has also been tested by incorporating random faults and treating these as no-flow boundaries (Cinar et al., 2007); there, the six faults incorporated randomly within the model volume caused a drop in the CO\textsubscript{2} injection rate of between 14 and 16\%. By analogy, the knowledge of how small-scale faults will affect CO\textsubscript{2} migration could ultimately improve the positioning of CO\textsubscript{2} injectors and aid monitoring of the CO\textsubscript{2} plume, amongst other operational advantages.

A structure-based base geomodel can incorporate fault-slip variation (i.e. potentially down to subseismic scale), fault arrays (i.e. multiple branch lines present), detail at the fault tip, presence of fault zones (i.e. inclusion of fault strands/planes) and allow an assessment to be made regarding their impact on CO\textsubscript{2} storage. These subtle structural features are all at or below the limits of seismic resolution (e.g. \( \sim 10 \) m at 2 km for vertical resolution) and can affect the extent and interconnectivity of the reservoir–aquifer system and, potentially affect CO\textsubscript{2} injection schemes (Figure 1.2a). It follows that CO\textsubscript{2} injection schemes require that the reservoir–aquifer system be reasonably extensive and/or interconnected, to avoid premature pressure buildup in the reservoir. In comparison, petroleum production withdraws fluids from the reservoir, so that the issue of extent and interconnectivity is still relevant but less critical to the success of the project. The IPCC report did not consider the extent and/or interconnectivity of the system to be a critical parameter for screening CO\textsubscript{2} storage sites (IPCC, 2005). Instead, the report indicated that an increase in pore pressure from injected CO\textsubscript{2} could
cause faults to open and possibly allow new subsurface fluid pathways to form. However, the extent and interconnectivity of the reservoir system has recently been shown to be of importance with regards to the injectivity performance and economics of a CO₂ storage project, as demonstrated through modelling (Sayers et al., 2006; Cinar et al., 2007).

The advantage of emphasising a structural interpretation and using a structure-based base geomodel up front is that it allows the impact of faults to be estimated early in a geomodelling assessment. This advantage is further emphasised by the fact that automatic fault extraction algorithms (e.g. GeoFrame’s ant tracking; Cox and Seitz, 2007) are now available, and these algorithms have fast turnaround times for interpreting faults in 3-D seismic cubes. Projects can be abandoned early in the project feasibility stage, should the level of faulting be unacceptable. In comparison, building complex sequence stratigraphic–depositional models is time-consuming, and models are constrained by the wide-spaced 1-D well control. However, stratigraphic–depositional models would be incorporated when a project progresses forwards (see Section 1.4). The structural-based base geomodel, and the workflow associated with it, form the foundation of the research strategy in this study.

1.2 Research objectives

The research in this study is based on the premise that faults, particularly minor faults, may directly impact the viability of CO₂ storage within carbon dioxide storage leads (CO₂SL)—this study is aimed at interpreting and modelling faults to address this premise (Figure 1.2). The impact relates to the fact that faults can act as a conduit to fluid-flow under certain geological conditions and as a baffle to fluid-flow in others (Barton et al., 1997). In the case of a reservoir being subjected to increased pore pressure consequent to CO₂ injection, the propensity of CO₂ to escape up or across a fault is very much dependent on the sealing across-fault, as well as the fault’s orientation relative to the maximum shear stress azimuth (Figure 1.2a; Streit and Hillis, 2002). Hence, a need exists to carry out a fault seal analysis that incorporates juxtaposition analysis, fault damage and reactivation modelling (Jones and Hillis, 2003).

The other related structural features that may also affect trap integrity within CO₂SLs, and considered in this study include, branch lines in fault arrays (Figures 1.2g–h), fault throw families (Figure 1.2f), fault zones (Figure 1.2i) and where fault tips arrest (i.e. the upper tip-line bound, Figure

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6 CO₂ storage lead: a potential location for CO₂ storage that has a trap(s) identified but that has not been comprehensively investigated (formally defined in Chapter 5).

7 A line of intersection between a master fault and a synthetic splay or between two segments of a multi-strand fault (Walsh et al., 1999).
1.2e). Establishing the latter of these is important because of the potential link between local hiatuses—unconformities and fault chronology (Figure 1.2b; discussed in the next two paragraphs). This potential link forms a fundamental part of this study, as any fault-seal breach that may occur as a result of increased pore pressure consequent to CO₂ injection will be at least partly dependent on whether the fault tip arrests at the base seal proper, within any potential thief zones, or propagates through the seal itself (Figure 1.2e). It is probable that a lower pore pressure (ΔP) would induce fault breach, should the fault tip extend into the seal as opposed to arresting at the base of the seal, thereby undermining the trap integrity.

Hiatuses and unconformities, linked to uplift and/or erosion, are often associated with a high percentage of fault tips that arrest at the unconformity (Sayers, 1993; Sayers et al., 2001). Other researchers have also demonstrated that fault sets can be disassociated across unconformities and/or megasequences (Lyon et al., 2004). Structural geometry and reconstruction principles have also been used to demonstrate the range of interpretations that are plausible causes of disassociated fault sets (Boult and Freeman, 2007). Disassociated fault sets are often seen when crossing megasequence boundaries, particularly where the basin’s tectonic phase changes (e.g. from syn-rift to rift-drift or syn-rift to sag phase). For example, faults F4.1–4.2 in layer 4 (Figure 1.2b) are disassociated from faults F3.1–3.3 in layer 3 where the fault tips arrest at unconformities H-4 and H-3, respectively.

It is also assumed, in this study, that not all pre-existing faults are reactivated when a basin undergoes multiple stretching and/or compression events; if so, remnants of structural events could potentially be identified by assessing where fault tips arrest relative to unconformities, minor or other. For example, early formation of a trap-door structure followed by differential compaction and/or drape could be misinterpreted as fault reactivation (e.g. compare faults F1.1–F2.1 and F2.2—Figure 1.2b). If faults F2.1 and F2.2 are not reactivated, then the faults and unconformity H-2 may only be

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8 The term hiatus is frequently used in this study and thus requires a formal definition. A hiatus is taken here as ‘A break or interuption in the continuity of the geologic record, such as the absence in a stratigraphic sequence of rocks that would normally be present but either were never deposited or were eroded before deposition of the overlying beds’ (Jackson, 1997, p298). The term 'lacuna' could also be used to more closely represent relevant geological features discussed in this study, where a lacuna is 'a period during which sedimentation was either nil or, more likely, was replaced by erosion' (Jackson, 1997, p353). However, to avoid confusion, the term 'hiatus' has been consistently used.

9 An ‘unconformity’ is defined as ‘the structural relationship between rock strata in contact, characterized by a lack of continuity in deposition, and corresponding to a period of nondeposition, weathering, or erosion (either subaerial or subaqueous) prior to the deposition of the younger beds, and often (but not always) marked by absence of parallelism between the strata ... ’ (Jackson, 1997, p689).

10 Herein, ΔP refers to the maximum increase in pore pressure that a fault can sustain, consequent to CO₂ injection.

11 A megasequence is a stratigraphic sequence of rock with a time span greater than 50 My and controlled by tectono-eustasy (Emery and Myers, 1996). The base and top of a megasequence are bounded by regional unconformities.
representative of a period of subtle structural readjustment within the basin. Alternatively, evidence from a fault like F1.2 (Figure 1.2b) would suggest a distinct structural event and/or localised faulting at basin hinge-lines. In such cases, evidence from regional fault systems and unconformities helps constrain the interpretation one way or the other. The latter is particularly relevant to trap integrity in relation to CO₂ storage since CO₂ will migrate through buoyancy drive to the base of a seal and then laterally along it. It follows that any fault tips in proximity to a CO₂ migration pathway could increase the breach-of-trap risk via top-seal by-pass, possibly via crack-tips ahead of the fault tip.

Improvements to the understanding of basin evolution and trap formation should naturally flow from the above objectives and, consequently, help in structurally assessing a CO₂SL. The primary objectives of this study are encapsulated through the research questions listed below.

- How does the inclusion of fault chronology, minor faults, fault arrays and fault zones affect the assessment of trap formation and integrity, as well as the modelling of fault seal integrity?
- How does the potential link between where fault tips arrest and where unconformities occur improve the understanding of fault chronology and basin evolution?
- How can a structure-based base geomodel complement the more common approach of modelling that emphasises the use of a sequence stratigraphic–depositional base geomodel?
- Is increasing the level of detail (e.g. by introducing minor faults) in a structural model warranted when carrying out a fault seal analysis and, if so, under what circumstances?
- What impact do the above have on geotechnical assessments of CO₂SLs?

### 1.3 Study area rationale

The offshore and eastern portion of the Gippsland Basin was chosen as the study area over other Australian basins screened (Browse, Malita Graben, Otway and Exmouth) for the following reasons.

(i) The availability of a high quality 3-D seismic dataset was sought; such a dataset enables greater flexibility in using seismic attributes to interpret fault systems. In addition, complex fault patterns are known to be present in the Gippsland Basin (Power, 2003). By way of an example from the Tuna Field, the complexity of fault arrays as interpreted from 2-D regional seismic data (Maung and Cadman, 1989; Maung, 1992), was superseded by the interpretation of a 3-D seismic dataset (Power, 2003). The structural interpretation was considerably altered through the understanding of the petroleum migration pathways (i.e. minor faults, relay ramps and fault splays), as could be expected.
Figure 1.2. Conceptual representation of key structural and stratigraphic factors critical to the fault/trap integrity in CO₂ SIs (a) map view, (b-1) cross-sectional views.

(a) Disassociation of fault sets. (c) Gipsiland Basin stratigraphy. (d) Juxtaposition of reservoir and seal, and shale smear. (e) fault tips and upper tip-line bound. (f) fault throw families, (g) hard-linked branch line, (h) soft-linked branch line and (i) fault zones.
(ii) Traps for CO₂ storage that are partly or wholly reliant on fault seal integrity were required so that the objectives of the research could be met. Additionally, lengthy CO₂ migration pathways were sought in order to capture a reasonable spread of fault networks. The eastern part of the Gippsland Basin was also found to be favourable in that the increased shale content in the Golden and Halibut Subgroups relative to the western part of the basin (Partridge, 1999) would increase the likeliness of favourable sand-on-shale windows across-fault; thus, making the study results pertinent to CO₂ storage.

(iii) The study area was chosen to test structural complexities related to CO₂ storage in the more favourable part of the sedimentary section in order to make assertions regarding CCS in less favourable sections and basins. For example, a CCS project envisioning CO₂ storage in the offshore Gippsland Basin, has a lower geotechnical risk compared to CO₂ storage onshore, as the upper reservoirs of the Latrobe Group are known to have both high porosity (Φ > 20%) and permeability (K > 1 darcy, Partridge, 1999). However, logistics are less favourable with projects being more capital-intensive when operating offshore (Hooper et al., 2005).

(iv) The availability of CCS-related industry-based feasibility studies in the offshore component of the basin was sought to facilitate research synergies and relevance. For example, two onshore Gippsland Latrobe Valley projects exist; the Monash and Hazelwood Power Station CO₂ capture projects (Hooper and Feron, 2006; Hooper et al., 2006). Additionally, the majority of CO₂ storage projects globally have previously been assessed with the view to store CO₂ emissions from coal-based power stations (IPCC, 2005). However, CO₂ storage from CO₂-rich petroleum gas fields had also proven to be a viable option (Sleipner Field, Norway, IPCC, 2005). The latter option could be viable in the Gippsland Basin, as gas discoveries there have associated CO₂ concentrations ranging up to 30% (O’Brien et al., 2008).

The eastern Gippsland Basin was also chosen as the study area in order to focus structural research towards CO₂ injection within aquifers that are located downdip of petroleum fields, many at

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12 For example, the first international release of offshore assessment permits for CO₂ storage was announced by the Honorable Martin Ferguson AM, MP, in Canberra on the 27th March 2009. Ten permits in all were released over the Gippsland and Otway Basins, and the Torquay, Vlaming, and Petrel Subbasins. Three permits (GIP-01, -02, -03) were released in the Gippsland Basin (Kalinowski, 2009); these permits are located on the southern margin of the Basin where potential conflict with the petroleum exploration will be minimised. The acreage is released under the Offshore Petroleum and Greenhouse Gas Storage Act 2006 that came into force on the 21st Nov. 2008. Further details regarding the permits were addressed at the APPEA conference held in Darwin in June 2009 (Barrymore and Mathison, 2009).

13 Injection in saline aquifers is assumed as fresh water aquifers may be off limits to CCS; ‘aquifer’ is used herein.
different stages of depletion. Consideration of CO\textsubscript{2} storage that combines near-depleted petroleum fields with an associated downdip storage option is optimal because of the incremental CO\textsubscript{2} storage and trapping potential. The view is taken in this study that it may be preferable to commence CO\textsubscript{2} storage in aquifers downdip, but during the mid-life of a petroleum field, for the following reasons.

(i) Injection into aquifers can involve CO\textsubscript{2} migration along lengthy pathways that are relatively undrilled. However, aquifers that are located well away from production are also located away, in many cases, from good data control. In comparison, aquifers located adjacent to near-depleted petroleum fields are better constrained.

(ii) Near-depleted petroleum fields can be used as CO\textsubscript{2} storage sites (Bradshaw et al., 2001b; Bradshaw et al., 2002). Data control is generally good within petroleum fields, although their CO\textsubscript{2} storage potential is limited and practical difficulties can arise if the initial reservoir properties are not reversible following production. For example, unfavourable changes to the reservoir can be caused by poroelastic effects (Streit and Hillis, 2002), or stress changes that alter fluid pathways, as has been demonstrated in giant field production (Rahim et al., 2003). In such a case, it is important to know how much pressure drawdown is acceptable before significant reservoir damage occurs in addition to understanding changes to the stress path that ultimately alter fluid flow. Therefore, using a downdip option in combination with a near-depleted petroleum field is more conducive to favourable outcomes in relation to a CO\textsubscript{2} storage project and, to the research in this study being pertinent. The above view is corroborated by the success of a near-depleted field in the North Sea that has performed to expectation following two years of CO\textsubscript{2} injection (Bert et al., 2009).

(iii) The existence of an updip structural trap, as is present in many petroleum fields of the Gippsland Basin, is important in that it is more conducive to CO\textsubscript{2} storage. The associated research will also be more acceptable and relevant to industry and government because such storage is more geographically contained and, less likely to interfere with petroleum exploration and production. For example, the competing effects of pressure drawdown from multiple fields as a result of petroleum production were demonstrated to be an important consideration for designing acceptable CO\textsubscript{2} injector programs (Hortle et al., 2010).

1.4 Study workflow

The workflow relating to the objectives described is illustrated in Figure 1.3, the associated methodology addressed in subsequent chapters and, summary given in what follows.
(i) Regional unconformities are reviewed (Chapter 2); local unconformities and hiatuses in the Latrobe and Seaspray Groups are then interpreted using age-depth plots, following which these are interpreted across the 3-D seismic data (Chapter 3). A structural interpretation is undertaken in the time domain giving particular attention to the interpretation of the fault trace, fault tip, fault arrays, fault zones and fault throw and, also determining where fault tips arrest in relation to unconformities. Sections, TWT slices and variance-attribute slices are used in the seismic interpretation. The fault chronology, fault mode, fault regime and fault network are determined; the basin evolution and trap formation pertinent to the study area are presented.

(ii) A fault seal analysis is best done in the depth domain\(^\text{14}\) where a depth-converted 3-D seismic dataset allows the juxtaposition of reservoir and seal across-fault to be compared directly against reservoir thicknesses. A new depth conversion program is developed to depth-convert both the horizons and faults interpreted (i.e. stored as ASCII), but also the seismic SEG\(^Y\)\(^\text{15}\) data (Chapter 4). The depth-converted 3-D seismic dataset is then exported out of the seismic interpretation platform (\textit{GeoFrame}™) to the fault seal analysis platform (\textit{TrapTester}™)\(^\text{16}\).

(iii) \(\text{CO}_2\)SLs are identified in the study area and a new terminology is introduced to screen them and facilitate their assessment (Chapter 5). Further, the top three screened \(\text{CO}_2\)SLs form the basis upon which a study subarea is defined.

(iv) A structural model is built (in depth) from the interpretation; it captures details of the fault throw and branch lines (Chapter 6). Then, the reservoir and seal intervals are assessed in 11 wells in order to define sand and shale lines pertinent to the generation of \(V_{\text{shale}}\) curves; the latter are then projected onto the fault planes to establish the juxtaposition of the intervals across-fault. The shale gouge ratio (SGR)\(^\text{17}\) is generated for each fault plane and then used to estimate the sealing potential across-fault for each of the reservoir intervals. The propensity of faults to reactivate, as a result of a pore pressure increase consequent to \(\text{CO}_2\) injection, is modelled (Chapter 7). The fault reactivation modelling is carried out on different fault-throw families\(^\text{18}\) with differing branch-line combinations analysed.

\(^{14}\) The depth domain refers to the seismic being interpreted in depth from the original seismic volume that was acquired and processed in the time domain.

\(^{15}\) SEG\(^Y\) refers to magnetic tape format ‘Y’ as defined by the Society of Geophysics (Barry et al., 1975).

\(^{16}\) \textit{TrapTester}™ is the fault seal analysis software owned by the UK-based company Badley Geoscience Ltd.

\(^{17}\) The shale gouge ratio estimates the sealing potential by shale smear (further explained in Chapter 6).

\(^{18}\) The term fault-throw family is used herein to differentiate faults according to throw (major, minor); see Figure 1.2f.
Figure 1.3. Geotechnical workflow for assessing CO₂ storage sites in this study—structure-based approach.
Geological modelling (top right) incorporates a sequence stratigraphic interpretation and, individual reservoir, seal and facies analysis, followed by populating of relevant parameters (e.g. porosity) into the geocellular model (i.e. one value for each parameter assigned to each cell in the model).
CHAPTER 2—REGIONAL GEOLOGY

2.1 Introduction
The Gippsland Basin extends both onshore and offshore in southeastern Victoria and is approximately bounded by longitudes 147°–149°30’E and latitudes 37°48’–39°48’S (Figure 2.1). The basin extends over an area of about 20,000–30,000 km², but would have covered up to 45,000 km² in the Cainozoic (Partridge, 1999). Water depths range up to 2 km but are generally less than 200 m. The basin is in the mature phase of the petroleum exploration cycle, following 40 years of production history with production in excess of one billion barrels of oil.

Well established regional unconformities include the Otway and Latrobe Unconformities (Partridge, 1999; Bernecker and Partridge, 2001) while other regionally significant unconformities include the Longtom, Seahorse, Marlin and Mackerel Unconformities (Partridge, 1999). Probable eustatic-driven hiatuses that are subregional in extent include the Marshall Paraconformity and the Bullseye Karst. Early interpretations defined the regional structure and chronology of tectonic events (Maung and Cadman, 1989; Maung, 1992); later, a regional seismic interpretation incorporated palinspastic reconstructions, deep seismic data (10 second) and one localised 3-D seismic grid to refine the interpretation (Power, 2003). Still later work established the present-day stress field (Nelson and Hillis, 2005).

In this chapter, the chrono, litho- and volcanostratigraphic schemes are reviewed to determine which scheme to adopt in this study. A review of the regional and local hiatuses, correlative unconformities, fault chronology and their associated structural, eustatic and/or sedimentological drivers follows. The review describes the basin’s structural framework, evolution and trap formation within, but also describes any structure-related anomalies that may have a direct bearing on CO₂ storage. No emphasis is placed on describing the basin’s lithostratigraphy for reasons previously outlined (Chapter 1). Spore-pollen zones are referred to but not elaborated on; a palynology chart is provided for further reference (Appendix 1—Figure A1.2). Any place names or structural elements referred to are illustrated in Figure 2.1.

2.2 Stratigraphy and bounding unconformities

2.2.1 Nomenclature
Chronostratigraphic correlations have been frequently used in southeastern Australia (McGowran et al. 1997; McGowran et al., 2004), the onshore Gippsland Basin (Holdgate and Sluiter, 1991; Holdgate and Gallagher, 1997; McGowran et al., 1997) and whilst chronostratigraphic charts have
NOTE:
This figure is included on page 18 of the print copy of the thesis held in the University of Adelaide Library.
been compiled for the offshore region (Norvick et al., 2001; Holdgate et al., 2003a; Norvick, 2005), it is more common for lithostratigraphic or palynology-based tops to be used offshore (Partridge, 1999; Johnstone et al., 2001) instead of seismic-based chronostratigraphic boundaries. Several combined lithostratigraphic–palynology-based schemes that cover both the offshore Latrobe and Seaspray Groups exist (Bernecker and Partridge, 2001; Bernecker et al., 2001; Wallace et al., 2002; Bernecker and Partridge, 2005), of which, Bernecker and Partridge’s (2001) scheme was adopted for pre-Early Oligocene stratigraphy in this study (Figure 2.2). Accordingly, lithostratigraphic tops are available in government databases (www.ga.gov.au—data and applications, energy, applications, petroleum wells applications), whilst chronostratigraphic tops are less common.

For the Seaspray Group, there are three common interpretation schemes available.

(i) Bernecker et al. (1997) subdivided the Seaspray Group into four units (I to IV) based on core and log analyses, while Partridge (1999) subdivided the group into four lithostratigraphic units: the Swordfish, Conger, Cod and Whiting Formations. These schemes resulted in the lithostratigraphic chart of Bernecker and Partridge (2001); Partridge’s (1999) nomenclature for the Seaspray Group is not consistently used by industry, nor has it been adopted in post-1999 well completion reports (WCR), nevertheless it has a number of positive characteristics. First, the subdivision of the Seaspray Group is more detailed than that commonly used in WCRs, which normally only differentiate the Gippsland Limestone and Lakes Entrance Formation. Second, Partridge’s subdivisions can have direct geological significance; for example, the base of the Conger Formation coincides with a sharp increase in the rate of deposition as interpreted in this study from age-depth plots (see Chapter 3). However, the primary advantage is that formation tops data based on Partridge’s (1999) scheme are publicly available; in this study, the scheme was adopted for post-Oligocene stratigraphy (Figure 2.2).

(ii) Feary and Loutit (1998) subdivided the Seaspray Group into four megasequences and 16 sequences using a sequence-stratigraphic approach that incorporated log data (Figure 2.3).

(iii) Holdgate et al. (2000) subdivided the Seaspray Group into the Angler, Albacore and Hapuku Subgroups. This subdivision was based on onshore stratigraphy and then extended to the offshore.

2.2.2 Strzelecki Group and Otway Unconformity

The Strzelecki Group ranges in age from Berriasian to Albian (c. 132–98 Ma); it was comprehensively described from onshore occurrences (Constantine, 2001) and also intersected offshore at depths of less than 2 km under the Northern Strzelecki Terrace. The dominant lithology is
sandstone with less common greywacke, mudstone, conglomerate and coal; interpreted depositional palaeoenvironments range from fluvial to lacustrine to rift valley-fill. The Strzelecki Group is subdivided into a lower predominantly quartz clastic section, overlain by an upper volcaniclastic upper section. An unnamed regional unconformity that separates these two megasequences was originally dated at 114 +/-4 Ma (Gilbert and Hill, 1994), but was later redated at c. 125 Ma (Partridge, 1999). The unconformity bounds large-scale faults and associated syn-rift section (Norvick et al., 2001). Several other sequence boundaries have also been interpreted within the Strzelecki Group; e.g. strz1, strz2a, strz2b (Power et al., 2001), IS (Willcox et al., 1992). The group’s thickness is in excess of 5 s TWT (Figure 2.4a).

The Otway Unconformity bounds the top of the Strzelecki Group and the associated hiatus spans from 98 to 92 Ma (Figure 2.2); the term top Strzelecki Unconformity has also been used (Lowry and Longley, 1991). The unconformity straddles the P. mawsonii and H. uniforma spore-pollen zones of the late Cenomanian–Turonian. Additionally, a number of sequence boundary/horizon names exist; for example, Strz2c (Power et al., 2001), TS (Willcox et al., 1992). Lastly, the unconformity merges with the overlying Longtom and Seahorse Unconformities over the Northern Strzelecki Terrace and Platform areas.

2.2.3 Latrobe Group (Emperor Subgroup and Longtom Unconformity)

The Emperor Subgroup was defined and named by Partridge (1999), subsequently published in Bernecker and Partridge (2001), and has since been fairly widely accepted (Thomas et al., 2003). The subgroup ranges from Late Cenomanian to top Turonian (c. 92.5–89 Ma); the base rests on the Otway Unconformity and is top-bounded by the Longtom Unconformity (Figure 2.2)—it incorporates the Kersop Arkose, Admiral Formation, Kipper Shale and Longtom Formation¹⁹. Deep water equivalents of these units have been described (Bernecker et al., 2001). The dominant lithologies of the Kersop Arkose are sandstone and arkose, interbedded with less abundant shale and siltstone.

The main palaeodepositional environment is interpreted to be an alluvial fan; the Kersop Arkose is restricted to structural hinge lines and escarpments along the basin margins. The dominant lithology of the Admiral Formation is sandstone, interbedded with shale and siltstone. The main palaeodepositional environment is interpreted to be an alluvial fan, with common fluvial-alluvial to lacustrine systems also present (Partridge, 1999). The dominant lithologies of the Kipper Shale are siltstone and shale, with some minor sandstone; a lacustrine palaeodepositional environment is

¹⁹ The Curlip Formation term initially used was later replaced by the Longtom Formation (Bernecker and Partridge 2005; Pers. Comm. Tom Bernecker July 2006).
Figure 2.2. Events chart of the Gippsland Basin - stratigraphy, basin phases and petroleum system elements.

The Longtom Formation is predominantly sandstone and conglomerate; interpreted palaeodepositional environments vary from lacustrine, shoreline to deltaic to fluvial, interspersed with coal swamps and alluvial fans. The subgroup’s thickness ranges up to 1.7 s TWT (Figure 2.4b). The Longtom Unconformity was recognised following an interpretation of the biostratigraphy in Longtom-1 (Partridge, 1999). Previously, the Longtom and Seahorse Unconformities had been interpreted. The Longtom Formation is predominantly sandstone and conglomerate; interpreted palaeodepositional environments vary from lacustrine, shoreline to deltaic to fluvial, interspersed with coal swamps and alluvial fans. The subgroup’s thickness ranges up to 1.7 s TWT (Figure 2.4b).
interpreted as one unconformity (Lowry and Longley, 1991; Willcox et al., 1992): this combined unconformity was also referred to as the *N. senectus* Unconformity (Lowry and Longley, 1991). Since 1999, the new interpretation has been accepted by a number of authors (Norvick et al., 2001; Power et al., 2001). The associated hiatus ranges in age from Late Turonian to Middle Santonian; c. 89–86 Ma (Partridge, 1999; Bernecker and Partridge, 2001; Bernecker and Partridge, 2005). The unconformity straddles the *P. mawsonii* and *T. apoxyexinus* spore-pollen zones, although two subzones of the *P. mawsonii* spore-pollen zone and, the oldest part of the *T. apoxyexinus* zone may be absent (Partridge, 1999).

**2.2.4 Latrobe Group (Golden Beach Subgroup and Seahorse Unconformity)**

The Golden Beach Subgroup was originally classified as a group or megasequence (Gilbert and Hill, 1994), reranked by Partridge (1999) and subsequently published in Bernecker and Partridge (2001). The Golden Beach Subgroup ranges in age from Santonian to Late Campanian; c. 86–75 Ma (Figure 2.2). The subgroup incorporates the Anemone and Chimaera Formations; deep-water equivalents have also been described (Bernecker et al., 2001). The dominant lithology in the proximal Chimaera Formation is sandstone, with minor shale and coal. The sandstone, shale and coal were deposited in a coastal plain palaeoenvironment, which incorporates occasional marine incursions (Partridge, 1999). The Chimaera Formation is often overlain by volcanics (see Section 2.3). The distal deeper water equivalent of the Chimaera Formation is the Anemone Formation. The Anemone Formation comprises siltstone and mudstone interbedded with thin sandstone beds that were deposited in a marine palaeoenvironment (Partridge, 1999). The subgroup thickens up to 1.5 s TWT (Figure 2.4c).

The Seahorse Unconformity bounds the top of the Golden Beach Subgroup. The unconformity straddles the *T. lilliei* and *F. longus* spore-pollen zone boundary, making it latest Campanian to basal Maastrichtian; c. 75–68 Ma (Partridge, 1999; Bernecker and Partridge, 2001; Bernecker and Partridge, 2005). The unconformity is not readily identified from well logs; however, it is often associated with volcanics that overlie the Chimaera Formation (see Section 2.3).

**2.2.5 Latrobe Group (Halibut Subgroup, Mackerel and Marlin Unconformities)**

The Halibut Subgroup term was defined and introduced by Partridge (1999), subsequently published by Bernecker and Partridge (2001) and, has since been fairly widely accepted (Thomas et al., 2003). The subgroup ranges in age from Late Campanian to latest Early Eocene; c. 75–49.5 Ma (Figure 2.2). It overlies the Seahorse Unconformity and is top-bounded by the Marlin Unconformity. The Halibut Subgroup incorporates the Volador Formation, Kate Shale and the Kingfish, Mackerel and Barracouta Formations; the latter three are bounded by the Mackerel Unconformity, which is overlain
NOTE:
This figure is included on page 25 of the print copy of the thesis held in the University of Adelaide Library.
by the Flounder Formation. Specifics of individual formations are listed below and are based on the work of Partridge (1999), unless otherwise stated. The lower part of the Halibut Subgroup thickens up to 1.3 s TWT (Figure 2.4d).

(i) The Volador Formation ranges in age from Late Campanian to upper Maastrichtian, c. 75–66 Ma. The formation consists of interbedded sandstone, shale and coal. A coastal plain palaeodepositional environment, with some marine incursions, is interpreted for this unit. 

(ii) The Kate Shale ranges in age from Late Maastrichtian to Early Palaeocene, c. 67–63 Ma. The maximum development of the Kate Shale is associated with a maximum flooding surface, c. 63 Ma.

(iii) The Barracouta Formation ranges in age from upper Campanian to latest Early Eocene, c. 75–49.5 Ma, and is present in the western, proximal part of the basin. The dominant lithology is sandstone, interbedded with siltstone and mudstone, likely deposited in nonmarine upper coastal to alluvial plain palaeoenvironments.

(iv) The Kingfish Formation ranges in age from lower Palaeocene to latest Early Eocene, c. 66–49.5 Ma. The dominant lithology is sandstone with secondary siltstone, shale and coal. The presence of coal distinguishes the Kingfish Formation from the Barracouta Formation; deposition is interpreted to have been in a coastal plain palaeoenvironment, with some marine incursions.

(v) The Kingfish Formation merges into the Mackerel Formation basinwards, with the latter ranging from middle Palaeocene to Early Eocene; c. 63–51.5 Ma. The dominant lithology of the Mackerel Formation is sandstone, with secondary siltstone and shale, all deposited on a siliciclastic shelf.

(vi) The Flounder Formation is latest Early Eocene, c. 51.5–49.5 Ma. It is predominantly muddy siltstone to mudstone, interbedded with minor sandstone. Occasionally, the percentage of sandstone ranges up to 85% (e.g. Kahawai-1). A submarine channel, or an incised valley fill, in an inner shelf setting has been proposed recently as a depositional setting (Ross, 2004).

The Mackerel Unconformity (Figure 2.2) has previously been referred to as the M. diversus unconformity (Lowry and Longley, 1991), lower Marlin Channel (Brown, 1986 cited in Partridge, 1999) and the top of the lower M. diversus (Johnstone et al., 2001). Holdgate et al. (2003a) referred to it as the Mackerel Unconformity and interpreted its presence from the absence of the middle M. diversus zone. However, the term Mackerel Unconformity is not widely used; Partridge (1999) refers
Chapter 2—Regional geology

to it indirectly, as the base of the Marlin or Tuna-Flounder Channels, but does not actually use the term. The unconformity does coincide with the base of the Marlin and Tuna-Flounder Channels, where the latter are present. The Barracouta, Kingfish and Mackerel Formations subcrop the Mackerel Unconformity in places; the unconformity becomes a disconformity elsewhere and, is then difficult to differentiate from the overlying unconformities where these merge within a thin-to-condensed section. The unconformity’s associated hiatus reflects the absence of the *P. asperopolis* and lower *N. asperus* spore-pollen zones; c. 52–51 Ma (Partridge, 1999).

Partridge’s (1999) defined Marlin Unconformity (Figure 2.2) has become accepted terminology by some (Power et al., 2003). Lowry and Longley (1991) previously referred to the unconformity as the middle *M. diversus* unconformity. However, some authors do not recognise it, believing it to be equivalent to the overlying Latrobe Unconformity (Holdgate et al., 2003a), as was the case earlier (Duddy and Green, 1992; Willcox et al., 1992; Holdgate et al., 2003). The unconformity bounds the top of the Marlin and Tuna-Flounder Channels or equivalent upper Halibut–lower Cobia Subgroups boundary. A short hiatus has been estimated at c. 49.5 Ma (Partridge, 1999), differing from the later estimate at c. 46.5 Ma (Bernecker and Partridge, 2001; Bernecker and Partridge, 2005). Only a thin-to-condensed section of Cobia Subgroup overlies the Marlin Unconformity surface, making the unconformity difficult to differentiate from the Mackerel and Latrobe Unconformities. The Marlin Unconformity can be differentiated outside of the channels from changes in the spore-pollen floras, sedimentation rate and distribution of coal (Partridge, 1999).

### 2.2.6 Latrobe Group (Cobia Subgroup and Latrobe Unconformity)

The Cobia Subgroup is Middle Eocene to Early Oligocene (c. 46–33 Ma) and comprises the Turrum, Gurnard and Burong Formations (Figure 2.2). These units are predominantly thin-to-condensed across the greater part of the basin. The Turrum Formation consists mainly of mudstone and shale, with minor sandstone, and is likely to be a submarine channel-fill. The lithology of the Gurnard Formation varies from glauconitic sandstone to mudstone, deposited in a shallow marine palaeoenvironment. The proximal Burong Formation is comprised of interbedded sandstone, siltstone, shale and coal, deposited in a lower coastal plain palaeoenvironment.

The Latrobe Unconformity was first recognised from outcrops in the mid-1960s (Partridge, 1999; Holdgate et al., 2003); it is occasionally referred to by horizon names; for example, TOL (Power et al., 2003) or TL (Willcox et al., 1992) and, has also been referred to as the Oligocene disconformity (Lowry and Longley, 1991). The Latrobe Unconformity forms the lower boundary of the carbonate wedge of the overlying Seaspray Group (Figure 2.2). Subcrop units include the Cobia Subgroup’s thin-to-condensed marine deposits, the Halibut Subgroup’s channel fills and the petroleum-bearing
reservoirs of the Barracouta, Kingfish and Mackerel Formations. The associated hiatus straddles the upper *P. tuberculatus* and upper *N. asperus* spore-pollen zones, and the equivalent *J. foraminifera* zone (Taylor, 1966) and, is diachronous across the Early Oligocene; c. 36–32 Ma (Partridge, 1999; Bernecker and Partridge, 2001). In places, the Latrobe Unconformity can merge with another hiatus offshore (the Marshall Paraconformity). The unconformity can be difficult to differentiate from the underlying Marlin and Mackerel Unconformities because of the thin-to-condensed Cobia Subgroup section and the merging of the unconformities. The Latrobe Unconformity is generally easy to recognise from logs and/or seismic data.

### 2.2.7 Seaspray Group (Marshall Paraconformity and Bullseye Karst)

The offshore Seaspray Group comprises the Swordfish, Conger, Cod and Whiting Formations (Partridge, 1999; Figure 2.2); alternative terms routinely used in WCRs include the Gippsland Limestone and Lakes Entrance Formation in the onshore regions. Other alternative terms include the Hapuku, Albacore and Angler Subgroups; the terms are used to represent proximal and distal facies present in both the onshore and offshore regions (Holdgate and Gallagher, 1997; Power, 2003). The Seaspray Group has also been divided, offshore, into 16 seismic sequences (Figure 2.3; Feary and Loutit, 1998); these can be keyed to Partridge’s (1999) and Holdgate et al.’s (2000) classifications. The Seaspray Group stratigraphy is important for differentiating fault chronology but not central to the study; as such, the formations are described briefly in what follows against comparative schemes; how they link to the seismic horizons defined in the study is described in Chapter 3.

The Swordfish Formation is Early Oligocene to latest Early Miocene; c. 36–16.5 Ma (Figure 2.2); it is dominantly hemipelagic microfossiliferous calcareous mudstone, with occasional silt and bioclasts present. The palaeodepositional environments are interpreted to vary from coastal plain, through shallow carbonate shelf, upper slope turbiditic and out to hemipelagic marine. The Swordfish Formation is equivalent to seismic sequence TL1 (Duff et al., 1991), petrographic core-based Unit I (Bernecker et al., 1997), seismic sequences 1–2 and part of sequence 3 (Feary and Loutit, 1998) and basal Angler Subgroup–zonules D and J (Holdgate et al., 2000).

A minor unconformity that tops the Swordfish Formation has also been mapped as the base of sequence SA4 (Feary and Loutit, 1998), the purple reflector (Holdgate et al., 2000) and base of unit II (Bernecker et al., 1997). The unconformity is also referred to as the mid-Miocene marker (Johnstone et al., 2001; Hart et al., 2006).

The Marshall Paraconformity occurs within the Swordfish Formation and spans part of the Oligocene (Figure 2.2); c. 32–24 Ma (Partridge, 1999). The paraconformity is recognised by the absence of
foraminiferal zone I-2, by the presence of most of foraminifera zones I-1 and J-1 and the late middle
*P. tuberculatus* spore-pollen zone (Partridge, 1999), and has been inferred in Flounder-5 by the
absence of the G, H1, and H2 foraminifera zones (ESSO, 1975), although there is some debate
about whether this interval was adequately sampled. The paraconformity has been occasionally
referred to as the Cobia event (Partridge, 1999). Seismic recognition of the paraconformity is difficult
because of its presence within a thin-to-condensed section with poor acoustic expression. The
paraconformity had previously been correlated to a global sea-level fall and ascribed to a tectonic
driver (Findlay, 1980). However, this interpretation was later discredited, as a global highstand event
was demonstrated (Carter, 1985). Eight transgressions in Victoria’s Latrobe Valley were later
demonstrated within the hiatus observed offshore, confirming that the paraconformity coincides with
a highstand (Holdgate and Sluiter, 1991). Therefore, the paraconformity represents a period of
nondeposition and a condensed section, occurring across a large part of the offshore Gippsland
Basin.

The Conger Formation spans from latest Early to Middle Miocene (Figure 2.2); c. 16.5–14 Ma
(Partridge, 1999). The dominant lithologies are muddy limestone and marly wackestone that contain
up to 30% bioclasts (Partridge, 1999). The palaeodepositional environments vary from coastal plain,
through shallow carbonate shelf, upper slope turbiditic and out to hemipelagic marine. The Conger
Formation is equivalent to the seismic sequence TL2 (Duff et al., 1991), petrographic core-based
Unit II (Bernecker et al., 1997), seismic sequences 4–5 and part of sequence 3 (Feary and Loutit,
1998) and upper Angler Subgroup–zonules D and J (Holdgate et al., 2000).

The Cod Formation spans from Middle Miocene to latest Middle Miocene (Figure 2.2); c. 14–10.5 Ma
(Partridge, 1999). Lithologies include wackestone and packstone that contain abundant bioclasts
(Partridge, 1999). Palaeodepositional environments are the same as for the Conger Formation. The
Cod Formation is equivalent to seismic sequences TG1–TG3 (Duff et al., 1991), petrographic core-
based Unit III (Bernecker et al., 1997), seismic sequences 6–11 (Feary and Loutit, 1998) and the

The Bullseye Karst is near-coincident with the transition from Middle to Late Miocene (Figure 2.2); c.
10.5 Ma (Partridge, 1999). The Bullseye Karst was initially interpreted from the Mackerel 3-D seismic
survey where circular features were seen on seismic TWT slices ranging in TWT from 820 to 868 ms
(Sanders and Steel, 1982). The Bullseye Karst coincides with the Cod to Whiting Formation
transition (Partridge, 1999), upper Albacore Subgroup (Holdgate et al., 2004), upper Unit III
(Bernecker et al., 1997) and base of sequence SE-11 (Feary and Loutit, 1998).
The Whiting Formation spans from earliest Late Miocene (c. 10.5 Ma) to Pleistocene (Figure 2.2, Partridge, 1999). Lithologies include bioclastic wackestone and packstone (Partridge, 1999). Interpreted palaeodepositional environments vary from coastal plain, shallow carbonate shelf, upper slope turbiditic, to hemipelagic marine. The Whiting Formation is equivalent to seismic sequences TG4–TG5 (Duff et al., 1991), petrographic core-based Unit IV (Bernecker et al., 1997), seismic sequences 12–16 (Feary and Loutit, 1998) and Hapuku Subgroup–zonule A (Holdgate et al., 2000).

The Lakes Entrance Formation contains mudstones with less significant sandy and glauconitic intervals. The name dates from the 1930s, when it was initially described from outcrop (Partridge, 1999). The Lakes Entrance Formation is roughly the proximal equivalent to the basal Swordfish Formation of Partridge (1999, Figure 2.2). The term Gippsland Limestone dates from the 1960s when the Limestone was described from outcrop (Partridge, 1999); it is the proximal equivalent to the upper part of the Angler Subgroup (Holdgate and Gallagher, 1997).

**2.3 Volcanostratigraphy**

Extrusive and intrusive volcanics have been intersected in various locations across the Gippsland Basin (Figure 2.2). Age-dating of volcanics helps to associate and/or ascribe drivers to geological events. Late Cretaceous to Pliocene volcanics exist on both the onshore and offshore regions.

No early Cretaceous (pre-Otway Unconformity) volcanics have been recorded in onshore Victoria, but Late Cretaceous volcanogenic detritus (c. 123, 103 Ma) has been described from the Otway Ranges (Duddy, 2003). Volcaniclastic sediments of the offshore Strzelecki Group are interpreted to originate from andesitic volcanic arcs, located to the east in the Tasman Basin (Norvick et al., 2001; Norvick, 2005). Longtom-2 intersected a volcanic unit within the Emperor Subgroup, termed the Turonian volcanics (Lanigan et al., 2007). Tholeiitic rocks spanning the Cenomanian to Santonian (c. 94.7–87.5 to c. 90.6–85.2 Ma) are described from onshore Poowong, Krowera and Latrobe Valley locations. These volcanics span the period of deposition of the Emperor Subgroup and are probably associated with breakup of the eastern margin. Santonian to Campanian volcanics are commonly intersected within the Golden Beach Subgroup (Figure 2.2) and, often intersected at the top of the Chimaera Formation/Seahorse Unconformity (Bernecker and Partridge, 2005). These volcanics are interpreted to form a top seal to the high CO₂-rich gas accumulation in the Kipper Field (Sloan et al., 1992; O’Halloran and Johnstone, 2001). The volcanics are described as basaltic lavas with chloritic alteration; tuffs are also present. Onshore, olivine basalts were sampled at two levels above the Seahorse Unconformity in the Latrobe Valley and Ballen Graben (Duddy, 2003).
The early Palaeocene Currajung Volcanics are present within the offshore upper Halibut Subgroup (Partridge, 1999; Figure 2.2); these volcanics have not been dated using isotopes but are constrained by the P2-to-P5 foraminifera zones (c. 61–55 Ma). Onshore, the Currajung Volcanics originate from the Currajung Monocline where they were dated at c. 58–57 Ma (Bernecker and Partridge, 2001; Power, 2003). Other volcanics, alkali and tholeiitic basalts, have been described from the onshore central Latrobe Valley and Rosedale areas, respectively (Duddy, 2003)—these locations trend with the offshore Northern Strzelecki Terrace (Price et al., 2003).

Onshore Early Eocene to recent volcanics are common and volcanic provinces are recognised, (Price et al., 2003). Volcanism occurred throughout the Cainozoic, although venting intensity varied. Two periods of intense volcanism occurred at c. 57–42 Ma and from c. 5 Ma to present; no volcanism has been recognised between c. 17–8 Ma. Offshore, the Thorpdale Volcanics have also been encountered in the Cobia Subgroup and, Bream volcanics in the Barracouta Formation (Figure 2.2; Bernecker and Partridge, 2001; Power et al., 2003; Bernecker and Partridge, 2005).

No post-Eocene volcanics have been recorded offshore (Figure 2.2). In comparison, onshore volcanics have been intersected near the equivalent base of the Seaspray Group in the Stradbroke, Alberton and Gelliondale Fields as well as in the Baragwanath Anticline (Barton et al., 1992; Chiupka, 1996). Onshore volcanics of equivalent age (c. 30–18 Ma) are found within the South Coast province (Price et al., 2003). Volcanics of Early Oligocene age (c. 36–25 Ma) have been found in the Latrobe Valley (Holdgate and Gallagher, 1997). Late Oligocene to Early Miocene volcanics have been recognised in the onshore Moe Swamp Basin (Barton et al., 1992; Chiupka, 1996). Duddy (2003) dated onshore basalts ranging from 26.1–20.7 Ma, although Bolger (1991) refers to these basalts as Late Oligocene to Early Miocene. Late Miocene to Pliocene volcanics (c. 7 Ma to present) are located onshore within the Western District and Macedon–Trentham provinces (Price et al., 2003)—these are commonly referred to as the Newer Volcanics (Wallace et al., 2005). A major peak occurred at 2 Ma, with a secondary peak at 5 Ma (Cupper et al., 2003); over 250 eruption centres exist in central and southwestern Victoria.

2.4 Fault systems

Lake Wellington Fault System

The Lake Wellington Fault System bounds the northern edge of the Northern Strzelecki Terrace and trends E–W for approximately 200 km (Figure 2.1). The average fault plane dips southwards at approximately 42° and the throw across it ranges up to 3.3 s TWT (Power, 2003). The Lake
Wellington Fault System was initiated via an extensional regime in the Early Cretaceous and later reactivated via a compressional regime in the Late Miocene and Pliocene.

**Rosedale Fault System**

The Rosedale Fault System (RFS) bounds the southern edge of the Northern Strzelecki Terrace and trends E–W for approximately 200 km (Figure 2.1). The fault plane dips southwards at approximately 50° and the throw ranges from 0.2 to 0.7 s TWT (Power, 2003; Power et al., 2003). The RFS was initiated in the Turonian (Figure 2.2) and may have been intermittently active during deposition, as implied by throw variability along strike, and at different stratigraphic levels. The RFS is the northern boundary of the Gippsland Basin’s Cretaceous depocentre. Multiple splays are described as branching into the RFS and connecting to thicker synsedimentary section downdip (Power, 2003; Power et al., 2003). *En echelon* fault and fold geometries are present at different stratigraphic levels, in part reflecting intermittent folding events during the Coniacian–Santonian, Oligocene and Miocene (Figure 2.2).

**Kingfish–Tuna Transition Zone**

The Kingfish–Tuna Transition Zone (KTTZ) is a deep-seated strike-slip fault trending NNE–SSW (Power, 2003; Power et al., 2003); the KTTZ has also been referred to as the Judith Transfer (Lowry and Longley, 1991). The KTTZ transects the northern part of the basin (Figure 2.1) and merges with a reverse fault linked to both the Angelfish and Batfish structures. The zone’s presence was initially interpreted from changes in dip of the top of the Otway Unconformity surface either side of the fault trace; however, fault offsets under the Northern Strzelecki Terrace were not identified. Conjecture exists with regard to the existence of the KTTZ, as interpretations of transfer faults across Australian basins in the late 1980s were based on inferences from conceptual rift models (Etheridge et al., 1985; Etheridge et al., 1988), often irrespective of whether fault planes were visible on seismic data. Further research is required to establish any relationship of the KTTZ to an underlying transfer.

**Darriman and Foster Fault Systems**

The Darriman and Foster Fault Systems trend WNW–ESE and bound the northern and southern edge of the Southern Strzelecki Terrace, respectively (Figure 2.1). Fault planes of the Foster Fault System dip to the NNE from 15 to 40°; the fault is potentially listric in nature (Power, 2003). In comparison, the fault plane of the Darriman Fault System dips to the NNE at 45°, this fault is interpreted by Power (2003) to be planar. The throw variation along strike and the Early Cretaceous section present against the footwall indicate that the Darriman Fault System was initiated after the Otway Unconformity and, may have been coevally active during deposition of the Emperor Subgroup.
sediments in the Turonian (Figure 2.2). Both the Darriman and Foster Fault Systems were reactivated in the late Cretaceous and again in the Cainozoic.

**Salmon, Red, Green-East, Green-West, Blue-West and Blue-East faults**

Smaller-scale faults (i.e. Salmon, Red, Green-East, Green-West, Blue-West, Blue-East) have been identified in the northern Gippsland Basin (Power, 2003; Power et al., 2003). These faults trend WNW–ESE and have throws up to 50–450 ms TWT, with most fault planes dipping from 50 to 60º to the SSW. This WNW–ESE fault trend is also the dominant fault trend in the late Cretaceous to Eocene depocentre and, associated faults form the major focus of this study (syn-sedimentary faults in Figure 2.1, discussed in Chapter 3).

**Seaspray Group faults**

Evidence of Tertiary compression had been noted in the Gippsland Basin (Young et al., 1991; Davidson, 1995), although these studies did not provide accurate dating. The reverse faults and folds were later reinterpreted and ascribed to the post-Oligocene Seaspray Group, where four late Tertiary compression events reactivated many of the structures located over the Northern Strzelecki Terrace (Figure 2.5; Johnstone et al., 2001; Power, 2003). Evidence of compression, in what is a margin sag regime, is difficult to account for (Norvick, 2005), although the event is generally attributed to the Southeast Asia–Australia plate collision (Johnstone et al., 2001; Power et al., 2003). Slumps as well as fault sets, possibly polygonal in style, have been interpreted in localised areas within the Seaspray Group (Feary and Loutit, 1998; Holdgate et al., 2000).

### 2.5 Basin phases

The review of the basin phases has been simplified by describing them as two phases. The basin tectonic stages (i.e. pre-rift, syn-rift, rift-drift, postrift, compression) are described in this section, with consideration given to the basin cycle (i.e. accommodation through thermal or tectonic subsidence and base level changes through eustasy or erosion uplift).

**Basin syn-rift phase 1**

An Early Cretaceous (c. 145.5–125 Ma) syn-rift overlies Palaeozoic basement; also, limited evidence suggests that an older syn-rift phase (c. 152–145.5 Ma) may also be present (Norvick et al., 2001; Norvick, 2005). Extension linked to the former syn-rift section has been estimated at 16% using palinspastic reconstructions (Power et al., 2001; Power, 2003; Power et al., 2003). Major depocentres, with up to 3.5 s TWT of fill, are located in the Central Deep and immediately north of the Southern Strzelecki Terrace. Secondary depocentres, with up to 2 s TWT of basin fill, are located in the Eastern Graben. A thin syn-rift section is present across localised highs and under the
Northern Strzelecki Terrace. The megasequence associated with this syn-rift phase naturally thins at the basin edges. This megasequence is stratigraphically equivalent to the lower Strzelecki Group and is top-bounded by a mid Cretaceous Unconformity. A TWT thickness map of the syn-rift section is unavailable; a map of the combined syn-rift and postrift-phase-1 section is used here to illustrate thickness trends (Figure 2.4a).

**Basin postrift phase 1**

A megasequence representative of an Early Cretaceous (c. 125–96 Ma) transitional-to-postrift basin phase, also referred to solely as a postrift phase (Norvick et al., 2001), overlies the earlier syn-rift phase (c. 145.5–125 Ma). This megasequence has been subdivided into three sequences strz2-a, strz2-b, strz2-c (Power et al., 2001); palinspastic reconstructions estimate extension at 6, 5 and 4%, respectively. This megasequence is stratigraphically equivalent to the upper Strzelecki Group and is top-bounded by the Otway Unconformity (Figure 2.2).

Extensional structures are interpreted in the eastern part of the Central Deep, as well as in the eastern and southern grabens (Figure 2.1). Crustal extension was estimated at 30% (Power et al.,
Faults within the postrift-phase-1 section were initially interpreted as trending WNW–ESE (Power et al., 2001; Power, 2003), but re-interpreted as trending NE–SW as a result of considering the merged syn and postrift sections (Power et al., 2001; Power, 2003). Grabens located adjacent to the Foster, Darriman and Rosedale Faults deepened during this phase (Power, 2003); as well, the Central Deep accumulated up to 3.5 s TWT of fill. Secondary depocentres, to the southeast in the Southern and Northern Grabens, accumulated up to 2 s TWT of fill. Thinning occurs across localised structures and to the north in the Eastern Graben, as well as along the basin edges. The Foster and Darriman Fault Systems that bound the Southern Strzelecki Terrace were active during this period and controlled the extent of basin fill to the south. The RFS bounds the northern part of the basin and was also active during this postrift phase.

Faults arresting at the Otway Unconformity help establish its origin as tectonic. Both listric and planar faults bound rotated fault blocks; the dip of the unconformity surface varies from 2° to 38° (Lowry and Longley, 1991). The interpretation of the unconformity surface is more problematic at depth and south of the Northern Strzelecki Terrace, as it has not been intersected in any wells. Fault systems associated with the breakup trend E–W (Power, 2003), similar to the infra-rift basin system interpreted for the Upper Jurassic (Norvick et al., 2001; Norvick, 2005). The large-scale Rosedale, Darriman and Foster Fault Systems bound the basin depocentre (Figure 2.1). The Strzelecki Group’s thickness ranges up to 5.3 s TWT (Figure 2.4a).

The Otway Unconformity has long been associated with breakup along the Southern Margins; originally interpreted to commence at c. 95 Ma based on magnetic anomaly 33o (Veevers, 1986). However, the age of breakup was later revised to c. 83 Ma following a reinterpretation of magnetic anomalies and lineations in the Great Australian Bight (Tikku and Cande, 1999) and a revised structural interpretation (Sayers et al., 2001). It follows that the entire postrift-phase-1 section predates initiation of seafloor spreading at c. 83 Ma in the Great Australian Bight, but also in the Tasman Basin, c. 84 Ma (Gaina et al., 1998). Therefore, the Otway Unconformity (c. 98–92 Ma) in the Gippsland Basin may actually be related to pre-breakup events on the Southern Margins in addition to being linked to the formation of island arcs that were located east of the present-day Lord Howe Rise (Norvick et al., 2001; Norvick, 2005). Onshore, the age of inversion and uplift is based on apatite fission track analysis (Duddy and Green, 1992). The AFTA data show that the Otway Unconformity correlates with commencement of hinterland denudation in onshore Victoria; c. 95 Ma (Weber et al., 2004). Uplift and denudation at c. 95 Ma has also been corroborated by an interpretation of the regional geology (Norvick et al., 2001).

**Basin syn-rift phase 2**
The Emperor Subgroup has been interpreted as forming part of a second basin syn-rift phase (Figure 2.2; Norvick et al., 2001; Power et al., 2001). Basin extension during basin syn-rift phase 2 has been estimated at 5% using palinspastic reconstructions (Power et al., 2001). The fault orientation changes from NE–SW in the preceding phase to E–W by the end of the second syn-rift phase (Power, 2003). Depocentres in the Central Deep and Northern Graben accumulated up to 1.8 s TWT of basin fill (Figure 2.4b). Thinning of the Emperor Subgroup occurs over the Tuna Field structure. A period of inversion terminated this second syn-rift phase (Power, 2003). Major areas that underwent inversion at this time were the SE Strzelecki Graben and, the Northern and Southern Strzelecki Terraces—the proto-Rosedale North fault that flanks the Northern Strzelecki Terrace, was initiated at this time.

The hiatus, c. 89–86 Ma, associated with the Longtom Unconformity (Figure 2.2) is correlated with pre-breakup tectonics in the Tasman Sea (Norvick, 2005) and predates the breakup of the Southern Margins estimated at c. 83 Ma, as estimated by Sayers et al. (2001). Complex, pre-breakup tectonics are to be expected at a quadruple plate junction, which was present at this time and incorporated the Campbell Plateau, the Lord-Howe, and the Otway and Ross-Sea branches of the junction (Norvick, 2005). The Longtom Unconformity defines the top of the second syn-rift phase (c. 92–86 Ma). At breakup, the maximum shear stress azimuth has been estimated at WSW–ENE with extensional fault systems orientated approximately E–W (Power, 2003, Figure 2.4b); the fault set trends at approximately 90° to the rift margin.

**Basin transition-rift phase 2**

The Golden Beach Subgroup was interpreted as a basin rift phase (Power et al., 2001; Power, 2003), but is also interpreted as a transition-rift phase accompanied by inversion (Norvick et al., 2001). This transition-rift phase is Late Coniacian to Late Campanian (c. 86–73 Ma) and represented here as part of basin phase 2 (Figure 2.2). The predominant fault trend changed from E–W to NW–SE by the end of this transition-rift phase (Figure 2.4c). The main depocentre was the Central Deep, underlying the Mackerel Field; secondary depocentres exist in the eastern Gippsland Basin. Thinning of the Golden Beach Subgroup megasequence is observed across some of the major fields as well as along the basin edges. Another area that underwent inversion occurs ENE of the Tuna Field and parallels the proto-Rosedale North Fault; the latter fault was abandoned at the end of the transition-rift phase but coevally with the initiation of the proto-Rosedale South Fault (Power, 2003; Figure 2.4c).

The basin transition-rift phase-2 section is top-bounded by the Seahorse Unconformity (Figure 2.2), which is difficult to identify in the basin depocentre but more easily recognised near basement hinge
areas. This unconformity coincides approximately with a change from slow to fast seafloor spreading in the Tasman Sea at c. 72 Ma (Gaina et al., 1998; Norvick, 2005). Onshore, a change from fast-to-slow hinterland denudation, is interpreted based on AFTA data (Weber et al., 2004).

### Basin postrift phase 2

Basin tectonic cycles that postdate the transition-rift phase 2 (c. 73–0 Ma) have previously been interpreted or referred to in different ways, as follows.

(i) Fault-controlled basin subsidence and/or sag related tectonism (c. 73–52 Ma) followed by flexural subsidence and/or sag (c. 52 Ma to present) has been suggested (Power, 2003); in this case, the Mackerel Unconformity forms the transition boundary. Depocentres, linked to fault-controlled basin subsidence, are located in the Eastern Graben and Central Deep with thickness ranging up to 1.3 and 1 s TWT, respectively (Figure 2.4d).

(ii) Two postrift transition phases (c. 73–36 Ma and 36 Ma to present) have also been interpreted (Norvick et al., 2001); in this case, the Latrobe Unconformity forms the transition boundary.

The accumulated fill representative of the basin-postrift-phase-2 section ranges up to 2.2 s TWT with its depocentre located in the Central Deep. The NW–SE-trending faults that dominate this phase are a major focus in this study (see Chapter 3). Periods of inversion are noted in the Early Eocene, Late Oligocene to earliest Miocene, Middle to Late Miocene and, from the Pliocene to Recent (Power, 2003).

The Mackerel Unconformity correlates, to within 1–3 Myr, with cessation of seafloor spreading in the Tasman Sea during the Early Eocene, shortly after c. 53.3 Ma (Gaina et al., 1998). Both the Mackerel and slightly younger Marlin Unconformities coincide with an Early Eocene compression event observed in the northern part of the Gippsland Basin (Johnstone et al., 2001; Power, 2003). In comparison, the Latrobe Unconformity coincides in time with drift of the Antarctic Plate; a compression event is also demonstrated in the Early Oligocene to Early Miocene, c. 35–22 Ma (Johnstone et al., 2001; Power, 2003). The Latrobe Unconformity also coincides with the formation of oceanic crust between Tasmania and the Tasman Rise in the early Oligocene, c. 36–32 Ma (Norvick, 2005). Although the Antarctic continent did not clear Tasmania until the Late Oligocene, both the Tasman and Drake passages open in the Early Oligocene (c. 30 Ma) allowing commencement of the circum-Antarctic polar current. Evidence for the opening of the Tasman Passage comes from a marked change in lithology recorded in five deep sea drilling program wells located in waters south of Tasmania (Exon et al., 1999), where the lithology abruptly changes from an organic- and glauconite-bearing, clayey siltstone–silty claystone to a foraminifer–nannofossil bearing ooze–chalk.
2.6 Petroleum prospectivity

A significant number of large oil, condensate and gas accumulations in the Gippsland Basin have been producing since the late 1960s (Brown, 1985; ESSO Australia, 1988). The majority of exploration wells targeted the upper Latrobe Group interval, from which most production has come (Brown, 1985). The 1998 cumulative production figures for oil, condensate and LPG/gas are 4.74 Bbbl (547.93 GL), 0.61 Bbbl (71.04 GL) and 4,837 Gft³ (136.96 Gm³), respectively (Malek and Mehin, 1998). Remaining recoverable reserves in 1998 were estimated at 0.86 Bbbl oil (99.67 GL), 0.3 Bbbl condensate (34.26 GL) and 4,780 Gft³ LPG/gas (135.34 Gm³). Oil- and gas-in-place recoverable reserves for undeveloped fields were estimated at 0.55 Bbbl (64 GL) and 4,513 Gft³ (127.8 Gm³), respectively. Undiscovered recoverable oil and gas reserves are estimated at 0.26 Bbbl (30 GL) and 706 Gft³ (20 Gm³), respectively.

Within the Latrobe Group, major oil producing reservoirs are present within the Mackerel, Kingfish and Barracouta Formations, while combined oil and gas producing reservoirs are present within the Flounder and Volador Formations and, the Chimaera Formation reservoirs are gas producing. Hydrocarbon indications have been recorded at other stratigraphic levels within the Latrobe Group (Admiral, Longtom, Barracouta, Burong, Gurnard, Korumburra and Turrum Formations); and there are some hydrocarbon indications present in the overlying Seaspray Group. Petroleum shows were compiled from Geoscience Australia’s online petroleum database (www.ga.gov.au—data and applications, energy, applications, petroleum wells applications). The majority of successful petroleum discoveries are in the Halibut Subgroup, with secondary discoveries in the Emperor, Golden Beach and Cobia Subgroups (Figure 2.2).

Occurrences of CO₂ are present throughout the basin in association with petroleum accumulations, many located along and adjacent to the Northern Strzelecki Terrace (Figure 2.6). Stratigraphically, CO₂ is present in the Admiral, Chimaera, Volador, Kingfish, Mackerel, Flounder and Burong Formations and, has been recovered from sands of the Tuna-Flounder Channel fill (i.e. Pers Comm. G. O’Brien – Jul. 2006).

The Department of Primary Industries Victoria—Minerals and Petroleum Division (DPIVIC) routinely promotes exploration through regular (usually annual) acreage release (Chiupka, 1996; Thomas et al., 2003). A brief summary of the reservoir–seal intervals, petroleum source rocks, generation and migration history, and play concepts are given below and supplemented by the petroleum systems chart (Figure 2.2). Petroleum systems will not be elaborated on as these are only peripheral to this study; a more comprehensive summary is available elsewhere (Bernecker et al., 2003).
Reservoir and seal intervals: Numerous petroleum-bearing reservoir intervals exist; these are summarised in Bernecker et al. (2003). Ranked from most-to-least prolific, these include the Halibut, Golden Beach, Cobia and Emperor Subgroups. The Kate Shale and Anemone Formations are distal marine shale units; they act as local and subregional seals (Figure 2.2). Other local/intraformational seals are present within the Cobia Subgroup (Gurnard and Turrum Formations), although sands within the formations may act as thief zones (Gibson-Poole et al., 2008b). Intraformational seals also exist throughout the Latrobe Group; again, the seals are more prevalent in the distal setting. The Seaspray Group is dominantly carbonate, relatively unfaulted within the depocentre, approximately 1–2 km thick; and acts as the regional seal. Nevertheless, some potential for petroleum entrapment is thought to exist (Power, 2003), though it has yet to be demonstrated. The Swordfish Formation forms the basal part of the Seaspray Group and is thin to absent in places.

Source rocks, generation, migration: Coals are the dominant source rock for oil generation in the basin (Moore et al., 1992), although marine source rocks have also been found (Gorter, 2001; Volk et al., 2001). Two periods of peak generation are postulated in the early Eocene and Late Oligocene (Figure 2.2; Thomas et al., 2003). Long-distance secondary hydrocarbon migration has been shown to have occurred both by modelling and by the presence of fields located on the Northern Strzelecki Terrace and Platform (O’Brien et al., 2008).
Traps: Traps are predominantly structural but include both faulted and unfaulted subcrop traps (Ryan and Vinson, 1997; Gross et al., 2008), combined structural–stratigraphic traps (Davies, 1983; Brown, 1985), combined volcanic top seal–fault traps (Sloan et al., 1992) and stratigraphic traps that overlie folds (Power, 2003; Power et al., 2003). ESSO Australia Ltd have used seismic attributes to delineate stratigraphic traps (Hinton et al., 1994), particularly, combined stratigraphic–erosional edges (Cousins, 1995; Djakic, 1996). However, the delineation and understanding of how faults contribute to the trap is currently becoming more prominent (Gross et al., 2008). Anticlinal traps exist but are mostly located in the southern part of the basin (Thomas et al., 2003). Of the aforementioned, structural traps with a subcrop component within the upper Latrobe Group have been the main drilling targets.

Play concepts: Offshore plays within the Strzelecki Group have been recognised for some time (Carmody, 1992) and similar plays have proven successful onshore (Holdgate and McNicol, 1992). However, the prospectivity of the Strzelecki Group is deemed high risk in the offshore, due to poor reservoir quality resulting from deep burial. However, the prospectivity may be enhanced under the Northern Strzelecki Terrace. There, quartz-rich sands, if present in the Lower Strzelecki Group as on the onshore (Constantine, 2001), could be charged from mature source rocks present in the Latrobe Group; these sands could be sealed by the volcanics of the Upper Strzelecki Group. Also, hydrocarbon migration pathways trend northwards from the source kitchens of the Central Deep located to the south (Bernecker et al., 2003).

A number of play concepts in the Emperor Subgroup have also been recognised for some time (Bernecker et al., 2002; Power, 2003). Under the Northern Strzelecki Terrace, a fault trap located in the Longtom Field and sealed by a volcanic layer was successful (Lanigan et al., 2007). Similar plays are a viable target, under or adjacent to this terrace, where the reservoir is buried at depths of investigation of less than 3,500 m. Irrespective of age or depth of burial, this play concept is becoming a focus for exploration in the eastern Gippsland Basin (Weber et al., 2004).

A number of play concepts have been tested in the Golden Beach Subgroup. A fault trap with a volcanic top seal similar to Longtom was successfully tested adjacent to the Northern Strzelecki Terrace, resulting in the discovery of the Kipper Field (Sloan et al., 1992). Interconnecting fault systems within the Latrobe Group and the presence of a top volcanic seal combined to make the discovery a successful play (O'Halloran and Johnstone, 2001).

The Latrobe Group subcropping beneath the Seaspray Group has been the dominant exploration target since the mid-1960s (ESSO Australia, 1988), although recent exploration strategy has shifted towards developing intra-Latrobe plays, particularly in the eastern part of the basin (Carmody, 1992;
Evans, 1992). Deep-water plays have also been identified (Bernecker et al., 2001) and are being promoted through acreage releases (Thomas et al., 2003).

Figure 2.7. Generic cross-section showing petroleum-bearing traps—offshore Gippsland Basin. See Figure 2.1 for field locations. Redrafted after Thomas et al. (2003).

2.7 Discussion

Scope exists to develop a consistent and detailed chronostratigraphic scheme that would incorporate both basin- and local-scale sequence boundaries/unconformities. A chart based on well control and widely spaced seismic lines had been initiated by Norvick et al. (2001) and Norvick (2005), although regional in nature. Lithostratigraphic nomenclature and palynology-based tops were independently developed by Partridge (1999) but these lacked significant seismic input. A seismic sequence stratigraphic interpretation would enable ties to be made to regional as well as local unconformities; thus, incorporating a greater component of structural knowledge and potentially resulting in alternative and enhanced interpretations.

Partridge (1999) revised and updated the palynology of the basal Latrobe Group, defining two new subgroups (Emperor and Golden Beach) and their associated terminating unconformities (Longtom and Seahorse). The basin’s tectonic phases were subsequently defined in a tectonostratigraphic framework and classified into basin syn-rift and postrift phase 1 and, basin syn-rift, transition rift-drift.
and postrift phase 2 (Norvick et al., 2001; Power, 2003; Power et al., 2003). However, the level of structural and tectonic detail within basin phase 2 is poor, as demonstrated below.

(i) Only one map exists to represent the transition rift-drift interval (Campanian–Early Eocene)—the interpreted TWT thickness map of the combined Campanian to Maastrichtian section (Figure 2.4d). A TWT thickness map of the lower Palaeocene to Early Eocene interval was not generated, but it is this interval that straddles the main reservoir interval of interest in this study.

(ii) Only one map exists to represent the postrift interval (Early Eocene to recent)—the interpreted TWT thickness map of the post-Oligocene section (Figure 2.5). A TWT thickness map of the Early Eocene to post-Oligocene section was not generated, but it is this interval that straddles the immediate overlying seal interval to the main reservoir interval of interest in this study. Detailed resolution of basin evolution is also impeded by the merging of a number of unconformities (the Mackerel, Marlin and Latrobe) where a condensed section is developed.

Thus, fault chronology in the Eocene to Oligocene interval is poorly constrained. Additionally, different fault modes (normal, reverse) and regimes (tension, compression) appear to be coevally active at this time. For example, latest Early Eocene normal faulting is dominant in the upper Halibut Subgroup, but coeval with localised reverse faulting (Power, 2003; Power et al., 2003). Although this period of faulting coincides with cessation of seafloor spreading in the Tasman Basin, the normal faults trend perpendicular to the Eastern Rift Margin (the latter estimated from magnetic lineations of Gaina et al., 1998), indicating that the tectonic driver for the faults may not be rifting on that margin. Determination of the origin of this fault trend is complicated by the fact that the Gippsland Basin is located between two active rift margins (the Southern and Eastern Rift Margins). Consequently, seafloor spreading associated with the Southern Rift Margin may have had a greater impact on the Early Eocene fault system following the shift from slow to fast seafloor spreading in the Maastrichtian, estimated at c. 72 Ma on the Eastern Rift Margin (Gaina et al., 1998).

An understanding of petroleum systems and geology at the regional scale helps highgrade which part of the basin is best suited for undertaking this study. For example, the major petroleum-bearing reservoir intervals are present within a Coniacian to Eocene clastic wedge, forming the bulk of the Latrobe Group; the regional seals of a carbonate-dominated sag margin wedge (the Seaspray Group) overlie the Coniacian to Eocene clastic wedge. Petroleum has migrated from the kitchen areas in the east and south, to traps in the west and north. Long-distance migration has been inferred by modelling and discoveries present onshore and on the Northern Strzelecki Terrace (O’Brien et al., 2008), all quite distant from the source kitchens. Long-distance migration is more prevalent in a basin’s proximal setting where the sand-to-shale ratio increases. More localised
sourcing and migration may occur in a distal setting (i.e. direction of the rift margin) where the intra-Latrobe Group shale (source rock) content increases. These features are in line with petroleum migration concepts in clastic wedges (White, 1980). It follows that an improved understanding of fault geometries and sealing across faults is increasingly important for ascertaining migration and/or trap issues in the more shaley part of a basin. Also, the eastern part of the Gippsland Basin appears favourable based on key fault patterns established (Power et al., 2003; Power, 2003), as well as from key palaeogeographic maps (Partridge, 1999; Bernecker and Partridge, 2005) and chronostratigraphic relationships in existence (Bernecker and Partridge, 2001; Bernecker et al., 2001).

2.8 Conclusions

The unconformities and seismic horizons to be interpreted in this study will be tied to and compared to the stratigraphic scheme of Partridge (1999) and Bernecker and Partridge (2001). Although the stratigraphic scheme of Holdgate et al. (2000) is referred to in the literature (Norvick et al., 2001; Power et al., 2003), it has not been adopted here, as formation tops in WCRs are not based on this stratigraphic scheme. The seismic-based sequence stratigraphic scheme of Feary and Loutit (1999) is not adopted, for the same reason given for Holdgate et al.’s (2000) scheme. However, the seismic interpretation along traverses will be linked to the stratigraphic scheme of Partridge (1999) and Bernecker and Partridge (2001), to facilitate the interpretation of the Seaspray Group where tops are infrequent (see Chapter 3).

The Otway, Longtom, Seahorse, Mackerel, Marlin and Latrobe Unconformities, as well as the Marshall Paraconformity and Bullseye Karst, are all recognised in the basin, withstanding that some terminology conflicts occur. The Cretaceous–Tertiary boundary, together with any unrecognised unconformities within the Halibut Subgroup, will be investigated to establish their relationship to the dominant fault trend of the Early Eocene (see Chapter 3). Further, although well mapped at the regional scale, there is an absence of structural detail within the basal Palaeocene to latest Early Eocene and Early Eocene to Oligocene intervals (c. 66–36 Ma); these intervals are critical, in that they coincide with the upper part of the main hydrocarbon reservoirs (also being considered for CO₂ storage in this study), as well as with the base of the regional seal. Any structural detail within this interval could be critical to understanding CO₂ migration within traps and could have ramifications for the structural assessment of any CO₂ storage sites.
CHAPTER 3—SEISMIC INTERPRETATION

3.1 Introduction

The number of publicly available geological, geophysical and engineering datasets for the Gippsland Basin is considerable. The datasets obtained for this study include well data (basic location, petroleum status, logging suites, well deviation); oil–gas and CO₂ compositional data; geological data (age-depth plots, formation tops); geophysical data (3-D seismic, checkshot, synthetics); and geomechanical data (rock properties, state-of-stress interpretations). The well, geological and geophysical data underpin the seismic-based structural interpretation undertaken in this study (Chapter 3) while the geomechanical data underpins the fault seal analysis (Chapters 6–7).

Partridge (1999) reworked the palynology and stratigraphy and subsequently established new nomenclature and palaeoenvironment maps for both the Latrobe and Seaspray Groups. From this work, a new sequence stratigraphic chart has become the industry standard (Bernecker and Partridge, 2001). Further, the tectonic, eustatic and/or sedimentological drivers for the supersequences were recognised, and associated sequence stratigraphy tied to the regional geology (Norvick et al., 2001). Formally recognised regional unconformities and fault systems are described in Chapter 2. In this chapter, both the regional and local hiatuses and associated unconformities are interpreted to provide a tectonostratigraphic framework, within which to improve the understanding of fault chronology and constrain basin evolution in the eastern Gippsland Basin and study area. Further, the improved fault chronology, particularly within the reservoir interval considered (Halibut Subgroup) and overlying seal intervals (Flounder Formation to Seaspray Group), helps elucidate trap integrity related to potential CO₂ SLs (see Chapter 5).

The regional structure and basin evolution have been interpreted from deep (10 s) reflection seismic data (Power et al., 2001). A 3-D seismic interpretation of the Tuna Field allowed structural detail to be incorporated into the regional interpretation. This regional interpretation was supported by palinspastic reconstructions that together helped establish the rotation of S_{Hmax} from a W–E azimuth in the Turonian, to a NW–SE azimuth in the Early Eocene (Power et al., 2003). The contemporary state of stress was determined using borehole imaging and calliper-based borehole breakout interpretations; S_{Hmax} is estimated at 139° ±1°N (i.e. NW-SE; Nelson, 2007), corroborating the seismic-based estimate. The seismic-based structural interpretation herein constrains the fault geometry, fault throw and fault arrays and, identifies fault zones; the impact on CO₂ SLs is assessed in Chapters 5–7.
Nevertheless, a detailed understanding of the later part of the Palaeocene to Eocene postrift basin phase and sedimentary section is lacking, for three reasons. First, part of the Early Eocene Cobia Subgroup is thin-to-condensed. Second, differing fault modes (normal, reverse and strike-slip) and associated structures coexist during the Early Eocene; these are represented in the upper part of the Halibut Subgroup. The fault chronology for these differing modes and their associated basin regimes is normally ascribed to one event; namely, cessation of seafloor spreading in the Tasman Basin. Third, the upper part of the Early Eocene Halibut Subgroup is partially removed in places because of uplift and erosion. However, the Palaeocene to Eocene postrift basin phase is all-important, as it coincides with the reservoir interval (Halibut Subgroup) considered here for CO₂ storage.

The study area, datasets and methodology are described in connection with the objectives described above. Specifically, the methodology used to interpret and quality-control over 800 formation tops and, hiatus markers from 85 wells is detailed. A new unconformity and seismic horizon nomenclature is introduced. The workflows used to interpret the seismic horizons and faults are discussed. These interpretations rely on both seismic TWT profiles and TWT slices of the seismic variance attribute. The workflows integrate the interpretation of fault traces, segments, contacts and horizon boundaries to establish fault arrays and geometries, differentiate fault families and classify faults according to fault throw.

### 3.2 Study area

The study area is bounded by longitudes 148°9′–148°46′E and latitudes 37°57′–38°36′S (Figure 3.1). The NE, SE, SW and NW corners of the study area are near the following petroleum wells or fields: Leatherjacket-1 (Lj-1), Blackback Field (Bb-1 to -3), Drummer-1 (Du-1) and Baleen Field (Ba-1 to -4), respectively. The northern part of the study area coincides with the Northern Strzelecki Terrace and in the southeast is the Bass Amphitheatre. The southwestern part of the study area forms part of the Gippsland Shelf Province that abuts the Central Deep. The regional Rosedale Fault System (RFS) trends E–W across the northern part of the study area, with terraces present down-dip.

The location of the study area was, in part, based on the extent of the Northern Fields 3-D seismic grid (Figure 3.2; also see Chapter 1). A study subarea was also defined within which to undertake a detailed fault seal analysis (Figure 3.1); it was chosen based on the resulting screening of CO₂ SLs (see Chapter 5). The NE, SE, SW and NW corners of the study subarea are approximately coincident with Admiral-1, Bignose-1, Angelfish-1 and the northern extent of the Tuna Field, respectively.
3.3 Datasets

Basic well and log data

Basic well data include well name, universal well indicator (UWI), latitude and longitude (GDA94 datum), the operating company, spud and release dates, permit, deviation information, petroleum shows and well status (Appendix 1: Table A1.1). The study area is extensively drilled; 58 wells and about 400 boreholes (side-tracks) exist within it, with 37 wells located outside of the 3-D seismic grid. Geological control was required from both sets of wells so that ultimately, the locations of 95 wells were entered into Schlumberger’s seismic interpretation workstation GeoFrame™ (Figure 3.1). An information summary regarding wells and loaded data is both tabulated (Appendix 1: Table A1.2) and displayed in map form (Appendix 1: Figure A1.1). Other basic data, such as kelly bushing (KB), water depth (WD) and total depth of the well (well-TD) can be found under individual tables for each well (Appendix 1: Tables A1.17–1.95). These initial digital data were provided by the DPIVIC. Additional well data were downloaded from the Geoscience Australia (GA) website (www.ga.gov.au—data and applications, energy, applications, petroleum wells applications), if missing from the DPIVIC data and, finally, if unavailable from GA, individual WCRs were used. The latitude and longitude for each well were cross-checked specifically against all of the above sources to identify errors and correct where necessary. However, a number of WCRs within the study area were either not open-file or unavailable (see Appendix 1: Table A1.2, Figure A1.1a).

Wireline logs for 46 of the 58 wells located within the 3-D seismic grid were provided by the Australian School of Petroleum (ASP, Appendix 1: Table A1.2). Logs for another 11 wells were unavailable at the ASP; these were donated by Occam Technology Pty Ltd. A log for one well was unavailable (see Appendix 1: Figure A1.1b for areal distribution of log availability). Logs for wells located outside of the 3-D seismic grid were not used; only age-depth plots were interpreted for these. Digital well-deviation data were obtained from Geoscience Australia. The x-y-z offset data were entered into the GeoFrame™ seismic-workstation data tables: offset data were calculated using the appropriate declination value. Only about 28% of the well deviation data were available for wells within the 3-D seismic grid (Appendix 1: Table A1.2). However, the majority of exploration wells were drilled near-vertical and depth errors caused by well deviation are small. The measured depth (MD) was systematically compared against true vertical depth (TVD) to gauge the potential for any mislocation of well markers.
Figure 3.1. Location map of the study area—wells, bathymetric image and key physiographic features.
Source of the backdrop image—www.dpi.vic.gov.au/minpet/geovic. See Appendix 1 (Table A1.1) for well names.
Geological and fluid data

Stratigraphic tops were sourced from the DPIVIC, Partridge’s (1999) Ph.D. thesis, Geoscience Australia’s petroleum database website (www.ga.gov.au—data and applications, energy, applications, petroleum wells applications) and WCRs. Over 700 stratigraphic tops were entered into a spreadsheet; conflicts between tops from the different sources were noted and ultimately resolved. Hiatus markers were separately defined, using age-depth plots and correlative unconformities inferred. Tops and hiatus markers were then entered into the GeoFrame™ seismic-workstation data tables. Geoscience Australia’s tops originate directly from WCRs, whereas tops from Partridge (1999) and DPIVIC’s database are reinterpretations that are based on palynological data. Accordingly, Partridge’s tops were given precedence; Dr Partridge is the head palynologist for the principal operator, Exxon Mobil Corporation. However, some modifications were occasionally required. Overall, 861 stratigraphic tops and hiatus markers from 79 wells were reinterpreted; these averaged at 11 tops per well for the study area (Appendix 1: Tables A1.17–1.95, Figure A1.1c).

Age-depth plots were available from Geoscience Australia’s petroleum database website (www.ga.gov.au—data and applications, energy, applications, petroleum wells applications) for 39 of the 58 wells that are located within the 3-D seismic grid part of the study area (Appendix 1: Table A1.2; Figure A1.1d). Plots were generated from the original palynology datasets; some accuracy limitations exist. Any revised palynology interpretations, carried out by the operator, are commercial-
in-confidence and thus unavailable. Nevertheless, age-depth plots provide a way to readily identify and/or confirm local and regional hiatuses and understand basin fill history (discussed in detail in Section 3.5). Consequently, age-depth plots are an efficient method for building and cross-checking a comprehensive well-top, hiatus-marker and seismic-horizon database (also discussed in Appendix A1.2).

Lastly, DPIVIC provided a spreadsheet with compositional percentages from over 1,000 petroleum fluid samples, including CO₂ collated across the basin (G. O’Brien—Pers. Comm., May 2006). These data were sought to ascertain the potential for processing CO₂-rich gas and storing it, thereby impinging on the potential relevance of CO₂-SLs identified in this study (see Chapter 5).

**Geophysical data**

The 2002 Northern Fields 3-D seismic dataset, used in this study, has been open-file since 2005 and was obtained through the intermediary of the DPIVIC. The full-offset, pre-stack, time-migrated cube was purchased from Spectrum Data. The 3-D survey navigation, acquisition parameters and seismic processing sequence are described separately (Appendix 2: Section A2.1).

The seismic data were loaded at 8-bit resolution to save hard-disk space, which was limited to 300 Gb with a 300 Gb backup. The data have a native 32-bit resolution that can be beneficial for stratigraphic work, but is not critical for a structurally oriented project. Ideally, a processing flow should preserve true amplitude to maximise resolution (Trappe and Hellmich, 2003; Brown, 2004). However, true amplitude preservation is more pertinent for stratigraphic interpretation than for structural interpretation; true amplitudes were not preserved in the 3-D seismic dataset used here. Nevertheless, comprehensive testing was carried out by the original seismic processing company (Veritas DGC Asia Pacific Ltd.) to remove sources of noise (e.g. water bottom multiples). Removal of coherent noise can have benefits for seismic attribute extraction; for example, time-based horizon attributes rely on the relative position of adjacent time picks, so that the removal of coherent noise improves the signal-to-noise ratio, particularly useful when using dip-based attributes (Neves et al., 2004). Suitable F-K or F-XY filters can also be applied to remove coherent noise, although further processing would be required to ensure amplitude preservation (Brown, 2004). A noisy 3-D seismic dataset can also break up the smooth operation of the autotracker, significantly increasing the time taken to carry out an interpretation, consequently, specialised smoothing filters have been developed to reduce the impact of noise on the autotracker (Fehmers and Hocker, 2003). However, the Northern Fields 3-D seismic dataset was high quality; the original processing flow was aimed at maximising the resolution within the Halibut and Golden Beach Subgroups, which in turn optimises the effectiveness of derived seismic attributes in the key reservoir intervals.
Checkshot data were available for 42 of the 58 wells located within the 3-D seismic grid part of the study area (Appendix 2: Tables A2.4–2.73). Seventy per cent of the publicly available checkshot data were downloaded directly from the Geoscience Australia petroleum-database website (www.ga.gov.au—data and applications, energy, applications, petroleum wells applications), while the remaining 30% were obtained from WCRs. As part of quality control, all checkshot data were graphed to isolate any spurious data as seen from depth offsets. Geograms, synthetics and/or vertical seismic profiles were available from WCRs as PDF files.

### 3.4 Local geology

Only a few lithostratigraphic references that specifically describe the geology of the eastern Gippsland Basin exist (Perrett and Garven, 1998; Johnstone et al., 2001; Fink et al., 2002). Stratigraphic controls, in combination with prominent geological features that are unique to the study area and relevant to the seismic interpretation, are briefly discussed below. The prominent geological features include the Early Eocene Tuna-Flounder and Marlin Channel systems that downcut into the Halibut Subgroup section (Partridge, 1999). Also, Campanian volcanics are prominent within the Golden Beach Subgroup (Figure 2.2). Lastly, inversion occurred in the Early Eocene and a significant section of the upper Halibut Subgroup was uplifted and eroded as a result (Johnstone et al., 2001).

The Strzelecki, Latrobe and Seaspray Groups have been extensively drilled, and stratigraphic control for the seismic interpretation is good. For example, within the study area, a total of eight wells intersected the deep Strzelecki Group, which is considered a high well count relative to other parts of the basin. However, the Strzelecki Group’s depocentre is located southwest of the study area (Power, 2003), so a large part of the Strzelecki Group’s stratigraphy has not been intersected in the study area.

Similarly, more than 10 wells intersected the Emperor Subgroup, primarily encountering the Kipper Shale and, to a lesser extent, the Longtom Formation (see Figure 2.2 for formation names). Wells that have drilled into the Golden Beach Subgroup mainly intersected the Chimaera Formation, but a few encountered the Admiral Formation and one drilled into the Anemone Formation and Kersop Arkose. This latter well, in combination with palaeoenvironmental mapping, established that the Anemone Formation’s depocentre was located seawards of the study area, although alluvial fans of the Kersop Arkose were shown to be located on the basin’s southern terrace and platform (Partridge, 1999). The Kingfish, Mackerel and Volador Formations and Kate Shale (Halibut Subgroup) are well constrained from drilling. In comparison, the Barracouta Formation’s (Halibut Subgroup) depocentre is located westwards of the study area in a more proximal basin setting and, has only been
interacted once in the study area (Partridge, 1999). The Kingfish and Mackerel Formations were uplifted and eroded in the Eocene and simultaneously downcut by the Tuna-Flounder and Marlin Channels where the high velocity of the channel fill causes a seismic pull-up (Gross et al., 2008). Similarly, the high velocity mid-Miocene channel fill, present within the carbonate wedge of the Seaspray Group, also causes pull-up of underlying structure (Holdgate et al., 2000). The Gurnard Formation (Cobia Subgroup) also has been frequently intersected by wells, although this formation is often thin-to-condensed and may be absent or unrecognised. The Burong Formation (Cobia Subgroup) has only been intersected twice, as the depocentre is located westwards, in a more proximal basin setting (Partridge, 1999).

The Seaspray Group (Swordfish, Conger, Cod and Whiting Formations) is well constrained from drilling because the group’s depocentre is located in the southern part of the study area. However, identification of individual formations can be difficult within the relatively homogenous lithologies of this carbonate wedge, and the up-hole section also is often undersampled. In such a case, the sequence stratigraphic interpretation of Feary and Loutit (1998) helped constrain the chronology of seismic horizons interpreted within the group.

Evidence of compression was observed at three stratigraphic levels: first, during deposition of the Emperor and Golden Beach Subgroups in the Coniacian to Santonian; second, during deposition of the Halibut Subgroup in the latest Early Eocene; and third, during deposition of the Seaspray Group from the Oligocene to Pliocene (Johnstone et al., 2001; Power, 2003). The compressional regime active in the Early Eocene resulted in uplift and erosion of the upper Halibut Subgroup in the study area (Flounder Field and to the southeast), whereas, in the western part of the basin uplift or erosion was less pronounced. Both the Coniacian to Santonian and Early Eocene inversion structures are observed within the thicker parts of the eastern Central Deep basin fill, also located within the study area. In comparison, Oligocene to Pliocene inversion structures are present only within the thinner section found on the Northern Strzelecki Terrace, at the northern extent of the study area (Johnstone et al., 2001; Power, 2003).

3.5 Methodology

**Formation tops and hiatus markers**

The basin’s formation tops (in the available digital tops database) do not conform to any one stratigraphic scheme, resulting in correlation conflicts and uncertainties. For example, Feary and Loutit (1998) interpreted wireline logs and applied a sequence stratigraphic scheme to the Seaspray Group, while Holdgate et al. (2000) used an onshore stratigraphic breakdown, which they then
extrapolated to the offshore. Partridge (1999) and Bernecker and Partridge (2001) used a consistent sequence stratigraphic system, but it differs from that used by industry (Johnstone et al., 2001).

A workflow that makes use of age-depth plots (Figure 3.3a) was devised for this study to obtain a large number of consistent quality-controlled formation tops/hiatus markers that are also tied to seismic data (Figure 3.3b). Only 61 age-depth plots were available (Appendix 1: Figures A1.3–1.33); interpreted hiatuses are found in Appendix 1 (Tables A1.3–1.16). The workflow used to define the tops and hiatus markers is briefly described here, and detailed in Appendix 1 (Section A1.2)—the formation tops and hiatus markers of 79 wells are tabulated accordingly (Appendix 1: Tables A1.17–1.95).

Steps 1–5: Formation tops were imported from three datasets whilst hiatus markers were interpreted from the age-depth plots. Depths derived from the three datasets were compared against depths derived from the age-depth plots and any inconsistencies resolved. In total, 861 formation tops and hiatus markers from 79 wells were interpreted and quality controlled.

Steps 6–8: The formation tops and hiatus markers that coincided with the seismically defined horizons were renamed according to the seismic horizon scheme (Figure 3.4). The remaining formation tops and renamed hiatus markers were imported into GeoFrame™'s seismic-workstation tops tables. The hiatus markers were compared against unconformities interpreted from seismic sections to provide a cross-check for the age-depth interpretation.

Several benefits arise from interpreting age-depth plots, as follows.

(i) Nondeposition and/or uplift and erosion can be readily ascertained, as regional hiatuses re-occur and are identified by overlaying and comparing individual age-depth plots. Subregional hiatuses, placed in the ‘probable–possible’ category, can be cross-checked against unconformities on the seismic data to ascertain authenticity and map the extent of the associated boundary.

(ii) A preliminary estimate of an interval’s thickness and depositional rate can be made; however, allowance for compaction needs to be carried out separately. Further, any eroded section (i.e. upper, middle, lower) can be readily assigned to the appropriate stratigraphic interval.

(iii) Closely spaced unconformities can be identified, often where they are potentially merging. Recognition of these potentially merging unconformities on age-depth plots allows criteria to be set up in GeoFrame™'s seismic workstation to aid interpretation and mapping (digitising the base versus the top of the sequence).
(iv) Age-depth plots are readily available from the Geoscience Australia website and interpretation less time consuming than perusing WCRs from 61 wells.

Seismic-horizon and fault nomenclature

A seismic horizon and fault nomenclature was developed in conjunction with the establishment of the Formation top/hiatus marker scheme (Figure 3.4), as explained in what follows.

(i) Any well-established formation top or hiatus marker nomenclature was retained (Figure 3.4, third to fourth columns from left). For example, the well-established Latrobe Unconformity was retained by abbreviating the seismic horizon to ‘LatrSS’; the first four letters of the seismic horizon abbreviation refer to the formation top or hiatus marker name, the last part represents the order of the sequence boundary as interpreted here (MS—megasequence, SS—supersequence, S—sequence). Next, seismic horizons were named according to Partridge’s (Partridge, 1999) stratigraphic scheme, so as to not introduce new nomenclature. Informally recognised hiatus-marker names were added where necessary and prefixed with ‘intra’.

(ii) Two hiatus markers (IFlouS, ICongS) were not tied to or interpreted from the seismic data, as the associated reflector quality was poor. However, these hiatuses are recorded because they can provide evidence upon which to constrain fault chronology. Formation tops that did not coincide with hiatus makers were not interpreted on the seismic section, because no structural information would be gained from the additional interpretation effort.

(iii) Seismic horizons have been interpreted on a negative-to-positive wavelet inflection point; representative of an increase in impedance contrast (Figure 3.4, Seismic polarity column). One exception is the Conger unconformity (CongSS horizon) that coincides with the base of a high-velocity carbonate wedge within the Seaspray Group; this horizon was interpreted on a positive-to-negative inflection point. The second exception is the Mackerel Unconformity (MackMS horizon) that was interpreted on a peak; the seismic polarity and phase conventions used for seismic horizons are elaborated in Appendix 2 (Section A2.2).

(iv) The age of each hiatus (i.e. the Age (Ma) column) is estimated from age-depth plots as interpreted in this study and supplemented by the stratigraphic scheme of Partridge (1999).

Seismic intervals (e.g. UHalSS1) and horizons (e.g. MackMS) will from this point onwards be abbreviated; for example, ‘UHalSS1 interval’ or ‘MackMS horizon’ will be used instead of ‘UHalSS1 seismic interval’ or ‘MackMS seismic horizon’.
Figure 3.3. Interpretation workflow. (a) Age-depth plot for Khehewal showing typical hiatus markers (see Figure 3.4 for hiatus definition). (b) Interpretation flowchart, numbers are workflow steps referred to in text. Reference datasets (well tops): Partridge (1999), Bemercker (Pers. Comm.), geoscience Australia website (www.ga.gov: data and applications, energy, applications, petroleum wells applications). Group IDs are used for minor faults that form clusters and have negligible throw. Fm - formation. (c) Subdivision of the seismic volume to extract variance cubes. Seismic slices of the variance attribute are interpreted at a 16 ms TWI interval above the baseline and, at a 288 ms TWI interval below the baseline. The average baseline varies with location block (e.g. B1, 1,700 ms), and is - 100 ms below the seismic MackMS horizon. (d) A typical section in the study area. See Figure 3.4 for horizon and fault nomenclature. (e) Typical fault traces interpreted using the seismic variance attribute and, fault cuts interpreted using TWT profiles. (f) Typical reflector and fault offsets shown on a seismic TWT slice.
A consistent colour scheme is used throughout this thesis to represent interpreted seismic horizons and faults (Figure 3.4, coloured rows). For example, a fault propagating through the Seahorse Unconformity (SeahSS horizon) but arresting at the lower Halibut unconformity (LHalSS horizon) would be coded as ‘LHal’ and coloured accordingly, whereas a fault arresting at the Seahorse Unconformity would be coded as ‘Seah’ with the appropriate, but different colour applied. The colour-coding aids in the differentiation of different periods of faulting, or to associate reactivated faults (see Chapter 1.

Seismic resolution—application to fault detectability

The terms major and minor faults are used extensively throughout this thesis to represent distinct fault-throw families. A major fault is one where the maximum fault throw is greater than one seismic wavelength ($>1\lambda$, in Figure 3.5a); a number of major faults transect the study area, these are more regional in extent. A minor fault is defined where the maximum fault throw has a corresponding seismic wavelength that ranges from $\lambda/4$ to $\lambda$. Minor faults are not always interpreted in regional assessments; these are also called near-resolution faults (Maerten et al., 2006). A subseismic fault is one where the fault slip has a seismic wavelength of less than $\lambda/4$.

Vertical seismic resolution is either defined by the limit of seismic event separability ($\lambda/4$) or the limit of seismic visibility ($\lambda/8$ to $\lambda/30$), both of which are dependent on the geological conditions as well as the depth of burial (Brown, 2004). The vertical and lateral resolution were initially estimated across the study area using the TWT cube (Appendix 2: Figure A2.12) and later revised using the depth cube. The availability of the seismic cube in the depth domain allowed for direct measurement of the seismic wavelength. The seismic wavelength was calculated at the centre of each location block (i.e. B1–B15 in Figure 3.3c–d) and midway within each of eight seismically defined intervals that mainly form the Latrobe Group (EmpeSS, GBeaSS, LHalSS1, LHalSS2, UHalSS1, UHalSS2, UHalSS3–CobiSS and SworSS). An average seismic wavelength was estimated using an upper and lower bound in cases where the seismic interval was significant (i.e. $\sim 4\lambda$ or greater). A best-fit RMS linear trend line was plotted from the data points represented by the $1\lambda$ seismic wavelength cases (Figure 3.5a). The seismic wavelength at different depths was divided by four to obtain the $\lambda/4$ resolution limit (e.g. $\sim 5$–25 m at depths of 1–5 km).

Horizontal seismic resolution is determined by ascertaining the dominant wavelength, burial depth and average velocity to calculate the Fresnel Zone radius (Figure 3.5b). Horizontal seismic resolution ranges from approximately 200–960 m for depths of 1–5 km, respectively (i.e. based on the Fresnel zone). Further explanation is given in the section ‘Interpretation of a fault zone’ that follows.
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<th>Megasequence</th>
<th>Seismic intervals</th>
<th>Seismic horizon</th>
<th>Formation top, hiatus markers, unconformities</th>
<th>Seismic polarity</th>
<th>Age (Ma)</th>
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<td>Upper Strzelecki Group</td>
<td>125–100¹</td>
<td>LStr</td>
<td></td>
<td></td>
</tr>
<tr>
<td>OtwaSS1</td>
<td></td>
<td>+/-</td>
<td>? Otwa</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>LStrMS</td>
<td></td>
<td>Upper Strzelecki Group</td>
<td>165–125¹</td>
<td>LStr</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Figure 3.4. Seismic sequence/interval, horizon, formation top, hiatus marker and fault nomenclature.

¹Age from Partridge (1999); ²Age estimated from age-depth plots, this study. Fm—formation, ?—uncertain. Nonformally recognised unconformities identified in this study are lower case.
Figure 3.5. Seismic resolution—application to fault detectability, (a) vertical and, (b) horizontal resolution. The seismic wavelength and Fresnel Zone radius are calculated at the centre of location blocks B1–B15 for each of the eight seismic intervals (EmpeSS, GBeaSS, LHaiSS1, LHaiSS2, UHaiSS1, UHaiSS2, UHaiSS3-CobiSS and SworSS). See Figure 3.3c for diagrammatic representation of location blocks.
Nomenclature for fault grouping

A scheme was devised to record and describe fault groupings. In the order of 200 faults were interpreted. These faults were identified by ID number, location and group type, the latter if non-stand-alone faults (Figure 3.3b). Group types include fault arrays, fault systems, fault clusters and fault zones. Faults are grouped into fault arrays where necessary, with the master fault distinguished from the splay(s). Nomenclature definitions follow.

(i) A fault array is defined here as an assemblage of faults that are kinematically linked by a common branch line(s) or intersection line.

(ii) A fault-system is a grouping of fault-arrays that are kinematically linked (e.g. Rosedale Fault System).

(iii) A fault cluster is a high density grouping of faults that are geographically localised, but where establishing any kinematic link is difficult, as the throws are minimal.

(iv) Faults do not always consist of a single clean fracture, so the term fault zone is used herein to refer to the zone of complex deformation associated with a dominant fault plane. Fault zones are described as belonging to two categories (Faerseth, 2006; Faerseth et al., 2007). The first category refers to a fault zone bounded by both external slip surfaces and shale smear along the multiple fault planes within the zone. There is a certain degree of symmetry of the internal fault-plane orientation and dip when compared to the external slip surfaces. Such fault symmetry would be a result of brittle as opposed to ductile deformation. The second category refers to a fault zone bounded by both external and internal slip surfaces and a complex zone that can include slivers of footwall/hanging wall rotated/not rotated and intermixed with fault gouge. Fault zone width is used in this study to refer to the width of the fault zone.

Additional fault parameters interpreted or measured include the fault mode (normal, reverse and strike-slip), upper bound of the fault plane tip (upper tip-line bound), strike and dip, fault trace length, fault-slip family (major, minor) and fault-province family (discussed in Sections 3.6.4–3.6.6). The fault-plane strike was plotted on stereonets at a specified inline interval. The status of the branch lines, whether soft or hard-linked, was not recorded in the study area because of time constraints but was assessed in the study subarea (Chapter 6). A soft branch line is used in this study to refer to the case where space exists near the intersection of two fault planes, possibly present because of plastic deformation (i.e. ductile strain). A hard branch line refers to the case where no space exists, possibly present because of brittle deformation (i.e. nonductile strain). Ultimately, viewing branch lines as
differing through nonductile versus ductile strain (Walsh et al., 1996), with resultant hard versus soft-linked branch lines, is paramount to ascertaining breach in a fault trap.

**Interpretation of a fault trace, segment, contact and boundary**

Seismic slices of the variance attribute and TWT seismic profiles were used to interpret the fault traces, segments, contacts and boundaries (Figure 3.3b), as follows. The seismic coherence attribute has been successfully used since its inception (Bahorich and Farmer, 1995); this attribute is also called seismic variance because of patent restrictions imposed by Conoco Phillips. The seismic variance attribute is calculated from the seismic volume and not dependent on bias that can occur in interpretation, thus complementing seismic-profile interpretation. Workflow steps 9–14 (outlined below) refer to the seismic interpretation; workflow steps 1–8 pertain to the interpretation of formation tops and, were previously discussed (see Section 3.5).

**Step 9:** Seismic slices of the variance attribute were generated from the SEGY volume at 4 ms intervals extending down to 5.5 s, the latter bounding the base of meaningful seismic reflectors. Horizon and fault segments were interpreted at 1.25 km intervals on both inlines and crosslines with, spot checks carried out on TWT slices. Fault segments were given a preliminary ID. This first-pass estimation of fault segments along seismic profiles was subsequently cross-checked against the fault traces as interpreted on seismic slices of the variance attribute.

**Step 10:** The seismic variance attribute was extracted during preliminary runs of testing in order to improve the resolution of the fault trace: the variance operator length, type and number of traces to be averaged were varied accordingly (expanded on in Appendix 2—Section A2.2). Generation of the seismic variance attribute requires input of the operator length ‘L’, operator type ‘+’ or ‘x’ and number of traces to be used in the summation (either 3, 5, 7 or 9). As a general rule, longer operator lengths and/or an increased number of traces will increase the signal-to-noise ratio. In such cases (high L, high number of traces), major discontinuities are better resolved, but at the expense of the minor ones. The ‘+’ operator may give better resolution over the ‘x’ operator in the case where fault trends are parallel to the inline and/or crossline direction. The ‘x’ operator provides better resolution for multiazimuthal fault trends. The seismic variance attribute is represented by Equation 3.1,

\[
\sigma_j^2 = \frac{\sum_{j=t-L/2}^{j=t+L/2} w_{j-t} \sum_{i=1}^{l} (x_{ij} - \bar{x}_j)^2}{\sum_{j=t-L/2}^{j=t+L/2} w_{j-t} \sum_{i=1}^{l} (x_{ij})^2} \quad \text{(Eq. 3.1)}
\]
where \( w_{j,i} \) is a triangular weighting function ranging from 0 to 1; \( x_{ij} \) is the seismic amplitude at a TWT of \( j \) on trace \( i \); \( \bar{x}_j \) is the average amplitude at a TWT of \( j \) for all traces \( i \); \( L \) is the operator length; and \( I \) is the number of traces in the operator.

Steps 11–12: The seismic volume (~ 30 Gb) was divided into 15 blocks\(^{20} \) (i.e. B1–15, Figure 3.3c) so as not to stretch the virtual memory capacity of the seismic workstation; allowing the screen to refresh quickly and facilitate interpretation. A seismic-slice roll increment of 16 ms enabled interpretation of fault tips to within ± 8 ms of any unconformity where fault tips might arrest. The baseline TWT from which to apply the 16 ms roll increment was taken as ~ 100 ms below the Mackerel Unconformity (Figure 3.3d). Similarly, the roll increment was kept constant at 288 ms below the baseline. The longer roll increment provided an adequate number of TWT slices from which to interpret the fault traces down to a TWT of ~ 4,500 ms.

Minor faults within the seal unit are too numerous (i.e. > 200) to assign individual IDs, so these have been grouped into fault clusters. Accordingly, all minor faults within a particular location block were grouped (e.g. minorB1-to-minorB15 for location blocks B1–15). The majority of fault clusters are present within the Swordfish Formation or equivalent seal interval; where this occurred, the fault ID was changed accordingly (e.g. minorB1-to-minorB1SworSS).

Steps 13–14: Major and minor fault segments were connected to horizons via fault contacts to ensure consistency of fault throw. Fault boundaries were created for each seismic horizon, followed by horizon gridding and contouring.

**Interpretation of a fault zone**

A number of factors were used to infer the location of fault zones, including where faults of differing modes coexist, and/or where the fault trace curves excessively. However, the interpretation of fault zones is limited by the horizontal resolution limitations of the seismic data. For example, broad fault zones can be interpreted from seismic data, whereas narrow fault zones are more problematic to detect; they are normally seen at the outcrop scale. The horizontal resolution at a given depth can be defined by the Fresnel Zone radius (Equation 3.2),

\[
\text{Fresnel Zone radius} = \left( \frac{V_{av}}{2} \right) * \sqrt{\frac{T}{F_{min}}} \quad \text{(Eq. 3.2)}
\]

where \( V_{av} \) = average velocity at MSL, \( T \) = TWT and \( F_{min} \) = dominant frequency.

\(^{20}\) The variance cubes are also used as a location reference throughout the thesis (e.g. location blocks B1 to B15).
The Fresnel Zone radius was calculated at the centre of location blocks B1–B15 and midway within each of eight seismic intervals (EmpeSS, GBeaSS, LHalSS1, LHalSS2, UHalSS1, UHalSS2, UHalSS3–CobiSS and SworSS; Figure 3.5b). The Fresnel Zone radius ranges from ~ 100 m at a depth of 1 km to ~ 700 m at 7 km. The Fresnel Zone radius averages to ~ 220 m within the reservoir and seal intervals of interest, equivalent to ~ 9 seismic inline, or 18 crossline intervals. Thus, fault zone widths of 220 m or less are difficult to resolve from the seismic data.

3.6 Results

3.6.1 Tectonostratigraphic framework

A number of new hiatuses and associated unconformities have been interpreted as a result of the additional hiatus markers recognised in this study, mainly as a result of the use of age-depth plots (Figure 3.6, ‘hiatus marker from age-depth plots’ column). The number of formation tops/hiatus markers has increased from 221 to 531, an increase of 140%. Formation tops available in pre-existing stratigraphic databases are shown in the column ‘Pre-existing Formation tops’. The column ‘Inferred hiatus markers’ shows the additional hiatus markers that could be inferred from the pre-existing formation tops (stratigraphic database). All these markers were cross-checked against the seismic unconformities and/or key reflectors (‘Inferred from seismic’). A few hiatus markers were also correlated from logs (‘Hiatus marker from logs’). The end result is that the interpreted seismic horizons are better constrained from the additional hiatus markers introduced.

The use of age-depth plots significantly improved the understanding of uplift and erosion, in addition to increasing the number of recognised hiatus markers. The number of merging unconformities (Figure 3.6) and evidence of uplift and erosion from age-depth plots can be cross-checked with seismic data and mapped subcrop edges (Figure 3.7). Uplift and erosion of the LHalSS1, LHalSS2, UHalSS1 and UHalSS2 intervals (see Figure 3.4 for nomenclature) resulted in recognition and mapping of subcrop boundaries that extend across approximately 50% of the study area. In addition to erosion caused by structural uplift, erosion also resulted from downcutting of the Tuna-Flounder and Marlin Channels.

Lastly, a thin-to-condensed section occurs in the Early Eocene to Early Miocene interval (c. 52–16.5 Ma). Palynology control of the Cobia Subgroup (a condensed interval) is limited, so the associated CobiSS interval is difficult to interpret on seismic data. The coarse interval used in palynological sampling is also limiting. However, a thicker section in the Marlin Channel is an exception for this subgroup. A second exception is the thick section present in the upper Halibut Subgroup (UHalSS3 interval) that forms the basin fill of the Tuna-Flounder Channel. Similarly, a third thicker, but
Figure 3.6. Merged unconformities of the study area—application to improving on well-tie control.
Coloured rows correspond to interpreted seismic horizons. Columns 'Pre-existing formation tops', 'Inferred hiatus markers' and 'Hiatus marker from logs': see Appendix 1 (Tables A1.17–1.95) for the well-by-well interpretation. Column 'hiatus marker from age-depth': see Appendix 1 (Figures A1.3–1.33) for original interpretation. Column 'Inferred from seismic' (reliability of the seismic pick): P—poor, M—medium, G—good, VG—very good. See Figure 3.4 for unconformity (Unc.) nomenclature.
Figure 3.7. Subcrop maps of the (a) ILHaS1, (b) ILHaS2 and (c) ILHaS3 horizons, (d - e) representative sections.

Source of subcrop and channel edges (see Chapter 5: Figures 5.8-5.9, 5.11). Selected faults imported from Figure 3.12. See Figure 3.4 for seismic horizon and interval nomenclature.
localised, section belonging to the Swordfish Formation exists and forms part of the channel fill called the Swordfish channel in this study.

### 3.6.2 Unconformity/horizon interpretation

#### Basement and Strzelecki Group unconformities and intervals

The basement (BaseMS horizon) and Strzelecki Group unconformities (LStrMS, OtwaSS1 and OtwaSS2 horizons; Figure 3.4) were identified solely from seismic data; these were interpreted under the Northern Strzelecki Terrace. The additional intra-Strzelecki Group unconformities (i.e. OtwaSS1, OtwaSS2 horizons) were interpreted within fault blocks immediately downdip of the Northern Strzelecki Terrace; these are possibly equivalent to Power et al.’s (2001) sequence boundaries ‘strz2a’ and ‘strz2b’. All unconformities are stratigraphically below the main reservoir and seal intervals of the Halibut Subgroup, and so, are not considered further.

#### Otway Unconformity (OtwaMS horizon)

The Otway Unconformity (upper Strzelecki Group) was identified in eight wells that are located across the Northern Strzelecki Terrace (Figure 3.6). This unconformity cannot be clearly defined from age-depth plots due to the relative absence of palynology specimens in the sandy lithology of the Strzelecki Group under the Northern Strzelecki Terrace. The Otway Unconformity is identified in sonic logs by an increase in acoustic impedance and is represented on seismic profiles by high-amplitude seismic reflectors and/or erosional truncation. The seismic interpretation of this unconformity is less reliable south of the Northern Strzelecki Terrace, where seismic reflectivity is reduced. The TWT to the OtwaMS horizon increases from north to south, from the Northern Strzelecki Terrace (~ 1 s) to the eastern Central Deep (~ 4.5 s; Appendix 2: Figure A2.6a). Many faults arrest at the Otway Unconformity (OtwaMS horizon) as seen elsewhere across the basin (Power, 2003)—these faults are not the focus of this study and are not considered further.

#### Longtom Unconformity (LongSS horizon)

The Longtom Unconformity was identified in 10 wells and recognised from age-depth plots in 2 additional wells (Figure 3.6). The age-depth plot recognition is, however, somewhat problematic due to the relative absence of palynological specimens in the sandy lithology of the Emperor Subgroup. The Longtom Unconformity was initially recognised adjacent to the Longtom Field and has been reconfirmed from seismic interpretation in this study. However, erosional truncation of the underlying seismic LongSS interval is not readily resolvable across the study area. The TWT to the LongSS horizon increases from north to south, from the Northern Strzelecki Terrace (~ 1.5 s) to the eastern...
Central Deep (~ 3.8 s; Appendix 2: Figure A2.6b). The EmpeSS interval is faulted, but faults propagate up to the Seahorse Unconformity (SeahSS horizon) and Mackerel Unconformity (MackMS horizon). Faults that arrest at the Longtom Unconformity (LongSS horizon) have been observed in other parts of the basin (Power, 2003).

**Seahorse Unconformity (SeahSS horizon)**

The Seahorse Unconformity has been identified in 10 wells and inferred in 2 other wells, all within or adjacent to the Northern Strzelecki Terrace (Figure 3.6). The unconformity has also been recognised in 8 other wells using age-depth plots (Appendix 1: Table A1.4). The age-depth plot picks are poor to average because of poorer palynological returns from sampling in the sandy lithologies of the Golden Beach Subgroup. One log pick was also made. Intra-Golden Beach hiatuses have not been recognised on age-depth plots, however, evidence of a hiatus exists at Kipper-2 at the top of the *N. senectus* spore-pollen zone (Bernecker et al., 2002). The unconformable nature of the SeahSS boundary is not readily apparent from seismic-reflector truncation across the study area, although erosion observed at the Longtom Field is an exception. The hummocky seismic facies of volcanic bodies can be identified readily in some areas of the study area, thereby further constraining the SeahSS horizon. Depth to the SeahSS horizon increases from north to south, from the Northern Strzelecki Terrace (~ 1,500 m) to the eastern Central Deep (~ 5,500 m; Appendix 2: Figure A2.7a). Approximately 20% of faults arrest at the Seahorse Unconformity (SeahSS horizon) although the majority propagate up to the Mackerel Unconformity (MackMS horizon).

The Chimaera Formation, which underlies the Seahorse Unconformity, is dominantly sandy. The lithology comprises sandstone (60–100%), siltstone (0–20%), claystone/shale (0–10%) and coal (0–10%). Shale is abundant at the base of the Chimaera Formation although conglomerate has also been reported in Tuna-4 (ESSO, 1985a). No sharp lithological transitions have been identified that could imply sequence boundaries. However, reworked surfaces have been interpreted using palaeoenvironmental indicators in Chimaera-1 (Baird, 1992). The thickness of the Chimaera Formation ranges from less than 10 m in some wells to more than 1,000 m in Chimaera-1. Volcanic horizons are present at or near the top of the Formation (Appendix 1: Table A1.96); the composition varies from dolerite, micro-gabbro, to olivine basalt and includes light brown to buff weathered volcanics and volcaniclastics.

**Intra-lower Halibut (ILHalS horizon)**

Lower Halibut Subgroup unconformities have been discussed briefly in the literature (Bernecker and Partridge, 2001; Weber et al., 2004; Hart et al., 2006) but no formal names have been applied. Fitall
and Cvetanovic (1991) interpreted a lower Halibut Subgroup unconformity in the Maastrichtian, at c. 71 Ma, with faults arresting at the unconformity. Maastrichtian and intra-Campanian seismic markers/horizons in Basker-1 (Bernecker et al., 2002) and the MA horizon that represents the intra-\textit{T. longus} marker (Willcox et al., 1992) have also been used as an alternative nomenclature.

The intra-lower Halibut unconformity (ILHalS horizon) has not been previously recognised and is newly defined in this study. The accompanying hiatus is based on age-depth plots from nine wells (Figure 3.6) and inferred from two other wells. The hiatus spans the Maastrichtian from c. 75–66.5 Ma (Appendix 1: Table A1.5) and consistently ranges from c. 68.3–68.2 Ma (Figure 3.8c). Note that two hiatuses have been interpreted in one of the wells (Grunter-1); palynological zone offsets are illustrated in Tuna-4 (Figure 3.8d). Reworked surfaces have also been interpreted in a number of wells (e.g. Bignonose-1 (Shell, 1994); Chimaera-1 (Shell, 1984a); Manta-1 (Shell, 1984b); Volador-1 (Shell, 1983)). In Bignonose-1, for example, the F, G, H-2, I-1, I-2 and J-1 Foraminifera zones are missing. Additionally, two minor unconformities at the base of the Roundhead Member (upper Volador Formation) were inferred from log correlations (Riordan et al., 2004). The hiatus there ranges from c. 68–67 Ma. The unconformable nature of the intra-lower Halibut hiatus is not readily apparent on seismic data. However, the unconformity is indirectly supported by high to low thickness gradient changes below and above it (discussed in Section 3.6.8). The lower Halibut Subgroup subcrops in the southeastern part of the study area (Figure 3.7a). Depth to the ILHalS horizon increases from north to south, from the Northern Strzelecki Terrace (~ 1,500 m) to the eastern Central Deep (~ 4,200 m; Appendix 2: Figure A2.7b). The ILHalS horizon is faulted, with most faults propagating up to the Mackerel Unconformity (MackMS horizon).

**Lower Halibut (LHalSS horizon)**

The lower Halibut unconformity is commonly referred to as the Cretaceous–Tertiary sequence boundary or KTSB (Power, 2003). The KTSB has not been formally referred to as an unconformity when describing the basin’s geohistory (Bernecker and Partridge, 2001; Weber et al., 2004; Hart et al., 2006) although Fittall and Cvetanovic (1991) do interpret an unconformity at the Volador Formation to Kate Shale transition. Partridge (1975, cited in ESSO, 1975b) inferred an unconformity at the KTSB in Hapuku-1 based on palynological frequency information that showed a downwards increase in the spore-pollen count and decrease in the dinoflagellate and architach count. Previous seismic markers that coincide with the LHalSS unconformity include the Maastrichtian and intra-Campanian, both identified in Basker-1 (Bernecker et al., 2002); the near KTSB or lower Kate Shale was also used in a regional study (Power et al., 2003).
The lower Halibut hiatus is readily identified from age-depth plots (Figure 3.6). Palynological zone offsets observed in 18 wells, and inferred in 25 other wells, suggest the presence of a study-wide hiatus having a range of c. 67.5–56.4 Ma (Appendix 1: Tables A1.6–1.7), but consistently with a range of c. 65.7–65 Ma (Figure 3.8c); a palynological zone offset is illustrated in Tuna-4 (Figure 3.8d). Further evidence for an unconformity coincident with the hiatus exists in the form of weak seismic onlap across the Southern Rosedale Fault Zone (described in Section 3.6.4). Additionally, the existence of an unconformity is indirectly supported by high to low thickness changes seen in the sediments below and above it (discussed in Section 3.6.8). The lower Halibut Subgroup subcrops in the southeastern part of the study area (Figure 3.7b). The depth to the LHalSS horizon increases from north to south, from the Northern Strzelecki Terrace (~750 m) to the eastern Central Deep (~3,750 m, Appendix 2: Figure A2.8a). The horizon is faulted, with most faults propagating up to the Mackerel Unconformity (MackMS horizon).

**Intra-upper Halibut (IUHalS horizon)**

Hiatuses within the upper Halibut Subgroup have been occasionally referred to when describing the basin’s geohistory (Bernecker and Partridge, 2001; Weber et al., 2004; Bernecker and Partridge, 2005; Hart et al., 2006). An unconformity at the top of the lower *L. balmei* foraminifera zone was interpreted in Sweep-1 (Bernecker et al., 2002). Seismic indications of hiatuses have also been recognised, for example, the mid-Palaeocene seismic marker in Admiral-1, Basker-1 and Kipper-2 (Bernecker et al., 2002). Willcox et al. (1992) correlated the ‘LP’ seismic horizon to the base of the upper *L. balmei* zone (c. 57 Ma) and interpreted this horizon as a localised unconformity. Fittall and Cvetanovic (1991) interpreted intra-Kingfish Formation unconformities, with faulting arresting at an unnamed unconformity; at c. 56.8 Ma. Partridge (1975, cited in ESSO, 1975b) demonstrated the presence of two disconformities within the upper Halibut Subgroup in Hapuku-1, using palynological frequency information and associated palaeo-zone offsets.

New intra-upper Halibut hiatuses have been defined in this study using age-depth plots (Figure 3.6); hiatuses c. 61–56.6 Ma, c. 55.5–55 Ma and c. 54.1–53.3 Ma are identified within the Kingfish and Mackerel Formations (Appendix 1: Table A1.8). Related hiatuses have also been inferred in an additional 11 wells. On balance, the main hiatus occurs at c. 55.5–55 Ma (Figure 3.8). The hiatuses outside c. 55.5–55 Ma may have significance, but their presence is only seen in a few wells. Seismic evidence for an unconformity linked to the main hiatus exists in the form of weak seismic onlap. However, the existence of the unconformity is indirectly supported by high to low thickness gradient
Figure 3.8. Age-space diagrams, (a) Seaspray Group end, (b) Hallibut - Cobia Subgroups; age-depth plots (b) Flounder-S and, (d) Tuna-4.

Age and sea level curves are after Haq et al. (1987); source of age-depth plots - www.gs.gov.au (data and applications, energy, applications, petroleum wells applications); see Appendix 1 (Table A1.1) for well names; see Figure 3.4 for hiatus nomenclature.
changes both below and above it (discussed in Section 3.6.8). The upper Halibut Subgroup subcrops in the southeastern part of the study area (Figure 3.7c). The depth to the IUHalS horizon increases from north to south, from the Northern Strzelecki Terrace (~ 1,000 m) to the eastern Central Deep (~ 2,900 m, Appendix 2: Figure A2.8b). Most faults in the UHalSS1 interval propagate up to the Mackerel Unconformity (MackMS horizon).

**Mackerel Unconformity (MackMS horizon)**

The Mackerel Unconformity has been identified in 31 wells (Figure 3.6), all located south of the Northern Strzelecki Terrace. A seismic marker representative of a discrete unconformity does not exist for the Northern Strzelecki Terrace as a whole because several unconformities merge here. The Mackerel Unconformity is best differentiated from the Marlin Unconformity within the Tuna-Flounder and Marlin Channels, where eight wells have intersected the channel fill. The unconformity was inferred from 15 other wells where Flounder Formation tops were available and also picked from age-depth plots of 26 wells and from 1 well log (Figure 3.6). The Mackerel Unconformity is by far the most significant unconformity in this part of the basin; it is readily apparent on seismic profiles from the erosional truncation of the sequences underlying it. The depth to the MackMS horizon increases from the Northern Strzelecki Terrace (~ 600 m) to the eastern Central Deep in the south (~ 3,500 m; Figure 3.9). The majority of faults arrest at the Mackerel Unconformity (MackMS horizon).

**Intra-upper Halibut (IFlouS)**

A second intra-upper Halibut hiatus is not formally recognised in the literature, although intra-Flounder Formation hiatuses were identified within the Tuna-Flounder and Marlin Channel fill (Evans, 1992; Partridge, 1999). Away from the channel fill, the Flounder Formation is thin-to-condensed (< 20 m), which makes the identification of any hiatuses difficult. In the study area, an intra-upper Halibut hiatus was recognised in Angelfish-1 and Flounder-4 (Appendix 1: Table A1.9). However, the associated IFlouS horizon was not interpreted in this study because of the absence of well ties and the difficulty involved in interpreting sequence boundaries within a channel fill.

**Marlin Unconformity (MarlSS horizon)**

The Marlin Unconformity is identified in 63 wells, although it can be difficult to resolve from the Mackerel and Latrobe Unconformities, where all three unconformities can merge (Figure 3.6). The thin to condensed section that is representative of the Flounder Formation and Cobia Subgroup limits the recognition of the Marlin Unconformity on seismic data; the Cobia Subgroup is generally less than 20 m thick and barely resolvable using seismic data. Merged unconformities are evident on some age-depth plots (Figures 3.8b, d). Neither TWT nor depth structure maps have been produced.
Figure 3.9. Depth structure—Mackerel Unconformity (MackMS horizon).
See Appendix 1 (Table A1.1) for well names.
Figure 3.10. Depth structure—Latrobe Unconformity (LatrSS horizon).
See Appendix 1 (Table A1.1) for well names.
because the Marlin Unconformity surface approximately parallels the overlying Latrobe Unconformity surface. The depth structure map of the LatrSS horizon substitutes for the Marlin Unconformity (MarlSS horizon) structure.

**Latrobe Unconformity (LatrSS horizon)**

The Latrobe Unconformity is identified in all wells south of the Northern Strzelecki Terrace. The unconformity cannot be identified reliably across the terrace because it merges in places with either the Mackerel or intra-Swordfish Formation unconformities (Figure 3.6). Depth to the LatrSS horizon increases from the Northern Strzelecki Terrace in the north (~600 m) to the eastern Central Deep in the south (~3,500 m); the surface dip on this unconformity also steepens in the southeastern corner of the study area (Figure 3.10). The LatrSS horizon is occasionally faulted, although most faults arrest at the underlying Mackerel Unconformity (MackMS horizon). Seismic reflectors of the SworSS interval downlap onto the LatrSS surface and the seismic reflectors immediately underlying the surface are subparallel to it (Figure 3.3e). Structural contours deviate from the regional trend east of the Turrum Field. There, the edges of the Marlin Channel are interpreted to control the sediment fill of the overlying Cobia Subgroup.

**Intra-Swordfish unconformity (ISworS)**

Intra-Swordfish Formation hiatuses are difficult to recognise from age-depth plots, because the formation is mainly thin to condensed. However, palynological zone offsets are interpreted in the thicker of the sections of 11 wells (Figure 3.6). Three hiatuses are interpreted: c. 27.5–23.2 Ma, c. 22.9–22.5 Ma and c. 18.8–18.7 Ma (Figure 3.8a; Appendix 1: Table A1.10). The latter hiatus is the one most consistently recognised in the wells. A hiatus spanning 10 Myr is identified in Bignose-1 where the F, G and H-1 foraminifera zones (Early Miocene) are absent (Shell, 1994). A hiatus spanning 12–14 Myr is identified in Volador-1 where the H-1, H-2, I-1, I-2 and J-1 foraminifera zones (Oligocene to Early Miocene) are absent (Shell, 1983). Intra-Swordfish hiatus boundaries have not been interpreted from seismic data because the section is thin-to-condensed.

**Swordfish unconformity (SworSS horizon)**

The top of the Swordfish Formation hiatus has been identified from age-depth plots in 25 wells (Figure 3.6). The sequence boundaries previously interpreted in the Cainozoic (Feary and Loutit, 1998) were then tied to the available hiatus markers (Appendix 1: Table A1.11). The SworSS hiatus is centred from c. 17.6 to 16.4 Ma (Figure 3.8a); coinciding with an increase in the depositional rate (Figure 3.8b).
The depth to the SworSS horizon increases from northwest to southeast, and there is a notable steepening of the gradient in the southeastern corner of the study area (Figure 3.11). Seismic reflectors in the overlying CongSS interval downlap onto the SworSS unconformity, while reflectors in the underlying SworSS interval, are seen to toplap (Figure 3.3e). The SworSS interval is relatively unfaulted. However, clusters of minor faults have been interpreted around the Marlin Channel as well as over the southern bounding fault of the RFS (discussed in Section 3.6.6). The depth structure contours deviate from the regional trend near the Marlin Channel, indicating that the channel has had some degree of depositional control on the Swordfish Formation sediments. The Swordfish Formation sediments are associated with generally low rates of deposition, although higher rates occur over the Swordfish channel.

**Post-Swordfish unconformities (CongSS, ICodSS, BullSS and IWhitS horizons)**

Additional hiatuses within the Seaspray Group were interpreted to help constrain the chronology of any reactivated faults (Appendix 1: Tables A1.12–1.16); correlative unconformities (CongSS, ICodSS, BullSS and IWhitS horizons) were then tied to the previously established sequence boundaries of Feary and Loutit (1998, see Figure 3.11 for traverse location). Age-depth plots show increased depositional rates in post-Swordfish Formation intervals (Figure 3.8b). Approximately 2 km of sediment was deposited from the Oligocene onwards (16.5 Ma to present), which equates to a minimum nondecompacted depositional rate of 0.25 mm/yr. Depth to all horizons increases from northwest to southeast, steepening in the southeast corner of the study area, similar to the SworSS horizon (Figure 3.11). Also, Mid-Miocene channeling is observed within the seismic intervals associated with the hiatus-related boundaries. However, the channeling is not central to this study, descriptions can be found elsewhere (Holdgate et al., 2000; Norvick et al., 2001; Norvick, 2005). In addition, the above horizons have subsequently been found to be only marginally relevant to the study objectives so that detailed descriptions, such as was done for other unconformities (e.g. IUHaiS), are not included.

**3.6.3 Fault interpretation—outline**

Interpreted faults are described in Sections 3.6.4–3.6.6 according to which seismic interval (or equivalent stratigraphic interval) they occur in (i.e. Strzelecki, Latrobe and Seaspray Group). However, the Latrobe and Seaspray-Group intervals are more pertinent as these correspond to the reservoir and seal intervals being considered. Also, fault planes that transect more than one seismic interval are described under the younger interval. Further, described below are the fault parameters (location, style, mode, family, array and zone), fault-province families (RFS, SRFZ, broad terrace and eastern Central Deep) and sub-province families (west and east Tuna, west and east Tuna).
Figure 3.11. Depth structure—Swordfish unconformity (SwordSS horizon).
See Appendix 1 (Table A1.1) for well names.
Flounder; Figure 3.12a), fault-throw families (major and minor faults), upper tip-line bound and fault chronology.

In total, in the order of 200 major and minor faults were interpreted, of which ~ 40% are stand-alone faults, and 60% are kinematically linked and subsequently grouped into fault arrays. Only those arrays located in the broad terrace province family, including west and east Tuna, west and east Flounder, are referred to as arrays as these are located within the thicker sedimentary fill and section of interest (Figure 3.12a). The fault-trend descriptions that follow make use of stereonet and maximum throw versus fault-trace length plots. The extent of the fault interpretation is overwhelming so that it is impractical to represent fault-trend information at all seismic-horizon levels. As a compromise, the fault-trend information is represented at the LHalSS horizon, midway within the Halibut Subgroup. Lastly, the tectonostratigraphic evolution is described: it incorporates the fault interpretation, the fault chronology and the TWT thickness mapping.

3.6.4 Fault interpretation (Strzelecki and lower Latrobe Groups)

Rosedale Fault System

Fault styles within the Rosedale Fault System (RFS) include up-to-basin and down-to-basin domino faults (Figures 3.13a–b) and also synthetic–antithetic fault arrays (Figures 3.13c–d). The fault mode varies from normal, to reverse, to combined reverse–strike-slip. Therefore, the RFS can be seen as comprised of numerous fault arrays with differing modes and geometry that form distinct structural domains within one large fault system that is up to 15 km wide (Figure 3.12a).

Thirty seven faults arrest at the merged Otway-to-Seahorse Unconformity surface (i.e. Seah faults). The above mentioned synthetic–antithetic fault array shows fault traces extending up to 10–15 km, with the fault-array width being 5–8 km (Figure 3.13c). The latter fault array is notable in that it has a wrench style and may have formed part of the incipient RFS. The array was relayed and now forms a distinct structural domain within the RFS. The strike distribution of Seah faults (sampled at a 100 m interval) is unclustered (Figure 3.12b), as expected from numerous fault arrays with differing modes and geometry. Average maximum fault throws and fault-trace lengths are less than 150 m and 15 km, respectively (Figure 3.12c).

Smaller-scale faults and fault arrays are likely to exist within the larger-scale faults that have been interpreted across the Northern Strzelecki Terrace. For example, many faults offset the OtwaMS horizon, but not all of these faults have been interpreted as the fault offset can be difficult to substantiate, particularly where the conjugate section is absent on the upthrown side.
Rosedale Fault, North Rosedale Fault, South Rosedale Fault zone
The RFS is bounded to the south by two fault arrays commonly referred to as the North Rosedale Fault and Rosedale Fault (Power, 2003; Power et al., 2003); these comprise normal, down-to-basin faults. The North Rosedale Fault and Rosedale Fault are recognised in this study by faults F54, F177, F37 and F66 and, faults F58, F55 and F214, respectively (Figure 3.12a). The North Rosedale fault, Rosedale Fault and RFS all extend beyond the 3-D seismic grid and study area (see Figure 2.1). In this study, the rock volume between the North Rosedale Fault and Rosedale Fault is considered to form part of a fault zone and, referred to in this study as the southern Rosedale Fault zone (SRFZ).

Within the study area, the fault throw within the SRFZ is modest, being less than 200 m, with fault-trace lengths less than 15 km (Figure 3.12c). However, the fault throw can sometimes be difficult to quantify because the displaced section is missing on the upthrown side of the SRFZ. A large fault-throw of 1,500 m has been estimated within the SRFZ, but outside of the study area (Power, 2003). Also, the SRFZ faults arrest at the Mackerel Unconformity (MackMS horizon), the latter merging in places with the Marlin (MarlSS horizon) and Latrobe Unconformities (LatrSS horizon).

Flounder deep and Turrum deep faults
Two deep-seated faults are inferred by the presence of hinge lines under the Flounder and Turrum Fields, and are referred to here as Flounder deep and Turrum deep (Figure 3.12a). These faults are important within the study area as they appear to have controlled sediment deposition within the Latrobe Group, as evidenced from the increase in thickness across them. These faults have not previously been recognised as being regionally significant (Power, 2003; see Chapter 2: Figure 2.4). However, they are present below the main reservoir and seal intervals of the Halibut Subgroup and so, are not considered further.

Eastern Central Deep and broad terrace
Changes in dip of the Strzelecki Unconformity surface have been shown to occur across a lineation that was subsequently interpreted as a NNE–SSW trending strike-slip fault (Lowry and Longley, 1991). This strike-slip fault was initially named the Judith Transfer (Lowry and Longley, 1991), then Kingfish-Tuna transition zone (Gilbert and Hill, 1994) and later abbreviated as the KTTZ (Power, 2003; Power et al., 2003). However, the KTTZ was not recognised in this study and does not propagate through the overlying Latrobe Group section. Further, only a few Otwa faults have been identified within the eastern Central Deep and broad terrace (F107, F130 and F168); however, not all
Figure 3.12. Subprovinces, fault families, fault zones, fault arrays and stand-alone faults - study area.

Fault-province families (broad terrace, eastern Central Deep, southern Rosedale Fault zone, Rosedale Fault System), subprovince families (east and west Flounder, east and west Tuna), fault arrays (A1 to A13) and stand-alone faults (F1 to F217). Some faults interpreted in this study have equivalents in Power’s study (2003) (i.e. Blue East, Blue West, Green East, Green West and Salmon). All strike, fault trace length and maximum throw are taken at the LH-11SS horizon.
Figure 3.13. Seismic interpretation of the Rosedale Fault System.
Seismic interval and horizon captions shown in parts a and d, respectively. See Figure 3.4 for seismic interval, horizon and fault nomenclature.
Otwa faults have been interpreted, especially where these are deeper than the reservoir and seal intervals of interest.

### 3.6.5 Fault interpretation (Latrobe Group)

**Broad Terrace (east and west Flounder fault families)**

Two fault arrays (A8–9) located in and adjacent to the Flounder Field form the east Flounder fault family (Figure 3.12a). These faults trend WNW–ESE, the strike being 120° ± 30°N and dip 60° ± 15°, to the NNE and SSW (Figure 3.12f). The maximum throw and average trace length for all faults...
are less than 200 m and 10 km, respectively (Figure 3.12g). Both normal and reverse mode faults are interpreted within this short-offset flower structure (Figures 3.14b, c). The west Flounder fault family (F11, F132 and F192) trends NE–SW (Figure 3.12a), the strike being 040° ± 20°N and dip near-vertical at 80° ± 10°, dipping to the SE (Figure 3.12f). The maximum throw and average trace length for the 3 faults are less than 150 m and 5 km, respectively (Figure 3.12g). The faults are reverse mode and are offset by the WNW–ESE-trending faults of the east Flounder family. The offset suggests that the west Flounder faults either predate the east Flounder faults, or were coeval with them (i.e. bearing in mind the offset is < 0.5 km), possibly within a transpressional system.

Immediately west of the Flounder Field, the mode and style of faulting changes to a more typically down-to-basin domino set of normal faults (Figure 3.14a). There, faults are on average approximately 10° shallower-dipping than for the east and west Flounder faults. To summarise, the east and west Flounder faults are together more complex than the domino set of normal faults that dominates the broad terrace area. Consequently, fault zones are interpreted adjacent to where east and west Flounder faults intersect (pink area in Figure 3.12a).

**Broad Terrace (east and west Tuna fault families)**

Three faults and one fault array (A4) that are interpreted adjacent to and east of the Tuna Field form part of the east Tuna fault family (Figure 3.12a). These faults trend WNW–ESE, the strike being 100° ± 40°N and dip 65° ± 10°, dipping to the SSW and NNE (Figure 3.12d). The maximum throw and average trace length for all faults are less than 70 m and 4 km (excluding the Rosedale Fault), respectively (Figure 3.12e). The minor *en echelon* splay faults (F57, 151–152, 179–181, 189 and 210), alternatively interpreted as Riedel faults, form part of the short-offset strike-slip system involving master fault F55 (Figures 3.15a–b).

One fault and three fault arrays (A10, 12–13) that are interpreted adjacent to and west of the Tuna Field form part of the west Tuna fault family (Figure 3.12a). Most faults generally trend WNW–ESE, the strike being 100° ± 40°N and dip 65° ± 10°, dipping to the SSW and NNE (Figure 3.12d). The maximum throw and average trace length for most faults are less than 60 m and 6 km, respectively (Figure 3.12e). However, exceptions exist within fault arrays A10 and A12–13 where F91 and F165 trend NE–SW (Figure 3.12a). Fault F165 soft-branches towards F55 but is offset by an east–west relay ramp (Figure 3.15a). The fault also forms the north-western limb of the Tuna Field structure, is arcuate in form, reverse in mode and curves towards the western limb of the West Tuna Field structure. Fault F165 subparallels F91, the strike being 060° ± 10°N and dip 65° ± 10°, dipping to the SE (Figure 3.12d); the fault trend of both faults is notably different in orientation to all other faults.
in fault arrays A10 and A12–13. Both the above faults are reverse mode faults (Figure 3.16) and are part of a regional compressional front (also discussed in Section 3.6.8), which includes the west

Figure 3.15. Fault zones associated with the east and west Tuna fault families.
(a) Seismic slice of the variance attribute at location blocks B10–B12 (see Figure 3.12a for location). (b–c) Seismic sections depicting en echelon fault arrays. F86—fault ID. See Figure 3.4 for horizon and fault nomenclature.
Figure 3.16. Seismic interpretation of an inversion structure—Tuna Field.
Seismic interval and horizon captions shown in parts a and c, respectively. See Figure 3.4 for seismic interval, horizon and fault nomenclature.
Figure 3.17. Seismic interpretation of an inversion structure—Flounder Field. Seismic interval and horizon captions shown in parts a and d, respectively. See Figure 3.4 for seismic interval, horizon and fault nomenclature.
Flounder fault family (F11, F132 and F192; Figure 3.17). It is suspected that the area between faults F91 and F165 would encompass a fault zone(s) based on the presence of intersecting fault trends, varying fault modes (reverse and strike-slip), as well as from curvature of the fault traces. Lastly, fault F35, within fault array A10, has previously been called ‘Fault Blue West’ (Power, 2003).

Broad Terrace (other faults)

Four fault arrays (A1–4) and 56 stand-alone faults extend across the broad terrace in addition to the faults and arrays that form part of the east and west Flounder, as well as east and west Tuna fault families (Figure 3.12a). These faults trend NW–SE, the strike being 130° ± 50°N and dip 65° ± 15°, dipping mostly SSW but also NNE (Figure 3.12h). The maximum throw and average trace length for all faults are less than 200 m and 15 km, respectively (Figure 3.12i). Some faults have been given official names in past studies; for example, Green East (F6), Green West (F8), Salmon (F9), respectively (Power, 2003). These faults tend to have longer fault-trace lengths (10–20 km) and greater throw (100–200 m). Major fault F56 splay s off master fault F55 (Rosedale Fault; Figure 3.15c), while other faults (some with minor throw) form an en echelon pattern (F19, F135 and F174) off it. Together, these faults form a larger scale fault complex that encompasses the Kipper Field structure.

Most of the broad-terrace faults are major, normal faults that form part of down-to-basin domino fault sets (Figure 3.18); most arrest at the Mackerel Unconformity surface (Mack faults). However, eight faults arrest at the Seahorse (SeahSS horizon) and lower Halibut unconformities (LHaISS horizon). Other faults (e.g. F6, F14, F20 and F90) are offset by relay ramps where fault zone(s) could be speculated on as being present.

There is no substantial evidence to support the existence of faults within the channel fill (intrachannel) of the Tuna-Flounder Channel (Figure 3.19a). However, although potential lineaments were noted, these are not an extension of the Mack faults up into the channel fill (Figure 3.19b). The lineaments are thought more likely to reflect facies boundaries, and possibly, associated compaction effects. The seismic variance character of facies boundaries is more diffuse along the trace length and, is also not consistent between variance slices.

An important ramification for potentially differentiating two sets of faults (i.e. Mack and intrachannel) is that the Latrobe Unconformity has long been recognised (Holdgate et al., 2003a) as a major megasequence boundary in the basin (see Chapter 2). Therefore, one should expect that a large proportion of the faults associated with a megasequence boundary to arrest at the boundary. The evidence here shows that faults do not arrest at the Latrobe Unconformity surface (LatrSS horizon),

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Figure 3.18. Seismic interpretation of Early Eocene normal faults. Seismic interval and horizon captions shown in parts a and d, respectively. See Figure 3.4 for seismic interval, horizon and fault nomenclature.
Figure 3.19. Fault trace interpretation, (a) near-base and, (b) below the Tuna-Flounder Channel.
but at the Mackerel Unconformity surface (MackMS horizon). The time gap between the Mackerel and Latrobe Unconformities is c. 16–20 Ma, a significant period during which fault reactivation could have occurred. However, nearly all faults arrest at the Mackerel Unconformity surface (MackMS horizon) so that the faulting clearly predates the Latrobe Unconformity.

**Eastern Central Deep**

Twenty five faults and three fault arrays (A5–7) were interpreted in the eastern Central Deep (Figure 3.12a). These faults trend NW–SE, the strike being 140° ± 40°N and dip 65° ± 15°, dipping NE and SW; the stereonet plot is similar to that for faults of the broad terrace family (Figure 3.12h). The maximum throw and average trace length for all faults are less than 500 m and 11 km, respectively; the crossplot is similar to that for fault of the broad terrace family (not shown, but similar to Figure 3.12i). Faults F39, F43, F45 and F46 have significant fault trace lengths, being 12, 9.9, 10.8 and 10.4 km, respectively. Almost all these faults arrest at the Mackerel Unconformity surface (MackMS horizon); although examples exist such as fault F29 that arrests at the lower Halibut Unconformity surface (LHalSS horizon) and, F47 at the Latrobe Unconformity surface (LatrSS horizon).

**Fault statistics**

Most faults arrest at the Otway (Power, 2003; in this study 1.6%), Seahorse (20%) and Mackerel Unconformities (67.4%; Table 3.1); thus, supporting three dominant associated faulting events. The arrest of fault tips at the Otway Unconformity is less well represented in this study as not all faults were interpreted (i.e. the unconformity lies below the reservoir intervals of interest). Some evidence for secondary faulting events exists; however, a greater sampling of faults across the entire Gippsland Basin would be necessary to substantiate this observation.

<table>
<thead>
<tr>
<th>Fault set</th>
<th>% faults</th>
<th>Fault no</th>
</tr>
</thead>
<tbody>
<tr>
<td>IWhi</td>
<td>0.5</td>
<td>190</td>
</tr>
<tr>
<td>Bull</td>
<td>0.5</td>
<td>129</td>
</tr>
<tr>
<td>ICod</td>
<td>2.2</td>
<td>127, 163–164, 172</td>
</tr>
<tr>
<td>Cong</td>
<td>3.3</td>
<td>7, 17, 64, 78, 83, 129, 166 plus fault clusters</td>
</tr>
<tr>
<td>Swor</td>
<td>0.5</td>
<td>131, Fault clusters</td>
</tr>
<tr>
<td>Latr</td>
<td>1.6</td>
<td>47, 56, 134</td>
</tr>
<tr>
<td>Marl</td>
<td></td>
<td></td>
</tr>
<tr>
<td>IUHal</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>LHal</td>
<td>2.2</td>
<td>29, 31–32, 121</td>
</tr>
<tr>
<td>ILHal</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>Long</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>Otwa</td>
<td><strong>1.6</strong></td>
<td>107, 130, 168</td>
</tr>
</tbody>
</table>

Table 3.1. Fault sets depicting the upper tip-line bound—application to differentiating faulting events. See Figure 3.4 for fault nomenclature.
3.6.6 Fault interpretation (Seaspray Group)

It is important to establish the occurrence of any reactivation by potentially linking the faulting in the Seaspray Group to faulting in the Latrobe Group (e.g. Mack faults). For example, Seaspray Group faults could sole out in the underlying UHalSS3 and/or CobiSS intervals; thus, potentially breaching local seals.

A first set of Seaspray Group faults is located in blocks B4, B7 and B15 (F127, F129, F163–164, F166, F168, F172 and F190; Figure 3.12). These particular faults are interpreted within the CongSS, LCodSS and UCodSS intervals, are characterised by low angle fault planes of approximately 2°. These faults sole out in the CongSS interval with upper tip-lines that arrest as shallow as the BullSS horizon (e.g. F129): these are interpreted as slump faults. An alternative interpretation is that the fault planes could be representing cut and fill boundaries associated with the mid-Miocene Channels. The above set of faults does not appear to have a tectonic origin and are possibly caused by localised slope instability resulting from basin margin flexure and/or sag.

The second set of Seaspray Group faults, located in blocks B1–B4, forms fault clusters above the edges of the Marlin and Swordfish channels (Figure 3.20a). All faults are normal, high angle but are randomly orientated. Fault planes intersect the SworSS horizon and arrest upwards within the overlying CongSS interval. Most fault planes arrest downwards within the SworSS interval but a few appear to propagate deeper down to the CobiSS and UHalSS3 intervals. However, most of these latter faults overlie buried escarpments and/or structural hinge zones, and as such, are not necessarily kinematically linked to the Latr or Mack faults (Figure 3.20a). The two distinct fault sets result from differing faulting events.

The third set of Seaspray Group faults involves fault clusters located in block B10 (Figure 3.20b). All faults are normal, high angle but are randomly orientated. Similarly to the second set of Seaspray Group faults, the fault planes arrest within the overlying CongSS interval. Faults form fault clusters that are distributed across the SRFZ and associated hinge lines and boundary faults (F58 and F54). The fault clusters are interpreted to have formed as a result of intermittent, short-offset, strike-slip movements along the RFS. The fault clusters are bounded to the southeast by escarpments that are formed as a result of Early Eocene inversion. Although the SworSS interval is less than 80 m thick, an escarpment is still recognised at the SworSS horizon (Figure 3.20b), immediately to the north of the Tuna Field. It was noted that escarpments could be differentiated from faults using the seismic variance attribute; escarpments have a diffuse amplitude zone, while faults have a well-defined high-amplitude black peak.
Lastly, some Latrobe Group faults have propagated by various mechanisms into the overlying Seaspray Group interval. For example, the upper tip-line bound for faults F64, F78 and F83 is in the CongSS, UCodSS and LCodSS intervals, respectively (e.g. F78 is shown in Figure 3.21); all these faults were initiated and/or reactivated in the Late Miocene. Of these, fault F78 bounds the northwestern limb of the Longtom Field structure; this is the youngest structure within the study area. Fault F47, located in the eastern Central Deep, has a relatively long fault trace of 12.3 km and, is one of the few faults that arrests in the SworSS interval. However, it is uncertain whether F45 resulted from localised reactivation, or may be linked to differential compaction and drape effects. Lastly, fault F56 splays off the Rosedale Fault and the upper tip-line bound extends up to the CongSS interval in one location (Figure 3.15c). However, it is uncertain if the fault was reactivated as a result of a specific regional tectonic event, or by more random intermittent movements along the RFS and/or SRFZ, maybe caused by flexure over basin bounding hinge lines. With the exceptions noted above, very few Mack faults (Latrobe Group) appear to have been reactivated, so that there is limited direct kinematic linkage between the Latrobe and Seaspray Group faults.

### 3.6.7 Fault regimes

Normal faulting is the dominant mode of the Otwa and Mack faults present, as demonstrated in Section 3.6.5. An extensional regime, therefore, would have dominated from the Cenomanian to the Early Eocene during deposition of the Latrobe Group reservoir and seal intervals with periods of maximum extension likely preceding, and/or coinciding with, formation of the Otway and Mackerel Unconformities.

The rift and fault-controlled basin-margin sag in the Palaeocene to Early Eocene, gave way to subsidence-driven, basin-margin sag in the post-Early Eocene. However, localised faulting could still occur during the basin-margin sag-phase as a result of differential flexure occurring across hinge lines (described in Section 3.6.6). The period when flexure may have occurred coincides with the CongSS interval, within which fault clusters are present (e.g. adjacent to the SRFZ, Figure 3.20b). In addition, the mid-Miocene Channels and the slump faults interpreted within the CongSS, LCodSS and UCodSS intervals may be indicative of shelf instability and/or localised faulting.

Several compressional events are recognised. A compressional regime could have accompanied the uplift and erosion associated with the Otway Unconformity. Compressional regimes have also been inferred during the Campanian and Early Eocene (Figure 2.2); these link to the formation of the Seahorse and Mackerel Unconformities, respectively. Early Eocene reverse faults and inversion structures were previously described within the Tuna, West Tuna and Flounder Fields (Power, 2003;
Figure 3.20. Minor faults in the CongSS–SworSS intervals, (a) seismic section and, (b) slice (variance attribute).
(a) No kinematic link between the Mack and Swor/Cong faults exists, (b) fault clusters overlying and adjacent to the South Rosedale Fault zone. See Figure 3.4 for seismic interval, horizon and fault nomenclature.
Power et al., 2003) and, have also been recognised in this study to be present at the Pilotfish-1 structure (Figure 3.22). However, the inversion structures and associated compressional regime are subtle in places as the syn-sedimentary section also subsided during the basin-margin sag-phase. A younger compressional regime was also demonstrated to be active in the Middle Miocene during deposition of the Cod Formation (Longtom structure). Areas outside of the study area (see Figure 2.5) have been demonstrated as having sustained a compressional regime during the Oligocene to Miocene (Power, 2003). Onshore, an even younger compressional event has been recognised in the Pleistocene (Dickinson et al., 2001; Dickinson et al., 2002).

Of these compressional regimes, the regime active during the Early Eocene is the most relevant, because of its association with the reservoir and seal intervals of interest. The axes of compression-related folds have been demonstrated in this study to trend NE–SW. It is also observed that the compression-related folds, from the oldest (Cenomanian) to the youngest (Pleistocene), appear to be
initially found in a more distal offshore location, and to retrograde to a more proximal onshore location. Unfortunately, conclusively demonstrating this observation is not possible because of the limited extent of the study area.

A strike-slip fault regime could be inferred within the Flounder Field, as implied by the offset of the NE–SW trending reverse faults (Figure 3.12). However, conclusively demonstrating whether any strike-slip fault movement forms part of a distinct regime, or is transpressional in its origin, is difficult. Similarly, episodic movements are inferred to have occurred from the Cretaceous to the present along the length of, and during formation of the RFS (Power, 2003). Lastly, the contemporary state of stress has been estimated as being intermediate compression to strike-slip, based on borehole breakout and drilling-induced, tensile fracture interpretations (Nelson et al., 2006b).

![Seismic interpretation of inversion structures](image)

**Figure 3.22. Seismic interpretation of inversion structures—Tuna, Flounder and Pilotfish Fields.** Seismic trace decimation—15. See Figure 3.4 for horizon and fault nomenclature.

### 3.6.8 Basin phases/subphases

Two basin phases were previously described in Chapter 2. Basin phase 2 can be further subdivided into subphases based on the seismic interpretation and the evidence brought forward in this study. These additional subphases allow subtleties of basin evolution to be recognised.
**Basin phase 1**
The LStrMS horizon (see Figure 3.4) has been interpreted under the Northern Strzelecki Terrace: it is related to the Lower Strzelecki Unconformity, which separates a Lower Cretaceous syn-rift section (c. 165–125 Ma) from a Valanginian to Cenomanian postrift section (c. 125–92 Ma). A horst-graben geometry is characteristic of the Strzelecki Group within its syn-rift and postrift sections. Numerous faults arrest at the Otway Unconformity (OtwaMS horizon), which is the upper bound of the basin phase-1 section.

**Basin subphase 2A**
Basin syn-rift subphase 2A is represented stratigraphically by the Emperor Subgroup (EmpeSS interval, Figure 3.23). The Subgroup/seismic interval is top-bounded by the Longtom Unconformity (LongSS horizon). The basin-scale extension associated with the syn-rift incorporates both half grabens and growth faults interpreted in the eastern Central Deep; there, the EmpeSS interval’s TWT thickness shows a depocentre with up to 1.25 s of basin fill (Appendix 2: Figure A2.9a).

**Basin subphase 2B**
Basin transition-rift subphase 2B is represented stratigraphically by the Golden Beach Subgroup (GBeaSS interval, Figure 3.23). The GBeaSS interval’s TWT thickness shows a depocentre (~1.25 s) located north of NW-SE trending faults (Appendix 2: Figure A2.9b). The subgroup/seismic interval is top-bounded by the Seahorse Unconformity (SeahSS horizon). The upper boundary of this subphase is marked by cessation of normal and growth faulting associated with the syn-rift and transition rift phases. Although extension dominates this subphase, compression is also noted directly east of and within the Teracllin structure, where a half graben has been inverted. Also of significance is that the majority of volcanic flows, intersected in offshore wells, overlie the Chimaera Formation (Appendix 1: Table A1.96), which coincides with the top of the GBeaSS interval and cessation of basin subphase 2B.

**Basin subphase 2C**
Basin rift-drift subphase 2C is represented stratigraphically by the lower part of the Halibut Subgroup (LHalSS1 interval, Figure 3.23). The term rift-drift has been used to reflect the fact that the Eastern Margin’s greater New Zealand–Lord Howe Rise plate was drifting in association with opening of the Tasman Basin, while the Southern Margins were rifting. The subgroup/seismic interval is top-bounded by the ILHalS unconformity (ILHalS horizon). Three key observations substantiate the presence of this subphase. First, the style of basin fill changes from a half-graben geometry to a fault-controlled, sag fill geometry (i.e. compare Figures A2.9b and A2.10a—Appendix 2); a half-
Figure 3.23. Events chart - Latrobe and Seaspray Groups.
graben geometry is characteristic of the underlying GBenaSS interval. Second, the LHalSS1 interval thickens into the eastern Central Deep (Appendix 2: Figure A2.10a). However, the seismic interval subcrops in the southeastern part of the 3-D seismic grid so that the thickness trends are less constrained there (Figure 3.7a). Third, 20% of faults arrest at the Seahorse Unconformity (SeahSS horizon), but not at the intra-lower Halibut unconformity (ILHalS horizon).

**Basin subphase 2D**

Basin rift-drift subphase 2D is represented stratigraphically by a lower part of the Halibut Subgroup (LHalSS2 interval, Figure 3.23). This subgroup/seismic interval is top-bounded by the LHalSS unconformity (LHalSS horizon). Evidence for recognising a basin rift-drift subphase is twofold. First, the basin fill style changes from a fault-controlled, sag fill to a subsidence-controlled sag fill, as demonstrated by the absence of faults arresting at the top of the LHalSS2 interval. The isopach map of the LHalSS2 interval show thickening into the eastern Central Deep (Appendix 2: Figure A2.10b). However, thickening of the LHalSS2 interval is less pronounced than for the LHalSS1 interval. Also, the LHalSS2 interval subcrops in the southeastern part of the 3-D seismic grid so that the thickness trends are less well constrained there (Figure 3.7b). Secondly, the end of the rift-drift basin phase is marked by a maximum flooding event; that is, stratigraphically equivalent to the Kate Shale (Partridge, 1999). The lithology changes from sandstone, shale and coal below the Kate Shale to a sequence dominated by sandstone above it (Figure 2.2). These features, along with evidence of a hiatus defined from the age-depth plots (Figure 3.7b), provide the basis for interpreting the interval as part of a distinct basin subphase.

**Basin subphase 2E**

Basin rift-drift subphase 2E is represented stratigraphically by the upper part of the Halibut Subgroup (UHalSS1 interval, Figure 3.23). The subgroup/seismic interval is top-bounded by the intra-upper Halibut unconformity (IUHalS horizon). The isopach of the UHalSS1 interval thickens into the eastern Central Deep (Appendix 2: Figure A2.11a). However, because this interval subcrops extensively within the study area, thickness trends are uncertain in places (Figure 3.7c). The seismic evidence that indicates uplift and/or erosion associated with the unconformity is limited to subtle truncational erosion. Further, few faults arrest at the intra-upper Halibut unconformity (IUHalS horizon), so that tangible structural evidence for this boundary is lacking. Nevertheless, the hiatus is defined on age-depth plots (Figure 3.7c), and these provide the limited basis upon which to interpret the interval as part of a distinct basin subphase.
Basin subphase 2F
Basin rift-inversion subphase 2F is represented stratigraphically by the upper part of the Halibut Subgroup (UHalSS2 interval, Figure 3.23). The subgroup/seismic interval is top-bounded by the Mackerel Unconformity (MackMS horizon). The isopach of the UHalSS2 interval thickens into the eastern Central Deep (Appendix 2: Figure A2.11b). However, the UHalSS2 interval subcrops extensively across the 3-D seismic grid so that thickness trends are uncertain in places. The Mackerel Unconformity is the most significant unconformity in this part of the basin, with the majority of faults arresting at the unconformity surface. An extensional regime exists and, an inversion regime is postulated to have been active during this basin subphase. Inversion is interpreted from the presence of inversion structures at the Tuna and Flounder Fields, as well as at Pilotfish-1 (Figure 3.22). These observations, together with evidence of a hiatus defined from age-depth plots, provide the evidence upon which to interpret the interval as part of a distinct basin subphase.

Basin subphase 2G
Basin rift-inversion subphase 2G is represented stratigraphically by the Flounder Formation or equivalent Tuna-Flounder Channel sedimentary fill (UHalSS3 interval, Figure 3.23). The formation/seismic interval is top-bounded by the Marlin Unconformity (MarlSS horizon). The thickness of the Tuna-Flounder Channel is up to approximately 450 m, outside of which the interval thins to less than 50 m (Figure 3.24). Faults do not appear to arrest at the Marlin Unconformity, although fault tips are difficult to recognise within the thin UHalSS3 interval (see Section 3.6.6). Structural inversion within the interval is recognised although alternatives for dating the event are possible (see Section 3.7). Nevertheless, the term rift-inversion has been applied to this subphase.

Basin subphase 3A
Basin sag subphase 3A is represented stratigraphically by the Cobia Subgroup (CobiSS interval; Figure 3.23). The subgroup is top-bounded by the Latrobe Unconformity (LatrSS horizon). Sediments of the Turrum Formation (Upper Cobia Subgroup) or equivalent Marlin Channel sedimentary fill have been interpreted as being younger than the sediments of the Tuna-Flounder Channel (Partridge, 1999). The CobiSS interval thins to less than 50 m, but within the Marlin Channel is up to approximately 280 m thick. Few faults arrest at the Latrobe Unconformity (LatrSS horizon). The above features, together with evidence of a hiatus identified from age-depth plots, provide the basis upon which to interpret the interval as part of a distinct basin subphase.
Figure 3.24. Isopach map—combined UHaISS3–CobiSS intervals.
See Appendix 1 (Table A1.1) for well-name abbreviations.
Basin subphase 3B

Basin sag subphase 3B is represented stratigraphically by the Swordfish Formation (SworSS interval, Figure 3.23). The formation/seismic interval is top-bounded by the Swordfish unconformity (SworSS horizon). The Swordfish Formation is characteristically a thin-to-condensed section (Partridge, 1999), which is consistent with an interpreted highstand, also demonstrated to extend onshore (Holdgate and Sluiter, 1991). The interval is less than 80 m thick within most of the study area. However, the SworSS interval thickens up to 650 m in the Swordfish channel, which is located in the southern part of the study area. There, it is probable that sediment deposition has been localised within the associated submarine canyon. This subphase is well defined as the depositional rates increase substantially above the associated SworSS interval.

Basin subphase 3C

Basin sag subphase 3C is represented stratigraphically by the Conger, Cod and lower Whiting Formations (CongSS, LCodSS, UCodSS and LWhiSS intervals; Figure 3.23). The CongSS to UWhiSS intervals thicken from 250 m in the north, to 2,250 m in the south (Figure 3.25). However, the combined intervals thin to the southeast of the study area, in a more basinwards setting, where depositional rates were reduced as is typical of shelf carbonate-wedge geometries (Bernecker et al., 1997). This subphase is characterised by significant Mid Miocene channeling, high depositional rates and margin slump-faulting. There is evidence for a compressional regime, based on reverse faults interpreted as part of the Longtom Field structure, as previously discussed (see Section 3.6.6).

Basin subphase 3D

Basin sag subphase 3D is represented stratigraphically by the upper part of the Whiting Formation (UWhiSS interval, Figure 3.23). There is no evidence of a compressional regime noted within the study area although evidence of inversion structures, in the Pliocene and Pleistocene sediments, are interpreted in onshore Victoria (Dickinson et al., 2001; Dickinson et al., 2002).

3.7 Discussion

Establishing at which unconformities fault tips arrest forms an important part of this study as unique faulting events (or episodes) can potentially be differentiated by this process. The differentiation of faulting events is important when describing the basin evolution, the chronology for trap-formation and to make associations to seal integrity at the fault tip. Also, fundamental structure-related issues are discussed in what follows, mostly in chronological order.
Figure 3.25. Time thickness map—combined CongSS to UWHiSS intervals.
See Appendix 1 (Table A1.1) for well-name abbreviations.
3.7.1 Tuna-Flounder and Marlin Channels

The Early Eocene, Mackerel Unconformity (MackMS horizon, c. 52–51.5 Ma), intra-Flounder unconformity (IFlouS hiatus, 50.5 Ma) and Marlin Unconformity (MarlSS horizon, 49.5 Ma) are coincident with deposition of the Flounder Formation sediments present in the Tuna-Flounder Channel. It is important to assess what the mode and timing of formation of the channel could tell us about traps subcropping against the channel base and, whether basin evolution and understanding of trap formation could be improved on.

Intra-Flounder Formation hiatuses and associated sedimentary sections have previously been identified within the Tuna-Flounder Channel complex (Partridge, 1999). Three Early Eocene channel forming intervals have been differentiated: the basal Flounder Channel interval (c. 52–51.5 Ma), the mid-Tuna Channel interval (c. 51.5–50.5 Ma) and an upper unnamed Channel interval (c. 50.5–49.5 Ma), with the latter top-bounded by the Marlin Unconformity. The IFlouS hiatus identified in this study would be equivalent to one of the above time-distinct ages and, would merge with the Mackerel and Marlin Unconformities at the edges of the Tuna-Flounder Channel complex.

Four modes of formation have been proposed to explain the multiple cut and fill history of the Tuna-Flounder Channel fill. The first mode of formation involved multiple erosional events within a restricted marine environment (James and Evans, 1971). The second mode involved erosion and deposition by submarine incision due to uplift and aerial exposure (Brown, 1985), followed by filling of the incised valley complex through fluvial to submarine channel processes. The last mode involves erosion and deposition resulting from a eustatic sea level fall (Partridge, 1999).

Palaeoenvironments interpreted for the Flounder Formation show a transgressive event, from fluvial–estuarine at the channel base to fluvial–floodplain and shallow marine at the channel top. Nine sequence boundaries within the Tuna-Flounder Channel have been recognised (Ross, 2004). The thickness and distribution of each of these sequences is significant. For example, the oldest sequences are located on the flanks of the present-day channel depocentre, inferring that the thickest part of the channel, as preserved today, overlies an area that initially had limited or no deposition. An explanation could be that the present-day channel-axis depocentre was initially a structural high that later subsided. Furthermore, the channel fill distribution has been useful for imparting structural reasons to why fault tips arrest at the base of the channel (defined by the Mackerel Unconformity, MackMS horizon), as opposed to the overlying Marlin Unconformity (MarlSS horizon). That most faults arrest at the base of the Tuna-Flounder Channel shows that this faulting predated deposition of the channel fill. The observation made by Ross (2004) would be consistent with the claim in this study that uplift and erosion, and/or inversion, predated a margin subsidence
sag phase. Furthermore, the reverse faults and compressional basin phase and, approximately 1–2 km of pre-Tuna-Flounder Channel sediments estimated to have been removed in the study area would also have had to postdate the normal faulting observed. Similarly, the uplift and erosion would have had to postdate the thickening of the UHalSS1 and UHalSS2 intervals observed in the eastern Central Deep.

The Marlin Unconformity occurs post-Tuna-Flounder Channel fill, but its origin is uncertain; the unconformity has previously been interpreted as an erosional end-phase of the upper Halibut Subgroup reservoirs, including erosion of local sands in the Flounder Formation (Partridge, 1999). Most of this erosion occurred pre-Marlin Unconformity; also, only minimal extension occurred post-Mackerel Unconformity. Partridge (1999) further interprets the Marlin Channel to have been downcut as a consequence of two to three sea-level falls, which implies some eustatic control to sediment deposition. The channel is infilled by the Turrum Formation.

It is proposed that the hiatuses and associated unconformities could be representative of intermittent structural movements linked to the rift-drift and rift-inversion basin subphases, as opposed to reflecting depositional boundaries. The Marlin Unconformity is coincident with the last phase of erosion that followed short intermittent periods of compression, uplift and erosion and possibly, localised extension. The Flounder Formation section acts as a seal but is known to be locally sandy. The arresting tips of the Mack faults clearly point to an extensional structural phase pre-dating the channel fill. These faults do not propagate through the basal shales of the Flounder Formation, which reduces the chance of fault breaching for traps formed by reservoir subcrop against the Mackerel Unconformity.

### 3.7.2 Latrobe and Mackerel Unconformities

The Latrobe Unconformity is currently viewed as the most prominent tectonic boundary in the basin (Partridge, 1999; Holdgate et al., 2003). However, the evidence portrayed in this study would suggest that the Mackerel Unconformity is in fact the most prominent tectonic boundary. The fundamental question becomes: ‘Is the evidence presented here robust enough to ascertain that the most prominent tectonic boundary in the basin has been misrepresented in time by 16–22 Ma?’ If so, fault tips arresting at different megasequence boundaries also markedly change the perspective on trap formation and risk associated with fault leakage in CO₂ SLs. Additionally, fundamental changes would result regarding maturation modelling of source-rocks, the latter irrelevant to this study.

Several key geological events and/or transitions immediately precede or are near-coincident with the formation of the Latrobe Unconformity (Figure 3.23), as follows:
• the Bassian Rise cools in the Early to Middle Eocene, c. 52–45 Ma (Chiupka, 1996);

• the slow to fast seafloor-spreading transition in the Great Australian Bight occurs during the Middle Eocene, c. 45 Ma (Tikku and Cande, 1999);

• there is localised seafloor spreading, involving a micro-plate east of Tasmania, during the Late Eocene, c. 36 Ma (McGowran et al., 1997; Norvick, 2005);

• the Tasman Gateway opens up in the Early Oligocene, c. 30 Ma (Exon et al., 1999);

• maximum inland flooding has been demonstrated in the onshore Latrobe sequence during the Late Oligocene, c. 30–23 Ma (Holdgate and Sluiter, 1991), and;

• a compressional regime in the Oligocene was demonstrated (Young et al., 1991; Davidson, 1995) with dates later refined (Johnstone et al., 2001; Power, 2003).

It is difficult to attribute which, if any, of the above events may have initiated and/or directly contributed to the development of the Latrobe Unconformity. For example, the unconformity was formed after the slow to fast seafloor-spreading rate change occurred in the Great Australian Bight, in the Middle Eocene (Tikku and Cande, 1999). Holdgate et al. (2003a) places the Latrobe Unconformity below the Turrum Formation, while Partridge (1999) places the unconformity above it. Partridge (1999) acknowledges that the numerous unconformities and associated surfaces at, and adjacent to the Latrobe Unconformity, are imprecisely and poorly defined. Further, many unconformities merge at the basin edge (Holdgate et al., 2003a) making age-dating of unconformable surfaces potentially inaccurate.

The Latrobe Unconformity is taken in this study as the boundary between the marly muds of the Swordfish Formation and the underlying clastic sediments of the Gurnard Formation. The age is estimated as Late Eocene to Early Oligocene, c. 36–30 Ma, and consistent with Partridge (1999). Therefore, the unconformity coincides with a change in sedimentation from clastic to carbonate dominated (Bernecker et al., 1997). This change occurs after the opening of the Tasman Gateway during the Early Oligocene (c. 30 Ma) and, concomitant crustal stretching between the Antarctic plate and the Tasmanian mainland during the Late Eocene, c. 36 Ma (Norvick et al., 2001; Norvick, 2005). However, faults arrest at the Mackerel Unconformity, proving that the Latrobe Unconformity is not the prominent tectonic unconformity in the basin as had been previously proposed; for example, ‘uplift and erosion associated with this structuring led to the formation of the Top Latrobe Group Unconformity’ (Holdgate et al. 2003a, p150), and ‘these reflectors underlie the Tuna-Flounder
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3.7.3 Faults (Latrobe and Seaspray Groups)

It is possible that clusters of minor faults within the Seaspray Group connect with, or link to, Mack and Latrobe faults at locations where the SworS interval is thin. Such linkage, if any, could be a significant seal bypass risk to CO2 SLs.

Early seismic studies interpret no kinematic linkage between the underlying Mack faults and the faults observed in the Seaspray Group (Feary and Loutit, 1998; Holdgate et al., 2000). However, these interpretations are 2-D-based. Also, the origin of the Seaspray Group faults was ascribed to a compaction process involving syneresis (i.e. clay compaction through dewatering).

A later study tested whether these faults were related to a tectonic process (Power, 2003), or as the previous authors state, originating from a compaction process. The Seaspray Group faults were demonstrated to tip out above the Latrobe Unconformity. Fault throws were found to be on the order of 15 m near the Turrum Field, but up to 75 m near the Bream Field (Power, 2003). Further, the fault spacing is in the range of 50–500 m adjacent to the Turrum Field and corresponding fault lengths in the range 30–1,700 m. Furthermore, 305 fault trace-lengths and azimuths were plotted on rose diagrams, as well as against each-other. It was found that the spread of the fault trace azimuth was significantly more random within this fault set than for the underlying Mack faults interpreted in this study; where, the range of the azimuth and dip was narrow. Specifically, the majority of the fault strikes (Seaspray Group) were found to be nonsystematically scattered between 045° and 180° N (Power, 2003), a result that neither proves or disproves a tectonic origin for these faults. Also, and as confirmed in this study, the underlying Marlin Channel and channel edges have a NW–SE trend, which could influence the localisation and orientation of the Seaspray Group faults. However, no readily observable kinematic linkage to underlying Mack faults is seen on the 3-D seismic slices and sections interpreted in this study. Ultimately, the origin of the Seaspray Group faults remains
uncertain, and a syneresis origin cannot be ruled out as the origin of similar age faults has been ascribed to syneresis in the adjacent Bass Basin (Cummings et al., 2004).

Previous work illustrated that the highest density of Seaspray Group faults are located along the main petroleum migration pathways within the drainage cells of the Turrum and Marlin Fields (Power, 2003). This study has confirmed these earlier observations and, in addition, has also recognised the same pattern in the drainage cell of the Longtom Field. I conclude that the clusters of minor faults interpreted in the Seaspray Group are unlikely to pose a significant fault seal risk.

### 3.7.4 Fault zones

Fault zones interpreted in this study include: (1) the SRFZ, a zone located on the downthrown side of the North Rosedale Fault; (2) an area of intersecting fault trends intersecting and located on the western limb of the Flounder Field; and, (3) an area west of the Tuna Field paralleling faults F91 and F165. Also, nearly all of the fault zones interpreted in this study coincide with locations of inversion structures (Flounder and Tuna–West Tuna Fields). Prior to this study, no fault zones have been implied within the study area. However, since fault zones can affect the overall likelihood of trap breach, it is important to discuss the overall consensus of their presence.

Fault zone widths can be estimated from fault displacement measurements; these, when cross-plotted on a log–log scale demonstrate a linear relationship (Walsh et al., 1998; Bailey et al., 2005; Wibberley et al., 2008). These authors estimated fault-zone widths from fault displacement relationships for high porosity sandstones, cataclasites, clay-rich zones, phyllosillicate basement, and fractured zones around fault relays and in sandstone damage-zones. In this study, maximum fault displacement (throw) estimates for Mack faults are in the range of 50–200 m. Other studies confirmed maximum throws in the range of 150–250 m along the Rosedale Fault and, 50–150 m under the broad terrace (Power et al., 2003). Using the fault-displacement to fault-zone-width relationship, as used by Wibberley et al. (2008), a spread of fault throws in the range of 50–200 m would translate to fault zone widths estimated to be in the range of 0.1–30 m. Such estimates are well below the seismic resolution of ~ 200 m, as demonstrated from the Fresnel Zone radii (see Figure 3.5b). It follows that the interpreted fault zones cannot be confirmed by this methodology and, as far as the author is aware, there is no other method to cross-check the validity of the interpreted fault zones.
3.7.5 Fault regime and mode (Palaeocene to Early Eocene)

Maastrichtian to Early Eocene hiatuses and associated unconformities ILHalS, LHalSS, IUHalS and MackMS (Halibut Subgroup) were recognised in this study. Note that Mack faults dominate within the associated upper Halibut Subgroup section. The Mack faults result from a late Palaeocene to Early Eocene extensional regime, which is well recognised both within and outside of the study area. Within the study area, a compressional regime has been recognised from structures in the Tuna, West Tuna and Flounder Fields, and a strike-slip mode has also been postulated along and adjacent to the Rosedale Fault and in the Flounder Field. Fault-blocks trend WNW–ESE across the broad terrace with the fault network across fault-block complexes being less dominant. Pertinent questions that arise in relation to CO₂ storage are threefold.

(i) Do the hiatuses and associated unconformities, fault modes and where fault tips arrest aid in establishing a robust chronology of faulting events for the latter part of the Halibut Subgroup and, prior to the infilling of the Tuna-Flounder Channel? Are the fault modes and fault events short-lived, prolonged, synchronous, or nonsynchronous?

(ii) Is a transpressional fault mode an alternative to the combined strike-slip–reverse fault modes interpreted?

(iii) Can other studies corroborate, or not, the fault network across fault-block complexes as interpreted in this study?

The regional tectonic driver associated with faulting in the Early Eocene (Mack faults, Figure 3.23) has previously been associated with cessation of seafloor spreading in the Tasman Basin (Power, 2003; Power et al., 2003). Early Eocene faulting clearly postdates the rift-stage associated with breakup of the Eastern (Gaina et al., 1998) and Southern Margins (Sayers et al., 2001) and predates the transition from slow to fast seafloor spreading along the Southern Margins in the mid Eocene (Veevers, 1986; Tikku and Cande, 1999). However, it is not immediately obvious how to explain the dominant trend of the Mack faults (this study), which is near-orthogonal to the seafloor spreading axis in the Tasman Basin (Gaina et al., 1998). Additionally, why should regional-scale tectonic plate motions in the Early Eocene impose such a dominant trend in the Gippsland Basin, but be less pronounced in the adjacent Bass Basin (Cummings et al., 2004). Crustal thinning under the Bass Strait is unequivocal, as demonstrated from long offset seismic data (Collins et al., 1992). Similarly, cessation of faulting is linked to failure of renewed crustal stretching between Antarctica and Australia, as previously proposed (Hill et al., 1995). Thus, it is postulated in this study that the Mack
faults may have been initiated by crustal stretching between Antarctica and Australia, with crustal stretching on the eastern margins only playing a secondary role.

Fault-controlled basin subsidence has been interpreted to be active until the Late Maastrichtian where growth was demonstrated on a few faults (Power, 2003). Fault spacing reduces from 3–27 km, in the Santonian to Campanian, to 2–10 km by the end of the Maastrichtian (Power, 2003). Similarly, maximum fault throws reduce in magnitude to 40–240 m. Faulting occurs in the Maastrichtian, during deposition of the lower Halibut Subgroup sediments, but wanes in the Palaeocene, during deposition of the upper Halibut Subgroup sediments (Johnstone et al., 2001; Power, 2003). Maximum fault throws that were 40–240 m in the Maastrichtian reduced to 40–135 m by the Palaeocene. Power (2003) states that the Latrobe Unconformity is coincident with the cessation of fault-controlled subsidence; however, within this study area, only five faults were interpreted to offset the unconformity with throws of the order of only 30–40 m.

Strike-slip and/or transpressional tectonic models have been invoked to explain the repeated inversion that occurs in the Gippsland Basin from the Late Cretaceous to recent (Duff et al., 1991; Etheridge et al., 1991; Willcox et al., 1992). These models are based on regional observations and 2-D seismic data. Many of these models invoke plate tectonics as first order drivers but fail to consider second order drivers to explain the unique structural features of the study area. In the eastern Gippsland Basin, flower structures have been identified in the Tuna and West Tuna Fields and ascribed as originating from an oblique extensional regime (Power et al., 2003); but ‘with no requirement to invoke major strike-slip kinematics along the Blue West Fault’ (Power, 2003, p248), further and because there is an ‘absence of definitive kinematic indicators such as fault striations, it is difficult to ascertain the role of strike-slip kinematics in the evolution of the Tuna area’ (Power, 2003, p249). These observations and statements are consistent with the conclusions of this study, being that the strike-slip faults interpreted have short-offsets and, only caused a low level of overprinting of other faults. The probable intermittent and minor strike-slip fault movements proposed for the RFS and associated with the SRFZ, are consistent with the present-day state-of-stress measurements that estimate a borderline reverse–strike-slip fault regime (Nelson et al., 2006b). On the basis on the above, it is possible that a transpressional regime existed rather than a compressional regime.

Subsidence and flexural sag may have caused some inversion structures to be less evident, or their importance masked or misinterpreted (Figure 3.22). It is also possible that slip on some faults could

21 See Figure 3.12a for location.
have occurred across hinge zones because of margin sag and/or associated flexure; in such a case, the slip would not necessarily be linked to extensional or compressional events at the basin scale. Localised Early Eocene inversion structures are seen as originating from distinct tectonic events whose associated regime was active across the study area and basin. The existence of multiple hiatuses (ILHalS, LHalSS, IUHalS and MackMS) as interpreted in this study, as well as episodic fault movements demonstrated elsewhere in the Gippsland Basin (Johnstone et al., 2001; Power, 2003) could imply that the study area was subjected to several brief intermittent basin-wide tectonic movements/regimes during the Maastrichtian and Early Eocene. Indirect evidence for episodic fault movements comes from thermotectonic studies of the Bassian Rise that demonstrated two periods of cooling (O'Sullivan et al., 2000), c. 94 ± 2 Ma (Cenomanian), with the second at c. 65–45 Ma, both closely relating to the argument. Similarly, tectonic studies of onshore Victoria also demonstrated episodic movements (O'Sullivan et al., 2002).

The fault network across fault-block complexes is not discussed specifically in the literature but the principal stress direction ($\sigma_1$) has been demonstrated to have rotated clockwise by 60° from the Cenomanian (pre-Otway Unconformity) to the Early Eocene (Mackerel Unconformity; Power, 2003; Power et al., 2003). In this study, the dominant NW–SE trending Mack faults have been shown to cross-cut the pre-existing trends of Otwa and Seah faults as interpreted by Power (2003; compare Figures 2.4a–b with c–d). The Mack faults were, however, interpreted to have originated during the Maastrichtian, to have waned during the Palaeocene and to have again been reactivated in the Early Eocene. On the basis of the above, the above fault chronology would imply a significant degree of disassociation in the kinematic linkage between the Seah and Mack faults although further work would be needed to prove the degree of the disassociation.

### 3.7.6 Basin phases/subphases

About 20% of faults arrest at the Seahorse Unconformity (Campanian hiatus, SeahSS horizon), many within the RFS, while 67% of faults arrest at the Mackerel Unconformity (MackMS horizon). Also, the underlying Golden Beach Subgroup section is related to a transition-rift basin subphase (2B) while the overlying Halibut Subgroup section is related to fault-controlled rift-drift and inversion basin phases (2C to 2G). The geological uniqueness of different basin subphases and associated sections may impact on the inherent viability regarding their CO$_2$ storage potential as a result of differing fault sets present. Thus, it is pertinent to question how their CO$_2$ storage potential would compare.

The Golden Beach and Halibut Subgroups differ in that the former has been subjected to several phases of faulting.
(i) In the Golden Beach Subgroup, several fault episodes are reflected in the clockwise rotation (15–30°) of the maximum principal stress (σ1) from the Cenomanian onwards (Power, 2003). Fault trends have changed from an ENE–WSW to E–W azimuth during deposition of the Emperor Subgroup and from an E–W to NW–SE azimuth during deposition of the Golden Beach Subgroup sediments. The above are outlined in the following quotes: ‘faults that accommodated growth of the Emperor Subgroup were subsequently abandoned and cross-cut by the closely spaced planar normal faults that dominate the east Central Deep, with the exception of long-lived zones such as the Rosedale, Darriman and Foster Fault Systems’ (Power, 2003, p184) and ‘the Golden Beach Subgroup is the oldest sedimentary unit to display evidence of significant growth across a number of the principal planar E–W to NW–SE trending normal faults that characterise the structural style of the Central Deep’ (Power, 2003, p185).

(ii) The fault orientation (NW–SE azimuth) was maintained during deposition of the Halibut Subgroup sediments. The impact of the rotation and fault realignment are outlined in the following quote ‘It also shows that many new faults formed, breaking up the underlying pre-existing fault geometry into more closely spaced fault blocks, from dimensions of 3–27 km wide in the Santonian to Campanian to 2–10 km by the end of the Maastrichtian’ (Power, 2003, p187).

(iii) The presence of the Seahorse Unconformity, arresting Seah faults and the volcanics overlying the Chimaera Formation, as found in this study, are all consistent with a tectonic boundary driven by crustal stretching and seafloor spreading. The structural geometry changes from rotated fault blocks with half-graben fill during the Campanian to fault blocks with graben fill and separated by relay ramps during the Palaeocene to Early Eocene.

The implication of the above is that transition-rift basin subphase 2B and section (Golden Beach Subgroup) is more intensively faulted then the overlying fault-controlled rift-drift and inversion basin subphases 2C to 2G and section (Halibut Subgroup). A greater variation of fault-trace azimuths was demonstrated in this study and it follows that CO2 storage can be expected to be affected by pressure and/or reservoir continuity issues caused by the more intense fault compartmentalisation that would be expected. In comparison, the fault network across fault-block complexes of the Halibut Subgroup is reduced. These considerations have significant implications for regional assessments regarding CO2 storage, where distinct basin subphases and sections would be inherently superior to other subphases and/or sections from a purely reservoir-continuity risk.
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3.7.7 Geohistory (Late Eocene to Middle Miocene)

Late Eocene to Middle Miocene hiatuses at c. 35–32 Ma, 17.6–16.4 Ma and 14.5–13.9 Ma link to unconformities LatrSS, SworSS and CongSS, respectively. It is important to ascertain: (a) what the unconformities and the upper tip-line bound of any faults tell us about structural events that occurred during the sediment deposition associated with the seal intervals (as represented by the Cobia Subgroup, Swordfish and Conger Formations), and (b) question any potentially, associated seal bypass risk via faults.

A number of geological events occur during the combined time span, c. 35–13.9 Ma (Figure 3.23), as follows.

(i) A suite of Oligocene to Early Miocene (c. 30–16 Ma) third-order depositional systems and associated eustatic cycles has been interpreted on the Southern Margins (McGowran et al., 1997). Cessation of the second of the cycles (~ 14 Ma) coincides with the CongSS hiatus, at c. 14.5–13.9 Ma.

(ii) Significant volcanism is present onshore, commencing in the Late Oligocene, c. 27 Ma (Chiupka, 1996). Also, uplift has also been demonstrated in the highlands, at c. 25 Ma (Norvick et al., 2001; Norvick, 2005).

(iii) The collision of the Australian plate with the Southeastern Asia plates commences in the Early Miocene, at c. 25 Ma (Norvick, 2005).

(iv) Early Middle to early-Late Miocene compression (c. 16.5–10.5 Ma) has been demonstrated across the Northern Strzelecki Terrace and Northern Platform (Johnstone et al., 2001); later revised to Middle to Late Miocene, at c. 16.5–6 Ma (Power, 2003).

(v) Depositional rates increase above the Swordfish unconformity, c. 17.6–16.4 Ma, before reaching maximum rates during deposition of the Conger and Cod Formations, as demonstrated in this study.

(vi) The base and top of mid-Miocene channeling (Figure 2.3; Norvick et al., 2001) coincides with the Swordfish unconformity (SworSS hiatus) and Bullseye Karst (BullSS hiatus) identified in this study. Also, compaction-driven faulting has also been interpreted as occurring within the basal Seaspray-Group section (Feary and Loutit, 1998; Holdgate et al., 2000).

Both the SworSS and CongSS hiatuses are well defined by palynological zone offsets as interpreted from age-depth plots; and, increased rates of deposition are observed above both hiatuses. Additionally, the Conger unconformity (CongSS hiatus) coincides with the major lithological
changeover from mixed clastic-carbonate to carbonate lithologies. On this evidence alone, uncertainty remains as to whether the boundaries are eustatically or tectonically driven. In addition, the nonfault-controlled nature of the basin sag (subphases 3B–3D) within the Seaspray Group implies that the impact of extension, if any, is minimal. However, flexural effects resulting from margin subsidence and basin sag, along and to the south of the RFS, are interpreted to have caused localised slip on a few faults (e.g. F55, F56, F64 and F83). In addition, the fault clusters within the SworSS and CongSS intervals, that overly the SRFZ and associated structural hinge-lines, are interpreted to result from intermittent movements along the RFS during the Miocene.

From all the preceding observations and, when taking into account inversion structures present onshore (Dickinson et al., 2001; Dickinson et al., 2002) and overlying the Northern Strzelecki Terrace and Northern Platform (Johnstone et al., 2001; Power, 2003), it appears that the SworSS and CongSS hiatuses and associated unconformities may have a tectonic origin. However, there is no obvious evidence of compression in the main syn-sedimentary section of the Seaspray Group (Oligocene–Miocene); and fault reactivation south of the Northern Strzelecki Terrace is minimal. This last point suggests that the seal-bypass risk in the Halibut Subgroup reservoirs being considered here is small.

### 3.7.8 Geohistory (Middle Miocene to Pleistocene)

Middle Miocene to Pliocene hiatuses and associated unconformities include ICodSS (c. 13–12 Ma), BullSS (c. 11.7–10.3 Ma) and IWhitS (c. 5.8–4.9 Ma). It is important to ascertain what the hiatuses/unconformities, in combination with the upper tip-line bound and regional-scale compressional events, tell us about fault reactivation of Latrobe Group faults (Latr). Again as in the preceding section, fault reactivation is important in order to question any potential associated seal-bypass risk via faults.

Fault reactivation and compression in the post Mid Miocene is demonstrated in the study area by reverse faults and associated folding present in the Longtom Field structure—faults and folds have not been interpreted south of the SRFZ. At a regional scale, post Mid Miocene structures have been recognised across the Northern Strzelecki Terrace and Northern Platform, where structures subparallel the RFS (Power, 2003; also see Figure 2.5). Onshore, McGowran et al. (1997) recognised erosional surfaces and associated these to the compressional regime, but did not describe individual structures.

Further, a major hiatus (i.e. 10–5.5 Ma) was described (Dickinson et al., 2001; Dickinson et al., 2002) as the Moorabool Unconformity where this low angular unconformity separates folded Miocene
carbonate sediments from overlying silicilastic sediments; it has indirectly been dated at c. 10 Ma (Holdgate et al., 2004). This hiatus correlates with the Bullseye Karst (BullSS horizon), dated at c. 10.5 Ma (see Figure 2.2; Partridge, 1999), which in turn coincides with a global eustatic fall at c. 10.5 Ma (Haq et al., 1987). The systems tract architecture also changes from aggradational to progradational across the BullSS horizon, which implies some degree of eustatic control to the boundary (see Figure 2.3; Feary and Loutit, 1998). According to evidence, the Bullseye Karst is partly tectonic in origin, but also resultant from a drop in sea level (Partridge, 1999); although, it has been found in this study to not contribute to understanding fault reactivation of Latrobe Group faults.

Volcanics could also play a role in forming the post-Mid Middle Miocene to Pliocene BullSS and IWhitS hiatuses and associated unconformities, although largely indirect through thermal-based uplift. Outpourings of the Newer Basalts in central Victoria were initiated in the Late Miocene at c. 7.2 Ma (Holdgate et al., 2004) and, continued into the Quaternary (Birch, 2003). This volcanism correlates with other extensive volcanism seen in onshore NSW and QLD, as well as in the northern Tasman Sea and offshore Middleton Basin (Norvick, 2005). However, any link to Gippsland Basin tectonics is uncertain.

The Pliocene hiatus IWhitS (c. 5.8–4.9 Ma) coincides with a polyphase compressional event, which was initiated during the Late Miocene (Johnstone et al., 2001; Power, 2003; Power et al., 2003) and was intermittent in the Pliocene. However, no significant faulting or folding of Pliocene age has been observed within the UWhiSS interval associated to the study area.

Onshore, retrograding clastic shoreline barrier features present in inland Victoria indicate an overall regressive coastal onlap pattern occurring across the Pliocene to Pleistocene boundary (Sandiford, 2007). River channels incised under the offshore shelf have been imaged using the magnetic signature of the channel fill, suggestive of onshore uplift to initiate the channel load (Holdgate et al., 2003b). Onshore, uplift and folding occurred in the Middle Pleistocene, at c. 1 Ma. Further uplift is known to have occurred at 0.2 Ma, with simultaneous reorientation of shoreline strands (Wallace et al., 2005). Pliocene–Pleistocene compression has been corroborated by inversion structures observed at a regional scale across the Southern Margins (Hillis et al., 2008).

Summarising, the post-Mid Miocene compression observed onshore most likely extends out to the Northern Platform and Northern Strzelecki Terrace, with marginal to no effects observed in the main depocentre (eastern Central Deep). The onshore late stage Pliocene–Pleistocene compression phase might be expected to have a corresponding offshore equivalent, but none is observed within the 3-D seismic grid of the study area. However, the age-depth plots available cannot be interpreted for section younger than 3.2 Ma; thus, potentially obscuring subtle hiatus-based evidence.
3.8 Conclusions

The stratigraphic intervals potentially favourable for CO₂ storage, in order of priority, and based solely on the fault network estimated across the fault-block complexes are the upper Halibut Subgroup (UHalSS1 and UHalSS2 intervals), lower Halibut Subgroup (LHalSS1 and LHalSS2 intervals) and Golden Beach Subgroup (GBeaSS interval). Commensurately, the strike distribution of the Mack faults is clustered within a range of 115° ± 60°N indicating that fault-block complexes incorporating the Halibut Subgroup are non-compartmentalized based on strike distributions alone. The strike distribution of all Seah faults intersecting the Golden Beach Subgroup is unclustered, with strikes varying in azimuth from 20° to 160°N. In this case, the fault network across fault-block complexes is predicted to be increased and comparatively poorer fault compartment pressure communication is predicted for intra-Golden Beach Subgroup reservoirs. Clustered strike distributions are favoured, as the associated openness of the fault-block complexes is more conducive for pressure communication within a reservoir interval. Thus, CO₂ injection is expected to be more viable in the Halibut Subgroup section associated with the rift-drift and rift-inversion subphases (2C–2G) than, in the Golden Beach Subgroup section associated with the syn-rift and transition-rift subphases (2A–2B).

The majority of faults have been shown to arrest at either the Otway (OtwaMS horizon), Seahorse (SeahSS horizon), or Mackerel (MackMS horizon) Unconformities, reflecting three major faulting events. The Mack faults dominate the structural grain of the study area and trend WNW–ESE. The majority of these faults are down to the basin, domino-style fault sets. Hiatuses and accompanying unconformities ILHalS, LHalSS and IUHalS are interpreted to be due to short-lived episodic tectonic movements within the upper Latrobe Group and active during the rift-drift basin subphases (2C–2E).

Seal integrity issues will be associated with fault zones and fault seal analysis will be required when these zones form part of a potential CO₂ storage site. Four fault zones are interpreted, as follows.

(i) The SRFZ incorporates numerous fault blocks, is regional in extent and up to 5 km wide.

(ii) A second fault zone, located adjacent and parallel to the E–W trending Rosedale Fault (F55), extends from the Kipper Field to the Tuna Field. The fault zone incorporates multibranch splay faults that form en echelon fault patterns; both short-offset strike-slip and normal faults are interpreted.

(iii) A third fault zone, located downdip of the Rosedale Fault and associated with NE–SW trending reverse faults (west Tuna fault family), bound the western limbs of the Tuna and West Tuna Fields.
(iv) A fourth fault zone, located at the western edge of the Flounder Field, where NE–SW-trending reverse faults (west Flounder fault family) intersect WNW–ESE-trending faults that form a flower structure (east Flounder fault family).

The thickness of the Tuna-Flounder (450 m) and Marlin (250 m) Channel fills enabled some new interpretations to be made in this study. For example, the thickness of the sedimentary fill allowed the chronology for faults to be differentiated by distinguishing those arresting at or propagating through, the Mackerel, Marlin and/or Latrobe Unconformity surfaces. The seismic interpretation suggests that the majority of fault tips arrest at the Mackerel Unconformity, indicating that the Latrobe Unconformity (Late Eocene–Oligocene, c. 36–30 Ma) is not the dominant tectonic basin-bounding megasequence boundary, but is secondary in importance to the main tectonic Mackerel Unconformity (Early Eocene, c. 52 Ma), as established in this study. Thus, the main faulting event within this part of the Gippsland Basin is shifted backwards by c. 16–22 My. In the context of this study, the observation that most fault tips arrest at the Mackerel Unconformity provides greater assurance for trap integrity as fluid flow via faults is less likely if faults do not propagate into the overlying seal interval. Also important but not pertinent to this study, the observation impacts on the thermal maturation modelling of petroleum source rocks where heat-flow decay would also be shifted by approximately 16–22 My backwards in time.

The three local seals are the Flounder Formation, Cobia Subgroup and the Swordfish Formation, all mainly comprised of thin-to-condensed section. Low depositional rates are characteristic of the Early Eocene to Early Miocene (c. 52–16.5 Ma). Exceptions include the Early Eocene (c. 52–49.5 Ma) Tuna-Flounder and Marlin Channel fills and, Late Eocene to Early Miocene (c. 36–16.5 Ma) Swordfish channel fill. Depositional rates increase in the Middle Miocene from the CongSS hiatus (c. 13 Ma) onwards and carbonate deposition is more prominent. The combined thickness of the three seal intervals is of the order of 100 m or less, except within the Tuna-Flounder, Marlin and Swordfish channels where the channel fills thicken up to 450, 280 and 650 m, respectively. The local seal intervals can be sandy in places, so that both the top seal and fault-seal integrity would require interpretation. However, the 2.2 km of dominantly carbonate section overlies the three seals; this forms part of the Seaspray Group and is known to act as a highly competent regional seal.

Miocene to Pliocene reactivation of Early Eocene Mack faults is not commonly observed south of the Rosedale Fault, although known to occur across the Northern Strzelecki Terrace and Northern Platform (Power, 2003; Power et al., 2003). Some stand-alone Mack faults within the broad terrace area could have been reactivated or, alternatively, be linked to flexural effects across hinge lines. Localised clusters of minor faults occur in the Swordfish and Conger Formations; the absence of any
kinematic linkage between the fault clusters and the Mack faults reduces the chance of breach-of-trap above the Mackerel Unconformity. The integrity of the top seal for potential CO₂ storage within the upper Halibut Subgroup is concluded to be favourable when solely considering structure-based screening criteria and other criteria remain to be evaluated.
CHAPTER 4—DEPTH CONVERSION

4.1 Introduction

The 3-D seismic dataset was interpreted in the time domain across the study area; thus, enabling
gross inconsistencies in the structure and seismic horizons to be readily corrected if required.
However, the fault seal analysis to be carried out in the study subarea is best done by importing a 3-
D depth cube; this enables seismic sections to be visualised when addressing minor inconsistencies
in the structural interpretation. Thus, the depth conversion required in this study must incorporate
depth conversion of both the seismic horizons and faults (ASCII), but also the SEGY data cube.

Robein (2003) summarised 11 depth conversion methods, each method suited to differing objectives
and/or velocity data constraints. For this study, the two methods initially considered were single and
multi-polynomial functions; both estimated by trend-fitting checkshot data from multiple wells. The
first method uses a function that is trend-fitted to the checkshot data for all wells, a method that is
suited to a narrow datapoint envelope (Lyon et al., 2004). However, plotting the checkshots for the
69 wells used in the depth conversion showed a datapoint envelope with a variation in absolute
depth of the order of ± 250 m (Appendix 2: Figure A2.12), an error bar considered too large for
determining reservoir offsets across-fault. The second method uses multi-well polynomial functions;
thus, enabling time–depth variation between individual well locations to be accounted for. In addition,
in the case of the study area, the water depth varies in the order of 100+ m and, lateral velocity
variations occur in the Mid-Miocene channels of the Seaspray Group (Holdgate et al., 2000; Wallace
et al., 2002), as well as in the Tuna-Flounder and Marlin Channels of the Latrobe Group (Power,
2003; Gross et al., 2008). The multi-well polynomial-functions method allows pseudo-well locations
to be created and, polynomial functions to be fitted to stacking velocity data—the method was
consequently chosen for the depth conversion.

Most seismic interpretation workstations have software that can depth-convert ASCII fault and
horizon data (e.g. Schlumberger’s InDepth™ program). Unfortunately, software capable of depth-
converting SEGY data is publicly unavailable. Petroleum exploration companies usually tender-out
the time to depth conversion of SEGY data to seismic processing companies. An alternative was
sought here since funds were unavailable to tender-out the task.

Chapter 4 describes the optimal workflow, testing, enhancements and the error bar associated with
the depth-conversion software Witosoft® (program Depth converter) that was developed as part of
this study. Lastly, Appendix 2 (Sections A2.3–2.5) contains the checkshot data used, a description of the software frontends, depth error analysis and the input file incorporating all polynomial functions.

4.2 Methodology and context

Schlumberger’s InDepth™ program was considered for carrying out the depth-conversion but was found not to have the functionality to depth-convert SEGY data. An alternative was to extend the capability of the depth conversion Witosoftc software, from 2-D to 3-D seismic datasets and, incorporate both SEGY and ASCII data (P. Boult—Pers. Comm., Aug. 2008). Witold Seweryn of the Department of Primary Industries and Resources of South Australia (PIRSA) was subsequently contacted to assess the feasibility of extending the capability of the 2-D version of the software to 3-D and, a collaborative working arrangement was formed. All programming enhancements were made between Oct. 2008 and Feb. 2009 by Witold Seweryn; the programming language was C++. The software was initially test-run on a pilot area and finally applied to the study area. Quality-control checks were carried out and recommendations for software enhancements were forwarded to Witold Seweryn in order to achieve an acceptable error of fit.

The methodology adopted for depth conversion incorporated the use of checkshots to depth-convert the shallower part of the stratigraphic interval of interest (i.e. the Latrobe Group) and, stacking velocity-derived Dix interval velocities (Dix, 1955) to depth-convert below well-TD (Figure 4.1). Further, wells were grouped to provide an average Dix interval velocity for those wells that lacked stacking velocity data (Figure 4.2). Only the fault-segment and horizon interpretations along seed lines were depth converted (Figure 4.3a); the fault boundaries and horizon surfaces were recreated within the TrapTester™ project.

4.3 Results

4.3.1 Software development

The frontend menus to the Witosoftc software were designed to be user-friendly; the workflow can be broadly divided up into four parts, as follows.

(i) The operator chooses the 3-D versus 2-D seismic data options, similarly, the offshore versus onshore options, as well as the data type to be depth-converted (use Segy mpoly for SEGY data and ASCII mpoly for ASCII; Appendix 2: Figure A2.13a). Witosoftc’s input format is compatible with the output format from the seismic workstation and follows SEG standards for SEGY data (Barry et al., 1975).
(ii) An ASCII file containing seafloor TWTs along seismic inlines is then read in. An option also exists for smoothing the seafloor TWTs before depth conversion (Appendix 2: Figure A2.13b).

(iii) The file containing well locations and associated polynomial functions is input; these polynomial functions were derived by trend-fitting the checkshot and stacking velocity data. An option also exists to apply a constant velocity below the well-TD (Appendix 2: Figure A2.13c).

(iv) The program is then run, first to depth-convert the ASCII data, then the SEGY data.

4.3.2 Seafloor interpretation

Reliable seafloor TWTs could not be obtained for all grid points when using GeoFrame™’s horizon autotracker function because of reflector degradation; a result of low fold in shallow water depths.
Figure 4.2. Map showing distribution of checkshot and stacking velocity data—application to depth conversion. See Appendix 1 (Table A1.1) for well-name abbreviations.
(i.e. < 80 m). Varying the quality level, maximum dip and trace parameters did result in some areas being autotracked, but not others. An alternative seafloor autotrack method was sought. Seafloor TWTs were subsequently interpolated in GeoFrame™’s basemap module, using seed lines as control points (Figure 4.3a). Seafloor artefact TWTs occurred in places but these were subsequently minimised using the smoothing radius function available in the Depth converter program (Figure 4.3b). Seafloor TWTs were subsequently interpolated at all grid points (Figure 4.3c) and the resultant point grid was compared against DP IVIC’s seafloor depth map (Figure 3.1). Some seafloor artefacts remain visible in the interpolated TWT point grid but depth errors are small (i.e. < 20 m).

4.3.3 Interpretation—checkshot data

Second- to sixth-order polynomial functions were fitted to the input checkshot data and least squares regression fit coefficients (abbreviated as R²) obtained. The error of fit between polynomial-derived and checkshot depths is particularly sensitive to the number of decimal places in the polynomial coefficients. Here, 10, 12, 16 and 20 decimal places were used for second- or third-, fourth-, fifth- and sixth-order polynomial functions, respectively. Only one sixth-order polynomial function was fitted, at East Pilchard-1. Most polynomial functions ranged from third- to fifth-order (Appendix 2: Table A2.74). In addition, the error of fit was graphed to determine the residual trend and outlier points (Appendix 2: Figures A2.14–2.19).

The polynomial-derived depth, calculated at every sample location W_{x,y,TWT}, is based on an inverse-distance weighted summation of depths calculated from all the polynomial functions provided in the input file (Appendix 2: Table A2.74). For example, well-1, being located closer to a sample location W_{x,y,TWT} will result in the derived depth being weighted more heavily at well-1 than at well-2 located further away (Figure 4.1). For example, if the distance D2 = 2 * D1, then

\[
W_{x,y,TWT} = \frac{W_1}{D1^2} + \frac{W_2}{D2^2} = \frac{4 \times W_1 + W_2}{5} \quad \text{(Eq. 4.1)}
\]

where W1 and W2 represent depths obtained from the polynomial functions at well-1 and well-2, respectively.

4.3.4 Interpretation—stacking velocity data

To better represent the fault plane at depth, stacking velocities were used to depth-convert the SEGY and ASCII data (fault) below the well-TD (Group ‘chk+Dix’—Figure 4.2). Stacking velocity data were used exclusively in three cases, as follows.
(i) Checkshot data were unavailable for 12 of the 69 wells used in the depth conversion. Consequently, for these 12 wells a polynomial function, fitted to Dix interval velocities, was applied from seafloor to base of reflectivity (Polynomial function W1—Figure 4.1; group ‘Dix’—Figure 4.2).

(ii) Checkshot data were sometimes available outside the 3-D seismic survey grid, in areas where stacking velocity data were unavailable. In such cases, wells were grouped and an average interval velocity below well-TD was determined for those wells within the group that had no stacking velocity data (group ‘chk+av.Dix’—Figure 4.2). The wells were grouped according to both their geographical location as well as similarity of their sediment thickness profile, as interpreted from TWT data (see Chapter 3). The resulting average interval velocity is portrayed in the polynomial file input in the program Depth converter (maxvel—Appendix 2: Table A2.74).

(iii) The Tuna-Flounder and Marlin Channel fills are up to 450 m thick and, are known to have significant lateral velocity variations (Power, 2003; Gross et al., 2008), generally high velocity thereby causing pull-up of underlying seismic data (Gross et al., 2008). To adjust for these lateral velocity variations, pseudo-wells were placed in the Tuna-Flounder and Marlin Channels (4 and 2 wells, respectively) and, stacking velocities used to depth-convert at these well locations (group ‘Dix’—Figure 4.2).

The stacking velocity data were loaded into EXCEL spreadsheets. Two stacking velocity analyses on inlines and crosslines that straddle the well location were averaged and adopted for that well. Dix interval velocities were calculated from the stacking velocities and plotted at the mid-TWT point between data pairs. The derivation of Dix interval velocities is sensitive to the stacking velocity input, which in turn can lead to an unreasonable estimation of interval thickness. To compensate for this sensitivity, a 6th order polynomial function was fitted through the Dix interval velocity–TWT datapoints; the water layer velocity was anchored at 1,500 m/s at a TWT intercept of zero (Figure 4.4a). An average interval velocity was subsequently calculated using the 6th order polynomial and anchored at the mid-TWT point between data pairs; smoothed interval thicknesses and cumulative depth curves were then derived (Figure 4.4b). The TWT/depth pairs were then re-datumed from MSL to seafloor. Finally, a linear trend was fitted through these latter data pairs and a constant velocity obtained and subsequently input as ‘maxvel’ in the polynomial file (Appendix 2: Table A2.74). The polynomial-derived depths matched the checkshot data adequately, although some TWT offsets remain because of the fundamental difference in origin of both types of geophysical data (checkshot ~ stacking velocity).
Figure 4.3. Depth-conversion methodology for ASCII data.
(a) Seed-line interval used, TWTs were interpolated to all inline–crossline intersections. (b) Smoothing radius is applied to each data point located within the circumference, and (c) resulting interpolated seabed TWTs. See Appendix A1 (Table A1.1) for well names.
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Figure 4.4. Depth-conversion methodology—example using checkshot data for Scallop-1.
(a) Dix interval velocities, note offset between the checkshot and Dix interval velocity derived depths, the SworSS horizon represents the top of the interval of interest for depth conversion (i.e. top seal). (b) Depth–TWT plot showing checkshots above the well-TD and, the constant velocity fit below it.
4.3.5 Time to depth conversion

The 32 Gb SEGY time domain file (representing the whole study area) sampled at a 4 ms interval, converted to a 52 Gb SEGY depth domain file sampled at a 5 m interval. The 52 Gb SEGY depth domain file then reduced to a 13 Gb SEGY file when loaded in GeoFrame™ and TrapTester™. The reduced size was considered low enough to allow loading the entire SEGY. All fault segments were depth-converted first and then loaded into the GeoFrame™ workstation project; this permitted a quality-control of the depth conversion prior to input into the TrapTester™ project. Similarly, the horizon segments were loaded: Seabed, IWhitS, BullSS, ICodSS, CongSS, SworSS, LatrSS, MarlSS, MackSS, IUHalS, LHaiSS, ILHalS, SeahSS, LongSS, OtwaMS and BaseMS. No mislocation discrepancies were noted for the horizon and fault segments following the depth conversion, as shown by comparing their position relative to seismic reflectors (compare Figures 4.5a and b). For additional quality control, well-markers were checked against the depth-converted interpretation along seed lines at all well locations—no mislocation of seismic horizons was observed. The error analysis and quality of fit are discussed in Section 4.4.

4.3.6 Software runs

Five fundamental changes were collaborated on and made during the programming development period (Dec. 2008–Feb. 2009), as follows.

(i) The run time was reduced by approximately 50%.

(ii) Third- to sixth-order polynomials do not accurately extrapolate depth below well-TD: this drawback is itself an inherent property of higher order polynomials. As a result, the program was adapted to allow the option to input a constant velocity below well-TD (i.e. max time velocity; Appendix 2: Figure A2.13c) and equivalent maximum TWT (i.e. max Time). This constant velocity option was used for 57 of the 69 polynomial functions, while the other 12 polynomial functions did not require the option as these trend-fitted the stacking velocity data from sea level (i.e. not well-TD) to the base of the imported TWT data.

(iii) An early version of the program searched for the closest wells and associated polynomial functions within the octants surrounding each sample location \( W_{x,y,TWT} \), ignoring octants where wells did not exist. Depths were subsequently calculated using a maximum of eight polynomial functions. However, usage of the octant method was found to produce planar artefacts at the octant boundary. This method was consequently replaced by allowing the import of an
Figure 4.5. Depth-conversion results, (a) seismic TWT profile and, (b) seismic depth profile. Basal part of the depth section not shown, inline is 4594.
unlimited number of polynomial functions and applying an inverse-distance weighted summation.

(iv) The seafloor was initially smoothed by averaging the closest TWTs in each of the octants, equivalent to averaging the datapoints on the closest inlines and crosslines, here equivalent to a distance of 28 m. However, seafloor depth anomalies produced undesirable static shifts of both the depth-converted SEGY data and ASCII horizon and fault-segment depths. Therefore, the program was altered to allow the operator to increase the smoothing radius (Figures 4.3a–b). Unfortunately, increasing the smoothing radius increased the program run time. Ultimately, a compromise radius of 100 m was used herein. Lastly, an option to plot the smoothed water bottom grid was added to the program.

(v) Fundamental changes were made to the frontends to improve on redundancy of the shell descriptors.

4.4 Discussion

4.4.1 Error of fit—checkshot data

Second to 6th order polynomials were trend-fitted through the checkshot data (1,670 checkshot pairs) of 57 wells until the error of fit was minimised by increasing the order of the polynomial. The error of fit is equal to the polynomial-calculated depths minus the checkshot-derived depths. The error of fit at different depth levels is calculated (Appendix 2: Figures A2.14–2.19). Outliers ranged from -56 to +55 ms. The absolute depth error, when taking into account all wells, amounted to ±7 m at one standard deviation (1 Stdev) and ±23 m (at 2 Stdev), respectively (Figure 4.6). In these estimates, seafloor depths are not taken into account. The seafloor is the datum from which the polynomial functions are applied; the depth error at the seafloor is estimated at ±15 m. When seafloor depth errors are taken into account, the combined estimation error translates to a root mean square (RMS) error of approximately ±17 m (at 1 Stdev) and ±27 m (at 2 Stdev), respectively.

The source of the RMS error is then compared to other sources of error and/or approximations made, as follows.

(i) The limit of vertical seismic resolution (i.e. ¼ λ) ranges from approximately 7 m, at a depth of 1,500 m, to 19 m at a depth of 4,000 m (estimated from Figure 3.5a).

(ii) The range of depth accuracy needed to determine juxtaposition of reservoir/seal intervals across-fault also depends on the thickness of the reservoir intervals being studied. Herein, the upper Halibut Subgroup is dominantly sandy (see Chapter 6) with reservoir intervals of 100+ m.
not uncommon. It follows that depth conversion of thin, individual sandstone bed intervals is not sought.

(iii) The lower limit of absolute depth resolution is controlled by the depth sample interval (5 m); for this study, the accuracy of the depth-converted SEGY and ASCII data is no better than ± 5 m.

(iv) The minimum acceptable depth error is also dependent on whether we are considering absolute or relative depth. In the case of absolute depth, a ± 27 m depth prognosis error (at 2 Stdev) could translate to wells being potentially drilled outside structural closure. In the case of relative depth, a ± 27 m depth prognosis error translates to a smaller ‘relative’ depth error. However, the depth estimation away from well locations remains uncertain; this is because the depth conversion method used here only takes into account geometric and not true lateral, velocity variation.

The sonic velocities within the Seaspray Group are known to be laterally variable (Holdgate et al., 2000; Wallace et al., 2002), with velocity inversion known to be present above the SworSS horizon in the study area as a result of cut-and-fill channelling (Figure 4.5). Wallace et al. (2002) describe the lateral velocity variation as independent of sedimentary facies but, dependent on the sequence stratigraphy. Mechanical compaction is the primary control on sonic velocity changes during shallow burial, with pressure solution and calcite cementation mechanisms controlling changes at increased burial depths. The presence of these lateral velocity variations, as well as a lower checkshot count within the upper Seaspray Group, directly limits the accuracy of the depth conversion within the group. However, the underlying interval of interest, the Latrobe Group, is not affected at well locations.

Enhancements to the depth conversion used here could be achieved by taking into account checkshot drift correction relative to markers and outputting time-depth after correction. An improved method on the checkshot drift correction would be to construct a comprehensive velocity model with polynomial functions applied to individual layers, similar to what is routinely done when constructing absolute depth maps for the purpose of drilling target prognosis (Robein, 2003). The above enhancements, however, would require increased analysis time that could not be justified in this study, as only relative depth accuracy is sought.

4.4.2 Error of fit—stacking velocity data

Errors in the absolute depth estimation, where sourced from stacking velocities and Dix interval velocity-derived polynomial functions, can be on the order of ± 50 m; as evidenced from overlaying derived depth estimates with checkshot data (Figure 4.4b). The source of the depth error is twofold.
Chapter 4. Depth conversion

First, trend-fitting Dix interval velocities using a polynomial function is only one method to derive interval thicknesses and depths (Robein, 2003). However, the similarity in the interval velocities derived from each well group provides confidence in the reliability of the approach. Second, the closest four stacking velocity analyses were used to represent the stacking velocity versus TWT pairs at the well and pseudo-well location. The four analyses were not inverse-distance weighted, as is often done when seeking accurate layer thicknesses in petroleum exploration (Robein, 2003). However, the depth estimation required below well-TD only pertains to the positioning of the fault plane, which is less critical than the required accuracy over the reservoir interval.

![Figure 4.6. Bar graph showing depth-conversion errors—depth error plotted against checkshot count.](image)

Total number of checkshots is 1,670. The depth error range is -55 to 56 m, ± 7 m (at 1 Stdev) and ± 23 m (at 2 Stdev), respectively. Numbers represent the checkshot count as a percentage, rounded to the closest integer.

4.5 Conclusions

The absolute depth accuracy achieved is on the order of ± 17 m (at 1 Stdev) and ± 27 m (at 2 Stdev). This achieved accuracy implies that the SEGY data, horizons and fault segments will be shifted, marginally, relative to imported well markers. However, the interpretation, initially carried out in the time domain, was tied to the well markers so it follows that any error should be small. Importantly, the relative depth error will be less, on the order of only two to four times the depth sample interval (10–20 m). A more accurate depth conversion could be carried out using a velocity model and combined using multiwell, multilayer polynomial functions. However, the above depth conversion methodology is deemed acceptable with respect to this study’s research requirements.
The capability of the original 2-D version of the depth-conversion software Witosoftc has been expanded to include 3-D seismic interpretations and the SEGY cube. As a result of this study and program development by Witold Seweryn, the program Depth Converter can be obtained through Witold Seweryn’s website (http://users.chariot.net.au/~witek/t2d.htm). The enhanced version of the software Witosoftc is a result of the collaboration with Witold Seweryn that included testing and improving its functionality, as well as improving the programming to reduce run times. This software provides an alternative method for depth conversion using multi-well polynomial functions. It is anticipated that the software may be developed further to include the functionality to average stacking velocities.