THE PRE-CARBONIFEROUS GEOLOGY OF BASS STRAIT AND SURROUNDS

DAVID HUGH MOORE BSc

A thesis submitted for the degree of

Doctor of Philosophy

at Monash University in 2016

Main supervisor: Professor Peter Betts
Associate supervisors: Professor Mike Hall, Doctor Laurent Aillères

School of Earth, Atmosphere and Environment

Copyright notice

Abstract

This thesis examines the Proterozoic and early Paleozoic connections between Victoria and Tasmania. Many different models have been suggested, and I conclude here that the Selwyn Block model is the most appropriate as it satisfies both the geological and geophysical data. The model proposed that the Proterozoic cratonic crust of western Tasmania continues north under Bass Strait and lies unconformably below the Melbourne Zone in central Victoria. The Selwyn Block is the northern end of the Proterozoic micro-continent, VanDieland, which also includes western Tasmania, the west South Tasman Rise and the East Tasman Plateau.

In order to be able to extrapolate the major rock packages in Tasmania across Bass Strait, they first had to be determined in Tasmania. Seven Proterozoic zones were outlined—King Island, Rocky Cape, Burnie, Pedder, Tyennan, Sorell-Badger Head, and Glomar, with an eighth, Eastern Tasmania equivalent to the Paleozoic Tabberabbera Zone in eastern Victoria. Only the first three Proterozoic zones continue across Bass Strait, with the other four truncated either in Bass Strait or lying further south. Outcrops of rocks from the King Island and Burnie Zones are present in windows in Victoria but the Rocky Cape Zone is completely concealed. However, the presence in the mid-crust of the Rocky Cape and King Island zones can be seen in the enclaves and in the geochemical signatures of the Upper Devonian granites of central Victoria and in rare conglomerate clasts.

VanDieland was initiated inside Nuna, between Laurentia and East Antarctica, at about 1.8 Ga. Much of the sedimentation seen in the Rocky Cape Zone is the erosional products of the Grenville Orogeny. As Rodinia broke up, VanDieland began to be extended at about 760 Ma, and this continued until final separation from Antarctica at about 570 Ma. After this, it drifted ‘north’ as micro-continental slivers along the Terra Australis margin until about 530 Ma. It then re-amalgamated in a closing back arc system within the greater Ross-Delamerian Orogeny, although it did not accrete onto Gondwana, remaining perhaps 200 to 300 km outboard. In the Early Devonian, as VanDieland got closer to Gondwana, a Banda Sea-style subduction system retreated southwards outboard of its eastern margin. This accreted VanDieland into Gondwana

and gave rise to a crude, clockwise age distribution of the granites in Tasmania, from approximately 400 Ma on Flinders Island to 350 Ma on King Island.

Declaration

This thesis contains no material which has been accepted for the award of any other degree or diploma at any university or equivalent institution and that, to the best of my knowledge and belief, this thesis contains no material previously published or written by another person, except where due reference is made in the text of the thesis.

Publications during enrolment

I hereby declare that this thesis contains no material which has been accepted for the award of any other degree or diploma at any university or equivalent institution and that, to the best of my knowledge and belief, this thesis contains no material previously published or written by another person, except where due reference is made in the text of the thesis.

This thesis includes three original papers published in peer reviewed journals and one unpublished publication. The core theme of the thesis is is the nature of the pre-Carboniferous rocks under Bass Strait and how they might inform us about the geology of southeastern Australia. The ideas, development and writing up of all the papers in the thesis were the principal responsibility of myself, the candidate, working within the (insert name of academic unit) under the supervision of Professor Peter Betts, Professor Mike Hall and Doctor Laurent Aillères.

In the case of chapters 2 and 3, my contribution to the work involved the following:

Thesis chapter	Publication title	Publication status*	Nature and extent (%) of student's contribution
2	Towards understanding the early Gondwanan margin in southeastern Australia	Published	Did the research guided by supervisors, wrote almost all of the paper; (90%)
3	Fragmented Tasmania: the transition from Rodinia to Gondwana	Published	Did the research guided by supervisors, wrote almost all of the paper; (90%)
Appendix 3	Comment on "Early opening of Australia and Antarctica: new inferences and regional consequences" by Jensen Jacob and Jérôme Dyment	Published	100%

(The inclusion of co-authors reflects the fact that the work came from active collaboration between researchers and acknowledges input into team-based research.)

I have renumbered sections of submitted or published papers in order to generate a consistent presentation within the thesis.

Student signature: [redacted] **Date: 26 August 2015**

The undersigned hereby certify that the above declaration correctly reflects the nature and extent of the student and co-authors' contributions to this work.

Main Supervisor signature: [redacted] **Date: 26 August 2015**

Table of contents

LIST OF FIGURES

LIST OF TABLES

Acknowledgements

Many people and organisations have helped this dream become a reality. When I first approached Prof Ray Cas (then Head of School, now emeritus) about the possibility, his enthusiastic support gave me great confidence that my dream was not a folly. Without that support, I would never have started this study, which has been one of the most fortunate times in my life. The unwavering backing of Monash University at all levels, but particularly the School of Geosciences (now the School of Earth, Atmosphere and Environment) and the Faculty of Science have been a constant source of encouragement, even in the days when I was making little progress. My supervisors, Assoc Prof Peter Betts, Prof Mike Hall (emeritus) and Dr Laurent Aillères, have at all times been wonderful sources of guidance. Pete helped me through the huge advances in geology since I last was at University. Mike was invaluable as a source of knowledge on Tasmanian geology. The understanding I gained on my early field trip with him is the foundation on which this thesis is based. Laurent was always in the background helping solve the computing issues that those of us who grew up in a pre-P.C. age inevitably have. Many times, other members of the Structural Geophysics Group, particularly Stefan Vollgger and Tegan Blaikie, were my computer gurus. At my confirmation, Prof Chris Wilson asked me to consider Antarctica and its implications in my research. This led to a fundamental change in my thinking. To each of you I owe debts that can never be repaid.

My room-mates in 106 and G42a have also been a constant source of help, both technical and psychological. Zara Little, Nic Unland and Ravin Dao, who finished before me will never know what they gave me, but their contribution was substantial. More recent room-mates, Matt Edwards, Stefan and Nasim Kharazizadeh, and subsequently, Dan Uehara, Jack Simmons, Frediano Clos and Bill Howcroft have given me support when needed, space at other times, and ignored my many foibles.

My two student field assistants, Richard Hills and Joel Vergunst, and Honours student Jesse Franks who I accompanied to King Island, were crucial helps for me in gaining an understanding of the rocks. The questions they asked forced me to think more deeply about the rocks we were looking at and the wider context in which they sat. As the song in the Sound of Music says, “by your students you are taught”.

The School of Geosciences Administration has smoothed my path at all times. I particularly recall the efforts of Dr Caroline Venn and Rob Oakley in sorting out problems early in my candidature. My transition to becoming a student again was made much easier by their knowledge of the Monash administrative systems. Since then, the succession of staff through the front office has been universally helpful.

Outside of the University, many gave me invaluable assistance. Dr Tony Brown, Executive Director Mineral Resources at Mineral Resources Tasmania gave me free access to MRT data, a priceless resource compiled over 100 years of mapping by dedicated geologists. Discussions with Dr Clive Calver and David Seymour gave me insights into Tasmanian geology that I would never have obtained otherwise. The Information and Lands Services Branch, Department of Primary Industries, Parks, Water and Environment kindly gave me free access to their transport and hydrological databases.

Minerals and Petroleum Victoria was also extremely helpful. Working with the Geological Survey of Victoria for 13 years gave me the background geological knowledge of the state to attempt the chosen topic. Of particular value was the two years spent with the Basin Studies Group, interpreting the basement to the Gippsland and Otway basins, which started me looking south. The GSV was always an intellectually challenging place in which to work, with Fons VandenBerg, Clive Willman, Vince Morand, David Taylor and, in particular, Ross Cayley all thinking deeply about Victorian geology. Ross' thinking about eastern Australian tectonics continues to challenge us all.

As well as the generous financial support provided by Monash University, I gratefully acknowledge support from the Bicentennial Gold 88 Endowment Fund and the Geological Society of Australia, Victoria Division. The former assisted with travel and other expenses when visiting King Island, and the latter with support to attend the 2011 Fall Meeting of the American Geophysical Union. Both trips were of great value in building my knowledge.

I would never have arrived here without a long list of people who have helped my understanding of geology. The staff of the Geology Department at the Australian National University introduced me to the wonders of rocks. Prof David Brown, (then)

Dr Alan White and his new postgraduate, Dr Bruce Chappell were inspirational figures in their love of what they were doing. As much as anything, they taught me that doing great science could give great enjoyment. Not only that, they made their science understandable. Prof Peter Coney (University of Arizona, Tucson) gave me a whole new vision of geology, particularly tectonics. Before working with him at BHP Minerals, I was a geological stamp collector—gathering facts without any understanding of their relationships. He gave me a structure in which to place these facts and thereby enhanced my understanding of geology enormously. Sadly, all have now passed away. Dr Tom Whiting taught me how to interpret potential field data and then placed me in a position at BHP Minerals where I could exercise these new skills each day, thereby consolidating them and so giving me a new career.

Lastly I thank my wife, Sonya. Her forbearance has been beyond what anyone should have to endure. Even though she did not feel that doing a PhD was my best decision, she gave me absolutely unwavering support and love. No husband could possibly wish for more.

Chapter 1

Come my friends,
'Tis not too late to seek a newer world.

Ulysses; Tennyson, 1842

Chapter 1: Introduction

1.1 SCOPE OF THE PROJECT

Accretionary orogenic systems are a fundamental part of Earth history. They are taking place today in the western Pacific (Crawford et al., 2003b) and can be traced back at least to the Archean-Proterozoic boundary (Evans, 2013). The very nature of our planet, with a continental crust that incompletely covers its surface, demands that it loses heat at different rates in oceanic and continental regions, and that the heat build-up under continental crust will ultimately cause the breakup of any large continent formed. In turn, the spherical geometry of the planet means that every breakup event is the precursor to an accretionary event. Sometimes large pieces of continental crust are accreted, such as in the collision of India with Asia (Patriat and Achache, 1984). Elsewhere, continents accrete micro-continental and island arc fragments at their margins that are embedded in a matrix of sedimentary rocks largely derived from the continent itself. The Altaids and the Tasmanides are of this style (Gray and Foster, 2004; Glen, 2005; Wilhem et al., 2012).

Each style of accretion brings its own challenges in understanding. The collision of India with Asia has had major effects on the adjacent regions (e.g. Metcalfe, 2013). The gross setting for the Tasmanides is intrinsically simple, lying between the Gondwanan margin and the circum-Pacific subduction system that has been present since the start of the Paleozoic (Coney, 1992; Cawood, 2005). However, within this system there are many complexities, such as poor or non-existent along-strike correlation of micro-continental ribbons and their boundaries. These ribbons will almost always have had a strong competency contrast with the adjacent young sedimentary rocks, and so subsequent compressive events are likely to have caused the sedimentary rocks to have been thrust over the rigid micro-continental ribbons. Internal subduction zones may have existed only briefly, and they may have wrapped around micro-continental ribbons, as modelled by Moresi et al. (2014). Similar features are seen today in the Banda Sea (Hall, 2002). In a history spanning several hundred million years, younger events will have overprinted older ones, further obscuring understanding of the early history of the region. The speed of advance or retreat of the master subduction zone will also have had a profound effect on the nature of the orogen formed (Cawood et al., 2009). For example, retreating subduction zones extend the adjacent crust and generate back-arc basins, while advancing subduction zones cause material to accrete to the continental margin (Schellart, 2008). The Tasmanides underwent long periods of subduction zone retreat separated by shorter periods of advance, ensuring a complex history (Collins, 2002).

Within the Tasmanides, it has long been recognised that the pre-Carboniferous rocks of Tasmania and Victoria have very different geological histories (e.g. David, 1950). Victorian rocks are clearly part of the greater Tasmanides, typically with the oldest parts comprising Cambrian ocean floor basalt overlain by Ordovician meta-turbidites rocks that are in turn overlain by younger sequences that are meta-turbidite–dominated. Almost all were laid down in oceanic environments. Granitic bodies were intruded at about 490 Ma and from approximately 410 Ma onwards. In contrast, rocks in western and central Tasmania clearly go back to the Mesoproterozoic, and the Paleozoic rocks present were deposited in continental or marginal marine settings. These rocks are present as far north as King Island, but the only equivalent sequence in Victoria is a small area on the south coast at Cape Liptrap.

Many of the answers to this enigma lie under Bass Strait. But what is the basement there, or to rephrase the problem, what are the connections between the Early Paleozoic rocks of Victoria and the Early Paleozoic and Proterozoic of Tasmania? How did they get to be this way? Do we see any evidence of the Proterozoic Tasmanian rocks in Victoria, and are the Paleozoic events seen in Victoria reflected in events in Tasmania? How do these events correlate with events seen elsewhere along the Early Paleozoic margin of Gondwana? And why is western Tasmania so different to the rest of the Tasman Orogen?

1.2 PROJECT AIM

The aim of this thesis is to correlate the Proterozoic and Early Paleozoic rocks of Tasmania across Bass Strait into Victoria and to outline the effects seen in the Early Paleozoic rocks of the Lachlan Orogen of southeastern Australia. This includes the following tasks;

- Deciding on the most appropriate model from the many outlined over the last 20 years. For example, the Selwyn Block model (Cayley et al., 2002) is just one of many that should be tested against the geological and geophysical data. Without a potentially valid model, any subsequent steps are pointless;
- Subdividing the Pre-Ordovician rocks of Tasmania into strato-tectonic packages. Each package should have a unique geological history that can be linked together to form a single history that goes back to the formation of the earliest rocks known, and that explains their dispositions today;
- If possible, extrapolating these mostly Precambrian packages across Bass Strait into Victoria and outlining their effects on Victorian geology. These effects should be interpreted from both sedimentary and igneous rocks and explain the enigmatic magnetic and seismic responses seen in central Victoria. The effects should be examined both in areas where Proterozoic and Cambrian Tasmanian crust might continue across Bass Strait and also in the adjacent rocks of the Lachlan Orogen.

1.3 DESCRIPTION OF RESEARCH METHODOLOGIES

This thesis uses a multi-faceted approach to the research. Most critically, it relies on understanding the geology of the region. Both the Geological Survey of Victoria and Mineral Resources of Tasmania have 1:25,000 or 1:50,000 scale maps of many of the important areas of the region, such as the coastlines and adjacent areas. However, some significant areas have not been mapped or the mapping has not recognised features whose significance only becomes apparent when integrated with other data. Mapping away from the Tasmanian coastline is problematic, given the extensive thick vegetation cover, deep weathering and younger rock cover. In Victoria, only a few outcrops of the relevant rocks crop out on or near the coast. The remainder are covered by Cretaceous and younger rocks of the Gippsland and Otway basins. A critical aid in understanding was the field work carried out in Tasmania (2 months), King Island (3 weeks) and southern Victoria (approximately 1 week). Almost all of this was carried out along the coast.

As well, there is an extensive literature on the relevant rocks in both states. Most was written for other purposes, but often papers contained relevant isolated facts that together helped inform the larger interpretation. This necessitated literature synthesis of the often disparate observations and data into a single, coherent interpretation.

Synthesis involved collection of field based geological data, geological literature and externally collected databases, which included digital geological data from Victoria (Welsh et al., 2011) and Tasmania, (Mineral Resources Tasmania Data Management Group, 2011), and OZCHRON SHRIMP (Black, 2007) and Sm-Nd databases (Champion, 2013).

Hydrocarbon exploration data were rarely useful. Only 8 offshore oil and gas exploration holes reached basement in the region, and 4 of these were west of the Bass Basin. This meant that the geophysical interpretation of Bass Strait was poorly constrained in some areas. As well, the seismic data from hydrocarbon exploration were processed to optimise contrasts in the Bass Basin, meaning that there was little contrast in the basement data, if any data were present. In contrast, the 1995 Geoscience Australia/Mineral Resources Tasmania seismic survey, GA148, which circumnavigated Tasmania, was targeted at the older rocks and data was collected to 16 s, yielding information to approximately 48 km depth. These data yielded important information about the locations and dips of many of the major structural breaks that cross the Tasmanian coast.

Other geophysical data used included airborne magnetic surveys with flight line spacings between 200 m and 500 m across almost all of Victoria, much of Tasmania and all of Bass Strait. These data have been stitched together by Geoscience Australia in an Australia-wide stitch with a cell size of 0.00083333° (approximately 80 m); the stitch available in 2010 was used as the basis for this study.

The magnetic data has the advantage of providing a uniform, high-quality data set in which all other observations could be placed and hypotheses tested. The data mostly map the variable presence of just two minerals, magnetite and monoclinic pyrrhotite; magnetite is by far the dominant magnetic mineral. Many sedimentary rocks are all but non-magnetic and so the method generally fails to distinguish between different meta-sediments, although careful imaging of high frequency responses can yield valuable structural information from those that do contain traces of magnetite or pyrrhotite. In contrast, significant proportions of oxidised igneous rocks contain magnetite-rich phases and can be distinguished by the magnetite within them, which provide not only spatial information about the outlines of the bodies, but also indications about what the bodies might be. For example, many oxidised granites in western Victoria can be distinguished by their magnetic outer parts that surround an essentially non-magnetic core (VandenBerg et al., 2000). The method has the advantages of being able to obtain responses from buried bodies and that the shapes of the responses change with depth, so that images of deeply buried magnetic bodies can often be distinguished from shallower, less magnetic bodies.

Many of the post-Paleozoic rocks in the region, notably Cretaceous and Cenozoic basaltic units, are strongly magnetic. Their responses mask more subtle responses from underlying bodies. To some extent this problem can be overcome by image enhancement techniques, as the young basalt bodies often have strong high-frequency magnetic responses. Even so, using many different image types and carefully filtering out high frequency responses is only a partial solution where a young, strongly magnetic sheet overlies an older weakly magnetic body. Magnetic remanence also needs to be considered. Where magnetic modelling has been carried out, I have assumed that only the post-Paleozoic rocks may carry remanent magnetism. Older rocks have been metamorphosed, mostly to greenschist facies or above, and this is likely to have reconstituted the magnetite present (Dentith and Mudge, 2014). This assumption may not always be valid, but is likely to be so in most cases. Finally, when magnetic minerals are at temperatures above their Curie points (approximately 580° for magnetite and 320° for pyrrhotite) they become non-magnetic and magnetic interpretation is not possible.

Radiometric data was generally collected with the magnetic data, and in some areas of highly weathered outcrop, this data source provided useful additional information as to the locations of boundaries. However, where the rocks are covered by water or more than a few tens of centimetres of young cover, any radiometric data collected was of no value to the present study.

Topographic data also provided helpful supplementary information, since lithology has a strong control on topography. High quality topographic data are routinely collected during airborne magnetic surveys and were also collected over the submarine plateaux south of Tasmania. While non-unique, the data provided some control in areas where other data were inadequate. They were also valuable when imaged in conjunction with radiometric data.

Because density is an intrinsic property of all rocks and gravitational attraction is a function of density, then gravity data should provide a method of outlining different rock packages. Furthermore, it is less limited by the depth of the causative body than magnetic data since there is no equivalent to the Curie point. However, when compared to magnetic data, the typical range of variation in density is less than a factor of 2 compared to several orders of magnitude variation in magnetic susceptibility, the data is more costly to acquire and so often the stations are more widely spaced than is optimal. The gravity data used were from Geoscience Australia, and both the 2008 onshore bouguer/offshore free air and 2012 isostatic compilations were used. These data sets had cell sizes of 0.0083333° (approximately 800 m), implying an order of magnitude less precision than the magnetic data in the final grid. As well, the data quality varied significantly. In parts of northwestern Victoria, the nominal station spacing is 0.5 km (Moore and McLean, 2009), while in Bass Strait the gravity grid available was derived from lines up to 100 km apart. Furthermore, there appeared to be major difficulties in stitching together the onshore and offshore data sets as no surveys cross the land-water interface. Indeed, most marine surveys are carried out well away from coastal areas. Thus interpretation in these areas was problematic.

The data outlined above were combined into a geological interpretation that extends from 35°50’S to 51°40’S, and from 142°E to 152°E, approximately 1700x550 km. The interpretation of the most important part of this region, across Bass Strait, was tested in seven sections that forward modelled the magnetic and gravity data. Forward modelling is a method of constructing a cross section, populating polygons with rock properties (density and magnetic susceptibility) and comparing the calculated response with the observed response. The cross section geometry and/or rock properties are iteratively modified until there is appropriate match between the calculated and observed responses. The method relies completely on the quality of the input data, and all of the strengths and weaknesses outlined above for the magnetic and gravity data are present in the models. The uncertainties also include the geological model being tested, since the geological knowledge used to develop a model is mostly derived from near-surface observations and the sections generated were modelled to 30 km deep, where our understanding is much weaker than in the near-surface. Furthermore, the gravity response decreases as the square of the distance to the sensor increases, and the magnetic response decreases as between the cube and the fifth power (depending on the orientation of the source and the the Earth's magnetic field). This means that the uncertainties will correspondingly increase. In addition, because the models created were 2D models, fields largely derived from off-section bodies could not be appropriately modelled even though they may have caused significant variations along the modelled profiles. Because some of the sections modelled were over 500 km long and 30 km deep, some models reached the limits of the Gm-Sys® software, and no more surfaces could be included in the model. Even so, the models provided important insights into the crust under Bass Strait and adjacent areas.

1.4 ORGANISATION OF THESIS

Chapter 2 contains a wide-ranging literature review that synthesises the Pre-Carboniferous geology of Victoria and Tasmania. It then evaluates the many hypotheses that have attempted to link the geology of Victoria and Tasmania against this synthesis and the geophysical data, and concludes that the Selwyn Block model of Cayley et al. (2002) is the most likely model to fit the available data. However this model says little about the relationships between the Selwyn Block (that underlies central Victoria) and Tasmanian geology. The chapter was published in Gondwana Research (Moore et al., 2013).

Chapter 3 provides a synthesis of the geology of Tasmania and a geological history that is internally consistent and consistent with that of VanDieland (the Selwyn Block, western Tasmania, the west South Tasman Rise and the East Tasman Plateau, Cayley, 2011). It subdivides Tasmania into 7 zones, with another, the Glomar Zone, further south. The subdivisions provide the basis for extending Tasmanian geology across Bass Strait in Chapter 4. The history outlined suggests Mesoproterozoic linkages with both East Antarctica and Laurentia, thereby implying constraints on future Rodinian paleo-geographic reconstructions. It also suggests a mechanism for the generation of the Cambrian volcanic hosted massive sulphide deposits in western Tasmania. The chapter was published in the Australian Journal of Earth Sciences (Moore et al., 2015).

Chapter 4 extrapolates the subdivisions proposed in Chapter 3 across Bass Strait and into the Selwyn Block, thereby providing the first strato-tectonic map of pre-Carboniferous Bass Strait and the Selwyn Block. It places all of the pre-Ordovician rocks in central Victoria into a single framework that is consistent with that previously outlined in Tasmania. It gives a source for the exotic conglomerate clasts seen in the southeastern Melbourne Zone. It offers explanations for many of the otherwise difficult to explain features observed in some plutons of the Central Victorian Granite Province, including the presence of calc-silicate xenoliths, abnormal Ni and Cr values, the distribution of ε_{Nd} values and the orientations of some intrusions. It suggests that the deep magnetic responses seen in the Selwyn Block may be due to the metamorphism of an underlying mafic volcanic unit. The chapter also examines the gross faulting patterns in rocks in eastern Victoria and eastern Tasmania and concludes that the hypothesis outlined by Moresi et al. (2014) for southeastern Australia appears to be valid in Tasmania.

Chapter 5 collates the conclusions outlined above, presents a strato-tectonic map of VanDieland and provides a more wide-ranging synthesis of the southeastern Australian part of the early Gondwanan margin. It concludes with suggestions for future research, including the search for other micro-continental fragments along the early Gondwanan margin, the relationships between granite source terranes and their final compositions, and the relationships between VanDieland and the overlying Bass Basin.

PART B: Declaration for Thesis

Monash University

Declaration for Thesis Chapter 2

Declaration by candidate David H Moore

In the case of Chapter 2, the nature and extent of my contribution to the work was the following:

Nature of contribution	Extent of contribution (%)
Did the research, wrote almost all of the paper	90%

The following co-authors contributed to the work. If co-authors are students at Monash University, the extent of their contribution in percentage terms must be stated:

Name	Nature of contribution	Extent of contribution (%) for student co-authors only
Peter Betts	Supervision, guidance on text	
Mike Hall	Supervision, assistance with field work	

The undersigned hereby certify that the above declaration correctly reflects the nature and extent of the candidate's and co-authors' contributions to this work*.

Candidate's Signature	[redacted]	26 August 2015
Main Supervisor's Signature	[redacted]	26 august 2015

*Note: Where the responsible author is not the candidate's main supervisor, the main supervisor should consult with the responsible author to agree on the respective contributions of the authors.

Chapter 2

The student must remember, for all his consolation….. that failures are almost as important to the cause of science and to those who follow him in the science road, as his successes. It is as much to know what we cannot know, what we cannot do in any given direction – the first step indeed, toward the accomplishment of what we can do.

J.L.R. Agassiz 1896 Geological Sketches.

Towards understanding the early Gondwanan margin in southeastern Australia

ABSTRACT

This review synthesises the Proterozoic and early Paleozoic geology of Tasmania, Bass Strait and western and central Victoria. We examine the many different conflicting hypotheses that have been proposed to solve the paradoxical relationships between Tasmanian geology and that of mainland Australia, most notably the prevalence of Proterozoic basement of western and central Tasmania, whilst immediately across Bass Strait evidence of Proterozoic rocks is much more cryptic. We conclude that the Selwyn block model is the most satisfactory hypothesis to date, since it fits best with the obvious patterns in the magnetic and gravity data. This model proposes that the central Victorian Melbourne Zone is underlain by the northern extension of thin Tasmanian Proterozoic and Cambrian crust under Bass Strait, and that the Silurian to Middle Devonian Melbourne Zone was shortened along a décollement during the Tabberabberan Orogeny. The Ordovician rocks of eastern Tasmania correlate more closely with the Tabberabbera Zone than the Melbourne Zone in Victoria; however the Silurian and Devonian correlations are less certain. Major unresolved issues are the origins of the Proterozoic and Early Cambrian lithostratigraphic packages, tectonic models for their assembly during the Tyennan Orogeny, and how these models fit with those for mainland Australia.

2.1 INTRODUCTION

Understanding the relationships in Gondwana between Australia and Antarctica cannot be solved without an understanding of how the geology of Tasmania fits into the eastern margin of the supercontinent. Tasmania is often ignored, for example by Flöttmann et al. (1998), Betts et al. (2002) and Veevers (2007). Those who choose to include Tasmania in post-Rodinian reconstructions are challenged by the wide variety of incompatible and conflicting interpretations of Tasmania's relationships with the rest of the eastern Gondwana margin. One reason for these disparate interpretations is that major changes take place either beneath Bass Strait, a 300 km wide stretch of water separating Victoria and Tasmania, or under the Cretaceous and Cenozoic Gippsland and Otway basins that dominate Victorian exposures in the Bass Strait region.

The geology of Victoria and Tasmania are along strike from each other, and an Upper Cambrian to Lower Ordovician paleomagnetic pole reported by Li et al. (1997) places northwestern Tasmania within a few hundred kilometres of its present position with respect to the Australian mainland. Despite this, there are several fundamental differences that are difficult to reconcile. The oldest Tasmanian rocks known are the Mesoproterozoic meta-turbidites on King Island, and Proterozoic rocks are ubiquitous in northwestern Tasmania (Burrett and Martin, 1989; Seymour et al., 2007; Berry et al., 2008). By contrast, in Victoria, just 80 km north of King Island, there are no Proterozoic rocks known to crop out (Crawford et al. 2003a) (Figure 2-1). The Tasmanian Ordovician and Silurian rocks in the western and central parts are mostly terrestrial or marginal marine, whilst across Bass Strait the coeval Victorian Paleozoic rocks are largely deep marine turbidites (Burrett and Martin, 1989; VandenBerg et al., 2000; Crawford et al., 2003a). In Tasmania, the dominant Cambrian deformational and thermal event, the Tyennan Orogeny, took place from about 515 to 505 Ma, with less significant early events at about 520 Ma and later events extending to about 490 Ma (Berry et al., 2007). In Victoria and South Australia, the broadly contemporaneous Delamerian deformation was initiated as far back as 545 Ma, whilst the peak metamorphism is generally younger than 505 Ma (Morand et al., 2004; Miller et al., 2005; Foden et al., 2006; Turner et al., 2009). In Tasmania, the Tyennan metamorphism formed both blueschist and eclogite facies rocks, whilst in Victoria and southeastern South Australia the Delamerian metamorphism is typically of a moderate to high temperature-low pressure style, although intermediate pressure rocks formed along the Moyston Fault (Preiss, 1995; Meffre et al., 2000; Cayley et al., 2002; Phillips et al., 2002; Morand et al., 2004). In Tasmania, Crawford and Berry (1992) argued that the Cambrian subduction system was east dipping, whilst in Victoria and Antarctica, north and south of Tasmania, the geological record suggests a west-dipping subduction system (Finn et al., 1999; Miller et al., 2005).

This paper describes the geology of Victoria and Tasmania and outlines the various, often conflicting, hypotheses invoked to try to resolve the paradoxical relationships outlined above. We will outline the geological, magnetic, gravity and seismic data,

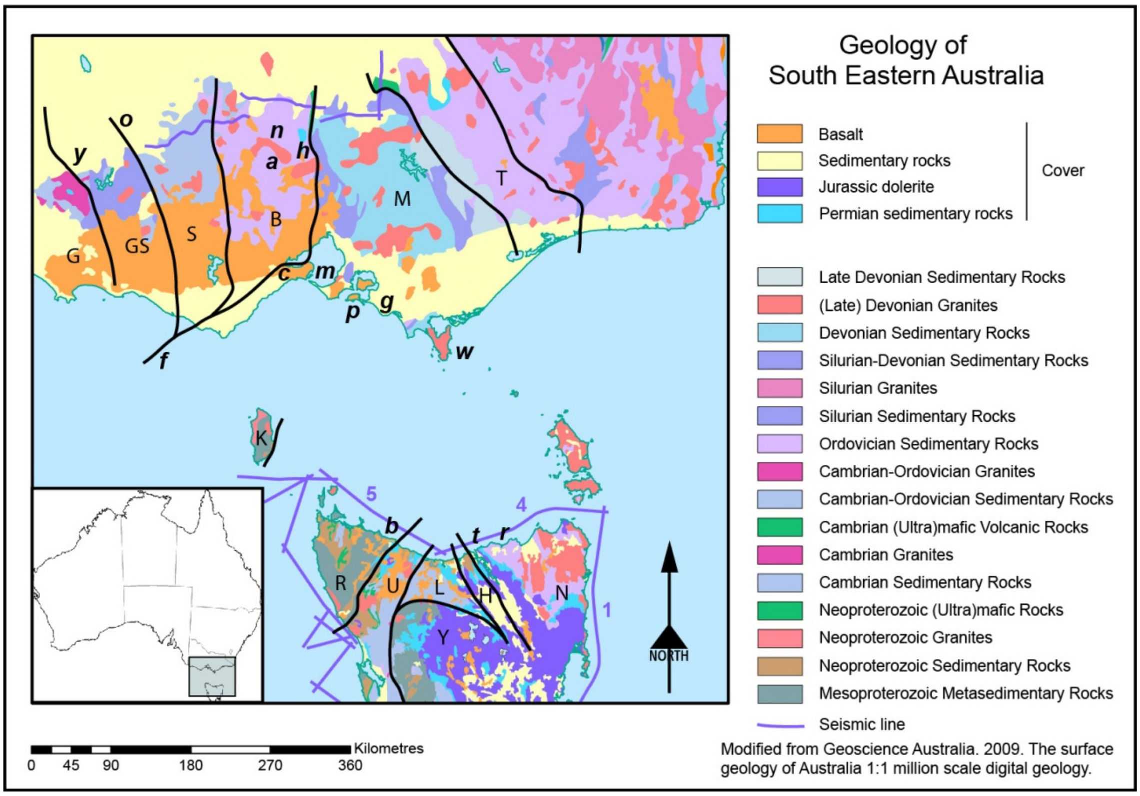


Figure 2-1. Geology of southeastern Australia showing the major Proterozoic to Lower Paleozoic structural zones and locations mentioned in the text. B marks the Bendigo Zone, G the Glenelg Zone, GS the Grampians-Stavely Zone, H the Badger Head Block, K King Island, L the Sheffield-Ulverstone area, M the Melbourne Zone, N northeast Tasmania, R the Rocky Cape Block, S the Stawell Zone, T the Tabberabbera Zone, U the Burnie area and Y the Tyennan Block. Locations mentioned in the text are *a*, Castlemaine, *b*, Black River, *c*, the Ceres Gabbro, *f*, the Bambra Fault, *g*, Wonthaggi, *h*, Heathcote, *m*, the Mornington Peninsula, *n*, Bendigo, *o*, the Moyston Fault, *p*, Phillip Island, *r*, Piper River *t*, the Tamar region, *w*, Wilsons Promontory, and *y*, theYarramyljup Fault. The Arthur Lineament forms the boundary between the Rocky Cape Block and the Burnie area. Seismic lines referred to in text are numbered.

much of which has been collected or reprocessed since 2000, to evaluate these hypotheses and to conclude which are most likely to be valid.

2.2 HISTORICAL DATA

2.2.1 GEOLOGY

The oldest rocks known from the region are from the southern tip of King Island, where SHRIMP analyses of zircons showed that the psammitic and pelitic rocks were deposited after 1350 Ma, whilst monazite dating showed that they were metamorphosed at 1290±20 Ma (Figures 2-1 and 2-2) (Black et al., 2004; Berry et al., 2005). These rocks have no known equivalents on Tasmania or Victoria (Crawford et al., 2003a; Calver, 2007). Mapping by Calver (2007) divided the basement of King Island into a western part of amphibolite facies metasedimentary rocks and an eastern part of greenschist to lower amphibolites facies metaturbidites. He believed that the eastern metaturbidites were the high-temperature low-pressure metamorphic equivalents of the western succession (Blackney, 1982; Calver, 2007).

In Tasmania, the Rocky Cape Group of the Rocky Cape Block are the oldest rocks mapped. These rocks are dominantly shallow marine to neritic metaquartzites and associated siltstones. These rocks are exposed in northwestern Tasmania, and the detrital zircons present indicate that the upper parts are younger than 1000 Ma and the lower parts no older than about 1430 Ma (Burrett and Martin, 1989; Black et al., 2004; Everard et al., 2007). Calver et al. (2010) correlated the *Horodyskia williamsii* 'string of beads' trace fossils present low in the Rocky Cape Group with similar fossils in the Bangemall Basin in W.A., where they are constrained to rocks aged between 1465 and 1700 Ma. The basement to the Rocky Cape Group is not exposed. The age of the earliest deformation is uncertain, but at the eastern edge of the Rocky Cape Block, rocks that may be younger than it are intruded by a granite with an age of 777±7 Ma (Turner et al., 1998; Calver et al., 2010). On northern King Island, Turner et al. (1998) used U/Pb zircon geochronology to document a second metamorphic event, the Wickham 'Orogeny', that took place at 760±12 Ma.

Several other sequences in western and central Tasmania also contain metasedimentary rocks where the youngest zircon populations range from 1400 to 1450 Ma (Black et al., 2004). All were apparently sourced from the same, probably Laurentian, source (Berry et al., 2001; Burrett and Berry, 2001; Black et al., 2004). Even where younger sources are present, e.g. Badger Head in north central Tasmania (1242±29 Ma) and the Jacob Quartzite further west (1010±45 Ma), the 1400 to 1450 Ma zircons are also present (Black et al., 2004). Chmielowski (2009) found older monazites ranging from 1000 to 1400 Ma, with a peak at 1376±7 Ma; several monazites also had overgrowths that grew at 1220±36 Ma. Whilst these give maximum ages of sedimentation, there are few other constraints. A K/Ar age by Crook (1979) on a dolerite intruded into wet sediments was recalculated by Black et al. (2004) as 711±16 Ma; they considered this a minimum age of sedimentation. The 777±7 Ma granite may also provide a constraint, although Holm and Berry (2002) considered that the rocks into which it was intruded were allochthonous.

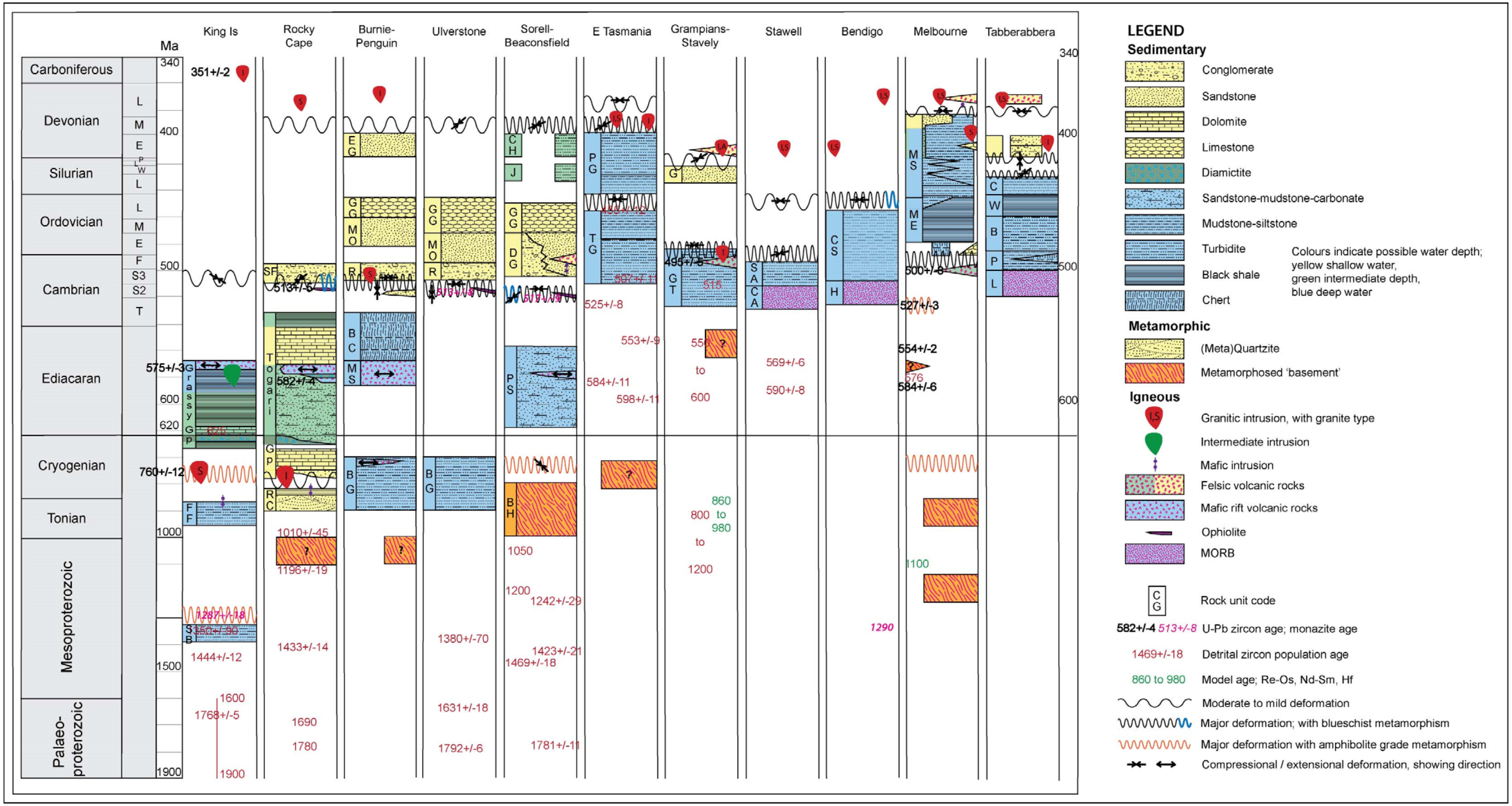


Figure 2-2. Tectono-stratigraphic columns for the Bass Strait region. Abbreviations used:- for King Is; FF, Fraser Formation; SB, Surprise Bay Formation; for Rocky Cape; Rocky Cape Group, SF Scopus Formation; for the Burnie-Penguin region, BC Barrington Chert, BG Burnie Group, EG Eldon Group, GG Gordon Group, MO Moina Sandstone, MS Motton Spilite, R Owen Group; for the Ulverstone region, BG Burnie Group, GG Gordon Group, MO Moina Sandstone, R Owen Group; for the Sorell-Beaconsfield region, CH Corn Hill Formation, DG Denison Group, J Johnson Creek Formation; for eastern Tasmania, PG Panama Group,TG Tippogoree Group; for Grampians-Stavely Zone, G Grampians Group, GT Glenthompson Sandstone; for the Stawell Zone, Carrolls Amphibolite, SA St Arnaud Group; for the Bendigo Zone, CS Castlemaine Supergroup, H Heathcote Volcanics; for the Melbourne Zone, ME Mount Easton Shale, MS Murrindindi Supergroup; for the Tabberabbera Zone, B Bendoc Group, C Cobbannah Group, L Lickhole Volcanics, P Pinnak Sandstone, W Warbisco Shale.

The Rocky Cape Group is unconformably overlain by the mostly shallow-dipping Neoproterozoic to Cambrian Togari Group, which occupies the Smithton Synclinorium (Calver, 1998; Black et al., 2004; Everard et al., 2007; Seymour et al., 2007). At Black River, the difference in dips is about 30° (Figure 2-3). The Togari Group includes quartzite, dolomite, siltstone, mudstone, conglomerate, both Cryogenian and Ediacaran (Sturtian and Marinoan) diamictite and rift-related mafic rocks ranging from basalt and dolerite to picrite (Calver et al., 2004; Everard et al., 2007). Equivalent aged rocks are present on the southeast coast of King Island, where Sturtian diamictite, dolostone, laminated shale, andesitic sills, tholeiitic pillow lavas, picritic pillow lavas and MORB tholeiitic basalts unconformably overlie the Mesoproterozoic rocks (Calver, 2007; Calver, 2009; Hoffman et al., 2009). Other correlatives have been inferred as far east as the Ulverstone Metamorphic Complex and the Port Sorell Formation, both in coasal central northern Tasmania (Figure 2-1) (Berry and Gray, 2001; Reed et al., 2002).

Figure 2-3. Unconformity at the base of the Smithton Synclinorium between the ?Mesoproterozoic Cowrie Siltstone and the Neoproterozoic Forest Conglomerate. It marks the beginning of a period of extension that generally continued until the Tyennan-Delamerian Orogeny. Location 40°50’48”E, 145°18’28”S.

The mafic volcanic rocks of the Togari Group and on King Island have a chemistry consistent with having formed in a rifting environment, and have been interpreted to have recorded the final Rodinian break-up (Meffre et al., 2004). A SHRIMP date on the andesitic sills yielded an age of 575±3 Ma and the picritic pillow lavas a Nd-Sm isochron age of 579±16 Ma (Calver et al., 2004; Meffre et al., 2004). The mafic rocks are moderately to strongly magnetic (to 600 nT) and lie at the western edge of a 50 kilometre-wide magnetic and gravity high that extends as far north as Phillip Island, just off the south coast of Victoria (Figures 2-4, 2–5). Here an isolated area of picritic basalt, dolerite and meta-cumulate rocks contains chromian spinels that are geochemically similar to those in the rift-related basalts on the east coast of King Island (Henry and Birch, 1992; Bushby, 2002).

The Arthur Lineament (Figure 2-1) is a prominent structure that forms the boundary between the Rocky Cape Block and the Burnie and Oonah Group rocks to the east. The Lineament is about 8 km wide at its southwestern end on the western Tasmanian coast, but in the northeast it is largely covered by Permian tillite where it crosses the north Tasmanian coastline (Holm and Berry, 2002). The rock-types present are metasediments, including variably deformed and metamorphosed dolomite, sandstone, siltstone, mudstone and conglomerate and quartzite, together with mafic schist, gabbro, granite, amphibolite and dolerite with E-MORB rift tholeiite geochemistry (Holm et al., 2003); they are mostly from the adjacent Proterozoic sequences although the mafic Bowry Formation may be allochthonous (Holm and Berry, 2002; Bottrill and Taheri, 2007). Some of the mafic packages have been metamorphosed to blueschist facies (700 MPa at 350° C) early in the Tyennan Orogeny (Turner and Bottrill, 2001; Holm and Berry, 2002). Both mapping and geophysical modeling suggest the eastern edge of the Arthur Lineament dips east whilst the western edge is more complex, with a near-surface east-dipping contact apparently overlying a deeper west-dipping break (Holm and Berry, 2002; Leaman and Webster, 2002). Holm et al. (2003) interpreted the rocks as mostly having been formed in a rift tectonic setting, perhaps at the 777±7 Ma age of the granite (Turner et al., 1998). In contrast, Calver and Walter (2000) interpreted rifting to have taken place between 650 and 550 Ma.

The Motton 'Spilite' forms another basalt-chert package on the central northern Tasmanian coast (Jago and Brown, 1989; Vicary, 2006; Vicary et al., 2008). The age of the basalt is poorly constrained; however Vickery (2006) considered it to be Early Cambrian, since clasts of the spilite are present in an overlying Cambrian conglomerate. The origin of the spilite is interpreted to be oceanic, based on its geochemical signature (Vicary, 2006).

What may be the oldest rocks in central Victoria are undated. At Ceres, near Geelong, an isolated outcrop of metagabbro has been metamorphosed to amphibolite facies and subjected to north-south shortening. Both events are unusual in Victoria, where the Pre-Carboniferous rocks have almost always been deformed during east-west shortening and the nearest outcrops of amphibolite facies rocks are about 150 km away (VandenBerg et al., 2000). Morand (1995) considered that the rocks were probably Cambrian, whilst Cayley et al. (2002) interpreted them to be one of the few outcrops of the Selwyn Block, the northern extension of the Tasmanian Precambrian and Cambrian cratonic crust across Bass Strait into southern and central Victoria. Another upper greenschist to amphibolite facies metagabbro is present at Waratah Bay to the immediate west of Wilsons Promontory. This lies below a Tyennan-Delamerian unconformity (Cayley et al., 2002).

The presence of other older rocks can be inferred in southern Victoria. On the Mornington Peninsula, at the southern end of Port Phillip (Figure 2-1), there is a 24 m thick sequence of Early Ordovician turbidites; however the underlying rocks are not exposed. Further north near Bendigo, the equivalent sequence is 550 m thick, and 450 m near Castlemaine. It suggests that the Mornington Peninsula section represents a starved sequence and that other older rocks must lie below it. Cayley et al. (2002) interpreted these older rocks to be part of the Selwyn Block. Other apparently starved chert and mudstone dominated successions crop out at Wonthaggi and north of Wilsons Promontory, in southernmost Victoria. The low K tholeiitic basalt found by Henry and Birch (1992) that crops out on the south coast of Phillip Island was also included in the Selwyn Block.

In western Victoria, detrital zircons aged about 580 Ma are present in metaturbidites deformed in the Delamerian Orogeny (Maher et al., 1997; Fanning and Morand, 2002). Tholeiitic basalts and related rocks are interlayered with the metaturbidites and Morand et al. (2004) interpreted them as within-plate basalts. Maher et al. (1997) recorded a metagabbro containing zircons with a SHRIMP age of 524±9 Ma, implying the host rocks must be older than this. In far southwestern Victoria, the Hummocks Serpentinite is an altered cumulate harzburgite with a Nd model age of about 700 Ma (Turner et al., 1993) with an error of approximy 200 m.y. (J. Foden pers.comm. in VandenBerg et al., 2000). All are consistent with a rift-related setting in the last stages of the Rodinian breakup, implying an Ediacaran or Cambrian depositional age for the sedimentary rocks.

Other MORB-type and boninitic volcanic and intrusive rocks seen in the hangingwalls of major faults in western and central Victoria may also have begun to form in the Early Cambrian (Crawford et al., 1984; VandenBerg et al., 2000). Near Heathcote, Jell (pers. comm. in VandenBerg, 1991) found Early Cambrian trilobites in the interflow sedimentary rocks. These volcanic rocks formed in a back-arc or island arc-related setting (Crawford et al., 1984; Squire et al., 2006). In central northern Tasmania, the Andersons Creek Ultramafic Complex occupies a similar stratigraphic position, since it seems to lie stratigraphically below the base of the Ordovician to Devonian Mathinna Supergroup of eastern Tasmania (Reed et al., 2002).

The oldest Delamerian orogenic events are recorded in southeastern South Australia, where Turner et al. (2009) found evidence for the onset of orogenesis at 545 Ma. However deformation in the region associated with this orogen became more prominent about 30 m.y. later, with the intrusion of the Rathjen Gneiss (ca 514 Ma) in South Australia and felsic dykes in the Glenelg Zone in far western Victoria (Foden et al., 1999; Ireland et al., 2002). Thus the main Delamerian deformation events broadly overlap in time with those of the Tyennan Orogeny and are part of the establishment of a much larger orogenic system, the Terra Australis Orogen, along the Gondwanan margin (Cawood, 2005).

The early Delamerian-Tyennan Orogeny was more prominent in Tasmania, where Berry et al. (2007) recorded a U/Th monazite age of 513±8 Ma and Black et al. (1997) a U/Pb SHRIMP age of 514±6 Ma for the metamorphism of the Forth Metamorphic Complex. Near the eastern edge of the outcropping Proterozoic rocks, the Settlers Schist, possibly a deformed 660 Ma granite (unpublished data by Black, 2007), has metamorphic monazites with an age of 517±9 Ma (Reed et al., 2001; Berry et al., 2007). Tectonic mélanges preserved on the north coast may also have been emplaced at about 515 Ma (Seymour and Vicary, 2010). Further west, westward obduction of forearc-derived mafic-ultramafic igneous rocks began at about this time (Crawford and Berry, 1992; Black et al., 1997; Stacey and Berry, 2004). This west-directed obduction has been used to propose an east-dipping subduction zone (Crawford and Berry, 1992).

In the western and central parts of Tasmania, including the Arthur Lineament, ages from monazite dating show that the peak of blueschist or amphibolite facies metamorphism in the Tyennan Orogeny was at about 510 Ma (Meffre et al., 2000; Turner and Bottrill, 2001; Berry et al., 2007). In the Franklin Metamorphic Complex, on the northwestern edge of the Tyennan Block, temperatures reached 650°C and pressures up to 1.9 GPa (Chmielowski, 2009). Three main phases of Tyennan deformation are recorded in the Arthur Lineament; two generations of isoclinal folding that indicate south-directed tectonic transport, whilst the third generation is dominantly of steep west-dipping thrust faults (Holm and Berry, 2002). Reed et al. (2002) recorded similar D_1 and D_2 but more intense D_3 deformation in the Badger Head Block in central northern Tasmania. Near the southern end of the Badger Head Block, Reed et al. (2002) recorded the presence of what seemed to be retrogressed garnet, suggesting that at least part of the Badger Head Block may have reached amphibolite facies conditions. Elsewhere metamorphism is mostly greenschist facies (Reed et al., 2002). Stacey and Berry (2004) interpreted that the initial Tyennan deformation was caused by collision from the northeast.

In Tasmania, the high-K calcalkaline andesitic to felsic Mt Read Volcanics and associated sedimentary rocks were emplaced into synorogenic regional graben at about 500 Ma (Black et al., 1997; Stacey and Berry, 2004). Stacey and Berry (2004) attributed this regional extension to north-south shortening in a second phase of the Tyennan Orogeny. At the same time, in Victoria similar volcanic and sedimentary rocks were being emplaced in the Stavely Belt and associated Dimboola Subzone in far western Victoria and in the Jamieson and Licola areas, windows through the Melbourne Zone (VandenBerg et al., 1995; Stuart-Smith and Black, 1999; Crawford et al., 2003a; Spaggiari et al., 2003b). Because the Cambrian volcanic outcrops in both Victorian regions are surrounded by sedimentary rocks to which they bear few relationships, the detailed settings of these volcanic rocks are less well known in detail (Crawford et al., 2003a). However, Crawford (2003a) considered that because the geochemistry of the volcanic rocks was so similar to the Mt Read Volcanics, it was likely that the general tectonic settings were also similar, with magmatism occurring on pre-existing continental crust.

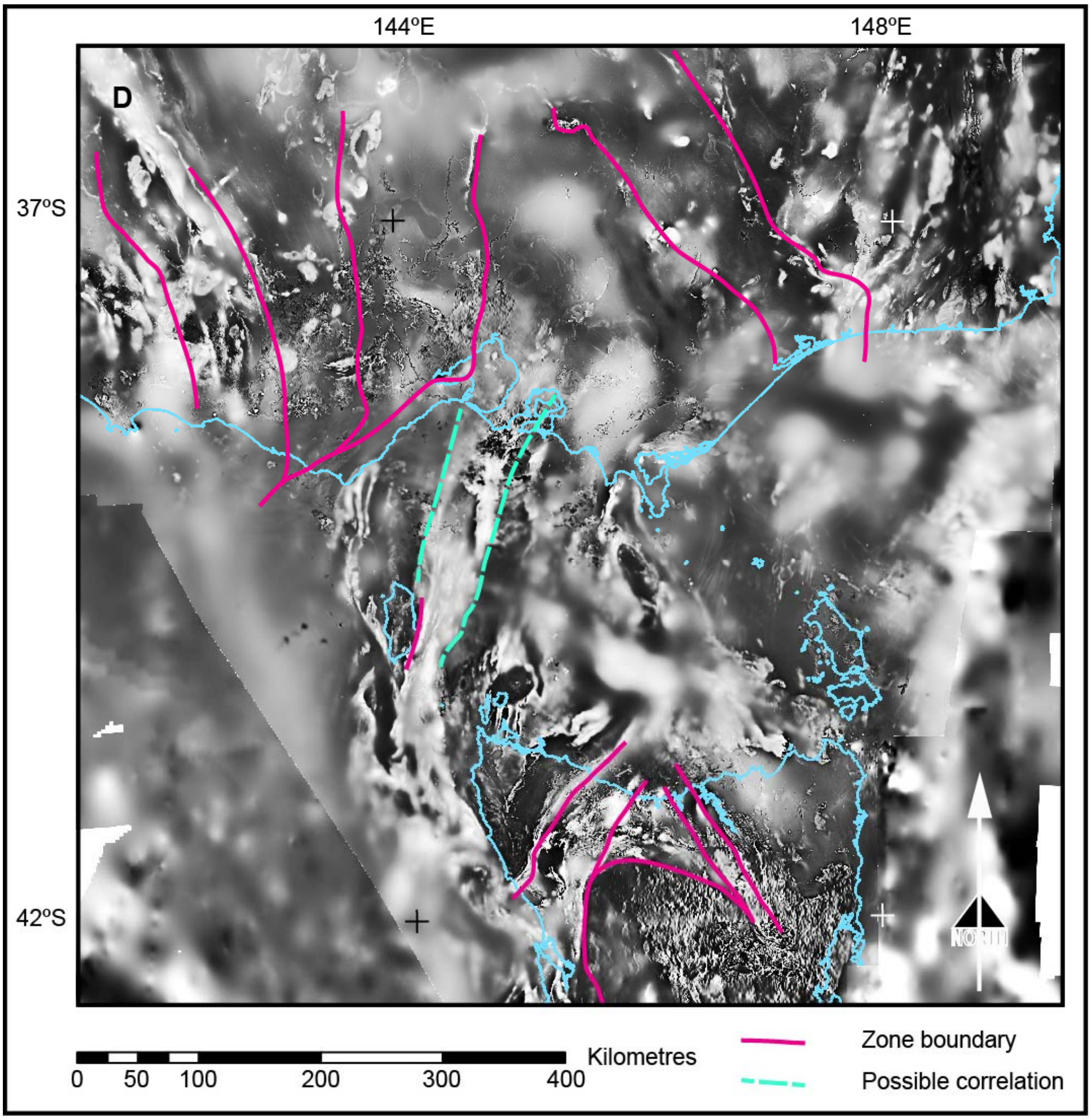


Figure 2-4. Total Magnetic Intensity image of the Bass Strait region overlain with the zone boundaries from Figure 2-1. White shows magnetic highs, black magnetic lows. D marks the Dimboola Subzone magnetic high. Magnetic data from the Department of Primary Industries (Victoria) and Geoscience Australia.

The effects of the Delamerian-Tyennan Orogeny after 500 Ma are widely recognized in western Victoria. The high temperature-low pressure Glenelg River Metamorphic Complex and associated plutons are present and several dating methods, summarized in Morand et al. (2004), suggest that peak metamorphism took place at about 495 Ma and post-tectonic intrusions were emplaced until about 490 Ma. Its eastern boundary is the Yarramyljup Fault (Figure 2-1), which separates the greenschist and low-pressure high-temperature amphibolite grade metasedimentary rocks from low greenschist or prehnite-pumpellyite grade sedimentary and volcanic rocks of the Grampians-Stavely Zone to the east (VandenBerg et al., 2000; Morand et al., 2004).

Although it does not crop out, the Dimboola Subzone is an important part of the Grampians-Stavely Zone. It is associated with a large magnetic and gravity high (Figures 2-4 and 2-5) that Direen and Crawford (2003a) interpreted as being caused by Ediacaran rift-related seaward-dipping magnetic basalts, whilst others have interpreted it as a buried island arc (Finn et al., 1999; Moore, 2006). The potential field expression of the Dimboola Subzone is approximately 250 km long and about 25 km wide and is covered by Silurian and Cenozoic sedimentary successions. Rare drill holes into the magnetic and gravity high intersected gabbro and felsic volcanic rocks (Knight et al., 1995; O'Neill, 1995) that VandenBerg et al. (2000) interpreted as a tholeiite-boninite sequence.

The Moyston Fault lies about 70 km east of the Yarramyljup Fault. It forms the boundary between the Grampians-Stavely Zone and the Stawell Zone (Figures 2-1 and 2-2). It is also the western boundary of the Lachlan Orogen (Cayley and Taylor, 2001; Miller et al., 2005; Cayley et al., 2011). In the hangingwall, the tectonic mélange of metasedimentary rocks and mafic metavolcanics have been metamorphosed to amphibolite facies; P-T equilibria suggest burial to about 15 to 20 km depth (0.7 to 0.8 GPa) at temperatures of 570±20°C, the deepest levels known to crop out in the Lachlan Orogen (Phillips et al., 2002). Seismic interpretations show that the fault dips east at approximately 45° (Korsch et al., 2002; Cayley et al., 2011). $^{40}Ar/^{39}Ar$ age dates on the micas present give an age of exhumation of 500 to 490 Ma, the age of the Delamerian and latest Tyennan Orogenies (Miller et al., 2005). The Moyston Fault is an important boundary at depth as it marks a change in Re-Os ages in mantle xenoliths, in the initial $^{87}Sr/^{86}Sr$ ratios in Pliocene to Holocene Newer Volcanic Group, and in teleseismic character (Price et al., 1997; Handler and Bennett, 2001; Graeber et al., 2002). Taylor and Cayley (2000) argued that the fault acted as a Type 1 backstop in the Tyennan-Delamerian and subsequent orogenic events in the region. In the outcropping areas, shortening was generally close to northwest-southeast, but magnetic and gravity data further north suggests that regionally the shortening was east-west and verging westward.

In Tasmania, the Owen Conglomerate records sedimentation during the late Delamerian-Tyennan Orogeny into evolving half-graben (Crawford and Berry, 1992; Seymour and Calver, 1998; Noll and Hall, 2005; Reed and Vicary, 2005). These deposits are widespread through much of western and central Tasmania east of the Arthur Lineament (Seymour and Calver, 1998; Noll and Hall, 2005). Equivalent deposits in Victoria are less obvious, although the time-equivalent Bear Gully Gritstone and the Wahroonga Breccia, both in the Melbourne Zone, have been correlated with the Owen Conglomerate (Cayley et al., 2002; VandenBerg et al., 2006).

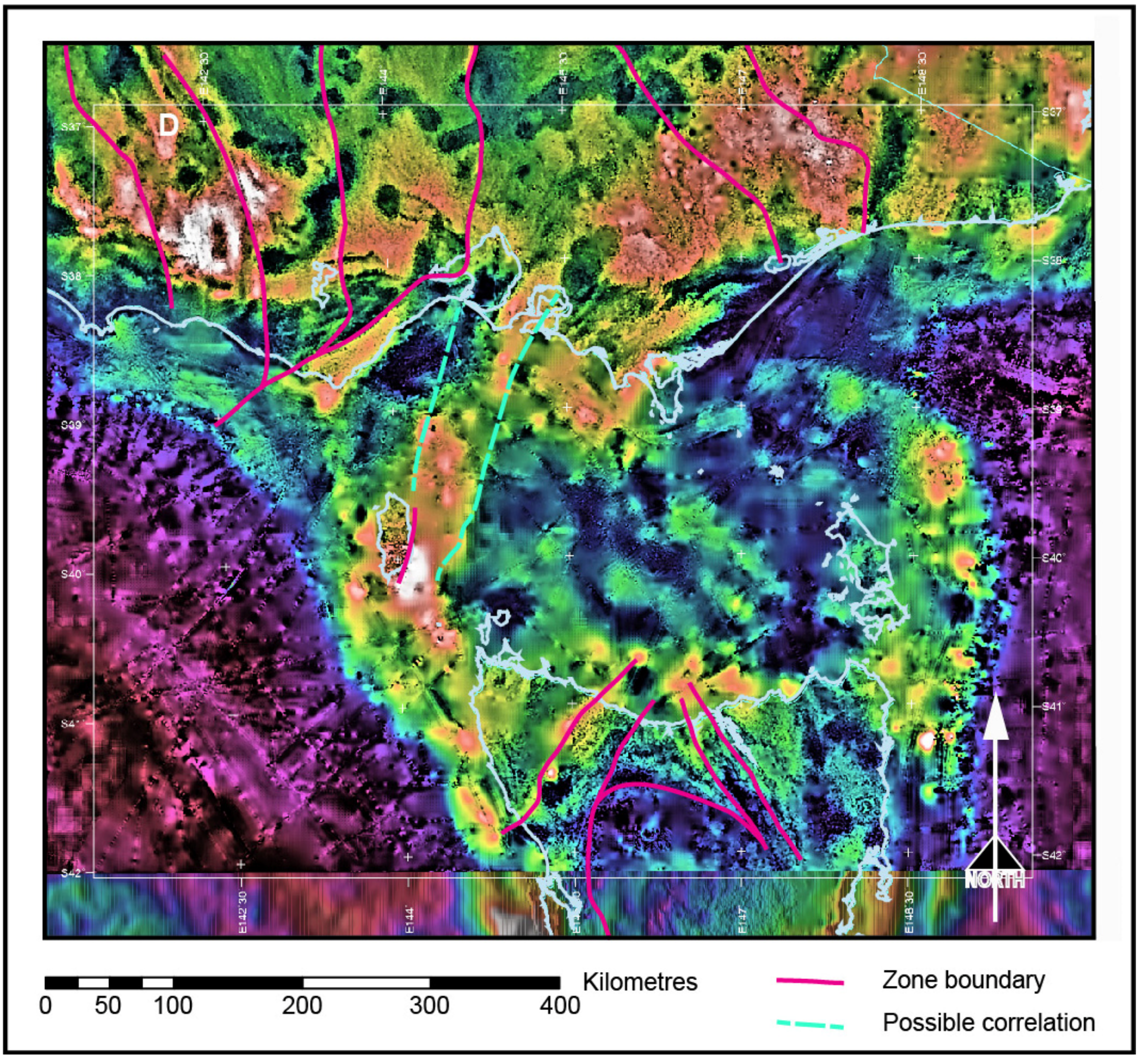


Figure 2-5. Isostatic gravity of the Bass Strait region. White and red indicate highs, blue and magenta lows. The zone boundaries are as for Figure 2-1. B shows the location of the Bambra Fault. Offshore data are of variable quality. D marks the Dimboola Subzone gravity high. Data from Department of Primary Industries (Victoria) and Geoscience Australia.

In Victoria, the eastern boundary of Delamerian-Tyennan deformation is not well defined, but is probably the Avoca Fault (Glen, 2005; Miller et al., 2005). To the east of the Avoca Fault, the sedimentary rocks present are Ordovician or younger. Using $^{40}Ar/^{39}Ar$ ages to date the metamorphism in the hangingwall, Miller et al. (2005) showed that the Delamerian-Tyennan deformation continued east of the Moyston Fault, and there is no other obvious structural break than the Avoca Fault. In northern Tasmania, the eastern boundary is in the west Tamar region, at the western edge of the Andersons Creek Ultramafic Complex (Reed et al., 2002). However we note the interpretation of seismic data from the central and southern east Tasmanian coast by Drummond et al. (2000) that showed a crustal unit with higher velocities below depths of from about 9 to 18 km (3 to 6 s two-way-travel time). They described the lower package as "suggestive of tilted fault blocks" and interpreted this to be the result of the extension of Proterozoic crust, on which Ordovician to Devonian sediments had been deposited (Drummond et al., 2000).

As a result of these orogenic events, an extensive turbidite succession, Coney et al.'s (1990) 'Ordovician mud-pile', was deposited throughout much of eastern Australia and adjacent regions (Ireland et al., 1998; Squire and Wilson, 2005). This mud-pile includes the Cambrian St Arnaud Group in the Stawell Zone, and the Ordovician metasediments in the Castlemaine Supergroup in the Bendigo Zone, the Adaminaby and Bendoc groups in the Tabberabbera Zone, all in Victoria, and the Tippogoree Group in the Mathinna Supergroup in eastern Tasmania (Table 2-1)(VandenBerg et al., 2000; Seymour et al., 2011). Two areas had no turbidite influx—the Melbourne Zone and western Tasmania. In western Tasmania and southernmost central Victoria, the Gordon Group limestones and equivalent shelf sequences are widespread (Burrett and Martin, 1989; Seymour and Calver, 1998; VandenBerg et al., 2000). In the northern and central Melbourne Zone, Early Ordovician thin cherts are sometimes visible in structural windows and Late Ordovician black shale is present where faults expose the deeper parts of the sequence (VandenBerg et al., 2006).

The Benambran Orogeny took place in the Late Ordovician and Early Silurian. In the Bendigo Zone deformation was restricted to upright folding with north-south axes and mostly east-verging faulting (VandenBerg et al., 2000; Cayley et al., 2011). Metamorphism was mostly sub-greenschist facies, although transitional blueschist rocks that formed at less than 450° C and 15 to 21 km are present in the Heathcote Fault Zone, at the eastern margin of the Bendigo Zone (Spaggiari et al., 2002; Wilson et al., 2009).

$^{40}Ar/^{39}Ar$ dating of the reconstituted micas and alteration products of the widespread gold mineralization at that time give ages of 460 to 440 Ma (Foster et al., 1999; Bierlein et al., 2001a; Fu et al., 2009).

In the Tabberabbera Zone, the Ordovician turbidite sequence appears conformable with the overlying Silurian Cobbannah Group (VandenBerg et al., 2000). There is also evidence of earlier deformation in the Wonnangatta Fault Zone area, where abundant tectonic mélange is present in the Pinnak Sandstone (Fergusson, 1987; VandenBerg et al., 2000; Willman et al., 2005). The mélange tends to confirm suggestions by Collins and Vernon (1992) and Gray and Foster (1998) that the deformation began early in the eastern Lachlan Orogen, whilst it was still in a deep marine setting. Paleontological evidence and age dates on the Omeo Metamorphic Complex, the metamorphosed equivalents of the Tabberabbera Zone, suggest that the main phase of the Benambran Orogeny took place at about 430 Ma (VandenBerg et al., 2004; Willman et al., 2005). Deformation verges towards the west, particularly west of the Wonnangatta Fault Zone (VandenBerg et al., 2000). Metamorphism associated with the Benambran Orogeny generally grades from prehnite-pumpellyite facies in the west to greenschist facies in the east, with transitional blueschist facies rocks present on the western margin (Spaggiari et al., 2002; Morand and McKnight, 2006).

In eastern Tasmania, the Tippogoree Group also appears to have been deformed in the Benambran Orogeny, with east-verging recumbent folds mapped (Reed, 2001; Seymour et al., 2011). These appear to lie below a faulted and folded unconformity (Seymour et al., 2011). Bierlein et al. (2005) also found evidence of sericite formation at about 425 to 430 Ma in $^{40}Ar/^{39}Ar$ age spectra from wall rocks in the Lefroy and Beaconsfield goldfields.

Neither folding nor faulting related to the Benambran Orogeny have been recognized in western Tasmania or in the Melbourne Zone. In western Tasmania, the only change in the Silurian was to more clastic, but still shallow water, sedimentation (Stacey and Berry, 2004). In the Melbourne Zone, provenance and current direction studies show that most sedimentary input came from sources to the south (Powell et al., 2003). There was no input from the east until the Emsian, despite the Tabberabbera Zone rocks now just to the east having been deformed in the Silurian Benambran Orogeny. This suggests that the two zones were not in proximity until the Devonian (Fergusson et al., 1986; VandenBerg et al., 2000; Cayley et al., 2002). The Early Silurian to the Middle Devonian evolution of the Melbourne Zone was dominated by deep water turbidite sedimentation, with a significant Silurian input of coarse, immature clastics at about 430 Ma and a finer grained sand pulse at about 420 Ma (VandenBerg et al., 2000). However in the southeastern Melbourne Zone, Devonian limestones unconformably overlie the basal Ordovician limestone. In the west, shallow marine, mostly clastic, sediments are present from the Late Silurian (VandenBerg et al., 2000).

In eastern Tasmania, the Panama Group, the upper part of the Mathinna Supergroup, was deposited from the Silurian to Early Devonian (Seymour et al., 2011). Like the Melbourne Zone, Seymour et al. (2011) described two sandier pulses of sedimentation, although unlike the Melbourne Zone, both pulses are mature. The older pulse has poor age control, but the younger one contains Early Devonian plant fossils and conformably overlies a siltstone containing a Ludfordian graptolite fauna, indicating deposition at approximately 420 Ma (Seymour et al., 2011). In western Tasmania, marginal marine clastic sediments were deposited. These grossly grade from sandstones in the Silurian to shales in the Early Devonian (Seymour and Calver, 1998). As in the Melbourne Zone, there were two sand pulses, one at about 430 Ma and the other near the Silurian-Devonian boundary, or about 416 Ma (Seymour and Calver, 1998).

The Bindian Orogeny took place in the Late Silurian and Early Devonian, at about 420 Ma. In the Tabberabbera Zone, dextral strike-slip faulting and steeply plunging folds are most prominent closest to the adjacent Omeo Zone (Willman et al., 2005). The faults strike southeast in the northern part of the zone, but progressively strike towards the east and lose displacement in the southern part (Willman et al., 2005). The Tabberabbera Zone also underwent north-south shortening late in the Bindian Orogeny (VandenBerg et al., 2000). There are no recorded deformation effects in Tasmania or in the Victorian Western Lachlan Orogen.

The Early Devonian Buchan Rift and Mitchell Syncline unconformably overlie the southeastern Tabberabbera Zone. Both were deposited in shallower marine conditions than the older rocks; they also contain volcanic deposits (VandenBerg et al., 2000). The Buchan Rift contains significant amounts of limestone, whilst the Mitchell Syncline mostly contains clastic deposits with variable amounts of calcareous cement, although there are also minor limestones (Orth et al., 1995; VandenBerg et al., 2000). The volcanism coincided with granitic intrusions in eastern Victoria. This igneous activity started immediately after the Benambran Orogeny, at 426±7 Ma, and continued through to the Tabberabberan Orogeny, from 380 to 400 Ma (VandenBerg et al., 2000; Fanning and Morand, 2002; Morand and Fanning, 2006, 2009). In the Bendigo, Stawell and the eastern Grampians-Stavely zones in western Victoria, U/Pb zircon ages of granites cluster at 410 to 400 Ma, some 40 m.y. younger than the Benambran Orogeny, the last major deformation in the region (VandenBerg et al., 2000).

In western Tasmania, much of the deformation in the Tabberabberan Orogeny was taken up by reactivation of pre-existing structures (Stacey and Berry, 2004). Where this did not take place, early north-northwest trending folding preceded northwest to north-northwest folding and thrusting (Seymour et al., 2007). In eastern Tasmania east of the Tamar River, only steeply dipping axial planes and dominantly shallowly southeast plunging folds in the Silurian to Early Devonian sedimentary rocks are seen (Seymour et al., 2011). Further east, Patison et al. (2001) mapped east-directed thrusting in similar aged rocks. They proposed that the entire Mathinna Supergroup in northeastern Tasmania was part of a west-verging recumbent fold system that subsequently was overprinted by upright folding and cut by east-verging back thrusts.

The Melbourne Zone was also extensively deformed in the Tabberabberan Orogeny. In the west, open folds are present, there is little cleavage developed and the dominantly east-west shortening is about 34% (Foster and Gray, 2007). Deformation increases towards the east and verges eastwards, so that in the easternmost parts, folding is recumbent to isoclinal, cleavage is intensely developed, and the northeast-southwest shortening is at least 70% (VandenBerg et al., 1995; Foster and Gray, 2007). In the northwestern Melbourne Zone, there is also late north-south shortening, giving rise to east-west faults and a dome and basin folding pattern (Gray and Mortimer, 1996). Further east, in the Buchan Rift, the Tabberabberan Orogeny also resulted in east-west shortening (Orth et al., 1995). Pre-Tabberabberan faults were reactivated, and new generally north-south reverse and thrust faults and the Murrindal Synclinorium formed (Orth et al., 1995).

The earliest post-Tabberabberan intrusive events seem to have taken place in eastern Tasmania and on Flinders Island, where foliated granites have U/Pb zircon ages of 399±1.5 Ma and 401±4 Ma (Black et al., 2005; Black et al., 2010). Similar ages, 398±3 Ma, have been recorded from a sheared granite in eastern Victoria (Morand and Fanning, 2006) and from Wilsons Promontory, 395±4 Ma (Elburg, 1996b). This is coeval with gold mineralization in northeast Tasmania and in the Tabberabbera Zone in Victoria (Bierlein et al., 2005; Willman et al., 2005). In northeastern Tasmania, granites were intruded until 376.5±3 Ma (Black et al., 2005). In western Tasmania, the oldest pluton has a U/Pb zircon age of 374±2 Ma with the youngest, 360±2 Ma on Tasmania and 351±2 Ma on King Island (Black et al., 2005; Black et al., 2010). Both S- and I-types are present in both regions. In Victoria, post-Tabberabberan granites are restricted to the Melbourne and eastern Bendigo zones and adjacent areas and in the far east (VandenBerg et al., 2000). In the Melbourne Zone, the earliest phase is preceded by an extensive mafic dyke suite and by gold mineralization (VandenBerg et al., 2000; Bierlein et al., 2001a). Elsewhere there is a dearth of U/Pb zircon ages, but where available typically indicate intrusive ages of about 370 Ma (summarised in VandenBerg et al., 2000; Bierlein et al., 2001a). Other less reliable methods with lower closure temperatures are generally consistent with this (VandenBerg et al., 2000). Both S- and I-types are present; they are both strongly reduced and have high Ba contents when compared to the granites hosted in adjacent zones (White, 2002; Rossiter and Gray, 2008).

2.2.2 GEOPHYSICAL DATA

Potential field data

Since the mid-1990s there has been an intensive program by the Victorian, Tasmanian and Commonwealth governments to conduct detailed airborne magnetic surveys and ground gravity surveys. As a result, large onshore areas have been covered by 200 m to 400 m magnetic surveys and offshore by 400 m to 800 m surveys, which have then been stitched into an Australia-wide grid by Geoscience Australia (Milligan et al., 2010).

This has allowed much more confident correlation of magnetic units in the region than the pre-existing data that was either digitized from analogue data collected in the early

1960s at up to 8 km line spacing or was nonexistent (Balfour, 1968). Ground gravity surveys range from a nominal 0.5 km to 15 km station spacing and ship-borne gravity data have also been widely collected and routinely integrated by Geoscience Australia into a single dataset (Nakamura et al., 2011). The new magnetic and gravity data now provide excellent tools for extrapolating units beneath Bass Strait (Figures 2-4 and 2-5) and are used in this synthesis to test the various hypotheses that have tried to explain the geological paradoxes relating to the Gondwanan margin between northern Tasmania and mainland Australia.

A strongly magnetic package of Ediacaran rift tholeiites crops out on the southeastern edge of King Island (Calver, 2007). Direen and Crawford (2003b) used regional magnetic data to forward model this package to be approximately 20 km thick and with easterly dips of between 50° and 70°. The new data show that this package continues northwards to the interpreted Cambrian basalt on the south coast of Phillip Island in southern Victoria (Henry and Birch, 1992). Regional gravity compilations show a gravity high coincident with the magnetic high. Further east, McLean et al. (2010) used the new magnetic grid to compare the magnetic anomalies sourced from under the eastern Melbourne Zone with the western Tasmanian ophiolites. In eastern Tasmania, near Piper River, Roach and Leaman (1996) used the regional magnetic and gravity data to interpret a west-verging thrust of an ophiolite package. The new magnetic grid shows that the ophiolite continues northwards offshore for at least 50 km.

Seismic data

Along the northern Tasmanian coast, Barton (1999) interpreted east-dipping reflectors in both the east and the west, and attributed most of them to west-directed thrusting. Where east- and west-dipping structures were present, the west-dipping structures were truncated by younger east-dipping structures. The thrust interpreted by Roach and Leaman (1996) may also have been imaged in the seismic data as a west-verging thrust stack at about 5 km depth (Barton, 1999). On the eastern Tasmanian coast, Drummond et al. (2000) interpreted that in the south, the Mathinna Supergroup was underlain by half-graben formed from Proterozoic crust, possibly the Tyennan Block. Further north, the Mathinna Supergroup appeared to be underlain by oceanic crust (Drummond et al., 2000).

There have been two major seismic surveys in Victoria. In 1997, seismic data were collected over the Moyston Fault and nearby areas in western Victoria. The data showed that the Moyston Fault dipped to the east (Korsch et al., 2002). In 2006, seismic data were collected from the eastern end of the 1997 survey (i.e. just east of the Moyston Fault) and extended into central Victoria to cover the Stawell, Bendigo, and Melbourne zones, finishing in the western Tabberabbera Zone (Cayley et al., 2011). The data showed that the deep crust under the Bendigo Zone was an imbricated series of west-dipping strongly reflective layers, suggesting that the Cambrian ocean floor basalt had shortened by stacking when the upper sedimentary rocks were shortening by folding in the Benambran Orogeny (Cayley et al., 2011). The Moyston Fault truncated the Avoca Fault, with the Cambrian rocks of the Stawell Zone emplaced over the Ordovician metaturbidites of the western Bendigo Zone. As well, the deep crust under the Melbourne Zone was characterized by strong sub-horizontal layering and a well-developed Moho unlike that seen elsewhere in the survey (Cayley et al., 2011).

Seismic tomography

Graeber et al. (2002), Rawlinson and Kennett (2008) and Rawlinson et al. (2010) used seismic tomography to define deep boundaries in Victoria and Tasmania. In western Victoria, Graeber et al. (2002) showed the Moyston Fault extended to a depth of 300 km. In Tasmania, Rawlinson et al. (2010) showed the Tamar Fracture Zone beneath the Moho was approximately 50 km to the east of its surface expression. In northwestern Tasmania, Rawlinson et al. (2010) showed that there was a significant east-dipping anomaly to the east of the Rocky Cape Block, which they interpreted as the relic of the Cambrian east-dipping subduction zone. The seismic tomography data also showed a step in the crust, with the Moho below most of the Dundas Trough approximately 5 km shallower than the adjacent Moho (Rawlinson et al., 2010).

2.3 PREVIOUS RESOLUTIONS

The early attempts to resolve the differences in geology between Tasmania and mainland Australia have focused on moving Tasmania either east or west along east-west faults either under Bass Strait or the Otway and Gippsland basins. Examples include Stump et al. (1986), Chappell et al. (1988) and Veevers and Eittreim (1988), who all hypothesized that Tasmania was positioned about 150 km east of its present position in the Mesozoic and moved to its present position during the Gondwanan breakup event. By contrast, Elliott and Gray (1992) suggested movement of 145 km to the southwest to make the space required for the Bass Basin fill.

In another fault-related model, Bierlein et al. (2005) argued that that the ages of metamorphic and hydrothermal micas in and around gold mineralization in Victoria and northeast Tasmania could be used to align the rock packages. The $^{40}Ar/^{39}Ar$ ages from the turbidite-hosted gold deposits there range from 385 to 395 Ma, whilst those from within the alteration areas of some mines tend to show ages of 425 to 430 Ma. Although the Victorian Bendigo Zone gold deposits formed above oceanic crust, they considered that the Tasmanian deposits formed above a (then) eastwards promontory of late Neoproterozoic crust from Gondwana that lay ‘south’ of the Victorian gold deposits. Since the ages of the deposits were close to late stage mineralization in the Bendigo Zone, they suggested that the Tasmanian deposits may have been along strike at that time and subsequently faulted eastwards on a sinistral strike-slip transform fault.

Powell and Baillie (1992) compared the Mathinna Supergroup and sequences in Victoria. On the basis of similar lithologies, north-directed paleocurrent directions and tectonic histories, they interpreted that the closest correlative was the Melbourne Zone. In order to align the two regions they proposed a southwest-striking transfer fault west of Wilsons Promontory in southern Victoria and across Bass Strait. Subsequently, Powell et al. (1993) suggested that the two regions might not be directly connected, but rather en échelon basins opened by the same large scale transtensional tectonic events. In this way, they resolved the issue that the Melbourne Zone verged to the east whilst the Mathinna Supergroup verged to the west. Patison et al. (2001) accepted Powell and Baillie’s (1992) direct correlation. Using new mapping and newly collected illite crystallinity data, they interpreted the outcropping Mathinna Supergroup as forming the lower limb of a west-verging nappe that had been repeatedly backthrust eastwards and with the lower detachment shallowing eastwards. This is somewhat similar to the detachment model for the Silurian and Devonian Melbourne Zone proposed by VandenBerg et al. (2000) and Cayley et al. (2002).

Cayley et al. (2002) proposed that the Selwyn Block was the Tasmanian pre-Ordovician crust that extended northward under Bass Strait to form the substrate to the Melbourne Zone. The reasons for this included:

a. Magnetic trends seem to cross Bass Strait from Tasmania to Victoria, and these trends seem to be reflected in the present topography in southern Victoria,
b. At Cape Liptrap, in southeastern Victoria, Early Ordovician limestone is present just above an unconformity, suggesting correlation with the Gordon Group in Tasmania. In Victoria the unconformity is marked by a thin chert layer containing quartz granules and grit, which was correlated with parts of the Owen Conglomerate in western Tasmania,
c. Western Tasmania and the Melbourne Zone were the only parts of the Lachlan Orogen that were not affected by the Benambran Orogeny,
d. The model could account for the Ceres Metagabbro, near Geelong, picritic basalts on Phillip Island (Figure 2-1), and the Cambrian felsic volcanic rocks in windows in the Melbourne Zone,
e. The granites in and around the Melbourne Zone have a different isotopic signature, and the many S-type granites are cordierite-bearing (suggesting derivation from continental crust). Both characteristics are different to the adjacent granites, suggesting a different source composition (Chappell et al., 1988; Gray, 1990),
f. The Melbourne Zone is a region of vergence reversal. To the west, in the Bendigo Zone, the thrusts verge east towards the Melbourne Zone. To the east, in the westernmost Tabberabbera Zone the thrusts verge west, also towards the Melbourne Zone, and
g. Both the western margin, the Heathcote Fault Zone, and the eastern margin, the Governor Fault, include outcrops of transitional blueschist facies metamorphic rocks (Spaggiari et al., 2002).

This correlation was carried further by McLean et al. (2010), who specifically correlated the magnetic signatures associated with the obducted ophiolites in western Tasmania with the deep responses under the Melbourne Zone.

Other authors have implicitly or explicitly correlated the Mt Read Volcanics in Tasmania with the Stavely Volcanic Group in Victoria. Scheibner and Veevers (2000) interpreted the Stavely Group and Mt Read Volcanics as direct correlatives that occupied the same relative positions as now and that were deposited in the same tectonic position. The Cambrian Jamieson and Licola volcanic rocks were correlated with similar rocks in the Mt Wright arc in far western New South Wales and were considered to be part of their interpreted 'Victorian Microcontinent', much further away from Tasmania than their present position. This microcontinent also included the Mathinna Supergroup. (Scheibner and Veevers, 2000).

Flöttmann et al. (1998) suggested that the Cambrian Kanmantoo Trough was bounded to the south by a left-lateral strike-slip fault, south of which was Antarctica, the source of the sediments that filled the trough. They interpreted this fault from the seismic data gathered to the southeast of Kangaroo Island, and implied that it may have continued south of Tasmania. This interpretation required the subduction in both western Victoria and Tasmania to be east-dipping. Foden et al. (2006) used new U/Pb zircon and titanite and $^{40}Ar/^{39}Ar$ ages to compare timings in the Ross Orogen in Antarctica with events in South Australia and western Victoria. They suggested that the 514±4 Ma age of the syntectonic Rathgen Gneiss and the 513±0.8 Ma titanite age of the Bungadillina Monzonite were the minimum age of the Delamerian-Tyennan Orogeny in South Australia. This overlapped with early ages seen in western and northern Tasmania, for example 510±3 Ma in the Heazlewood Tonalite. On Kangaroo Island, in the westernmost parts of the Delamerian Orogen, deformation ceased between 500 and 504 Ma, whilst to the east in western Victoria it continued until between 493±8 Ma and 491±8 Ma (Foden et al., 2006). This temporal relationship was used to interpret west-dipping subduction driven by ridge push, and terminated with slab break-off. Subsequent slab rollback resulted in the post-tectonic granites in southeastern South Australia and western Victoria (Foden et al., 2006).

Cayley (2011) proposed a detailed history of south eastern Australia and North Victoria Land from the Middle Cambrian (530 Ma) to the Early Silurian. He suggested that Tasmania, its adjacent submarine plateaus and the Selwyn Block were one entity in the Cambrian. This micro-continent was termed VanDieland and was positioned in the Proto-Pacific during the Early Cambrian. An east-dipping subduction zone was off the eastern edge of Gondwana and a west-dipping subduction zone was active along the immediate Gondwanan margin. VanDieland began colliding with the Ross Orogen section of Gondwana at about 515 Ma, the Ross-Tyennan-Delamerian Orogeny, giving rise to metamorphism in western Victoria and Tasmania, whilst the east-dipping subduction zone caused the emplacement of the ophiolite slices in western Tasmania (Crawford and Berry, 1992; Cayley, 2011). From 495 to 500 Ma, there was a single zone of felsic volcanism extending from northernmost Victoria Land through the Mt Read Volcanics to the Jamieson-Licola region and thence to the Stavely Volcanic Group, which at this time was along strike from central Victoria. In the Early Ordovician, VanDieland migrated away from Gondwana, drawn northeast by a north-dipping subduction zone outboard of Gondwana (Cayley, 2011). This allowed the subduction zone outboard of the Ross Orogen to re-establish. By 470 Ma, in the Middle Ordovician, VanDieland had travelled into the proto-Pacific, about 350 km to the northeast of Victoria Land and generally east of its modern-day position with respect to western Victoria (Cayley, 2011).

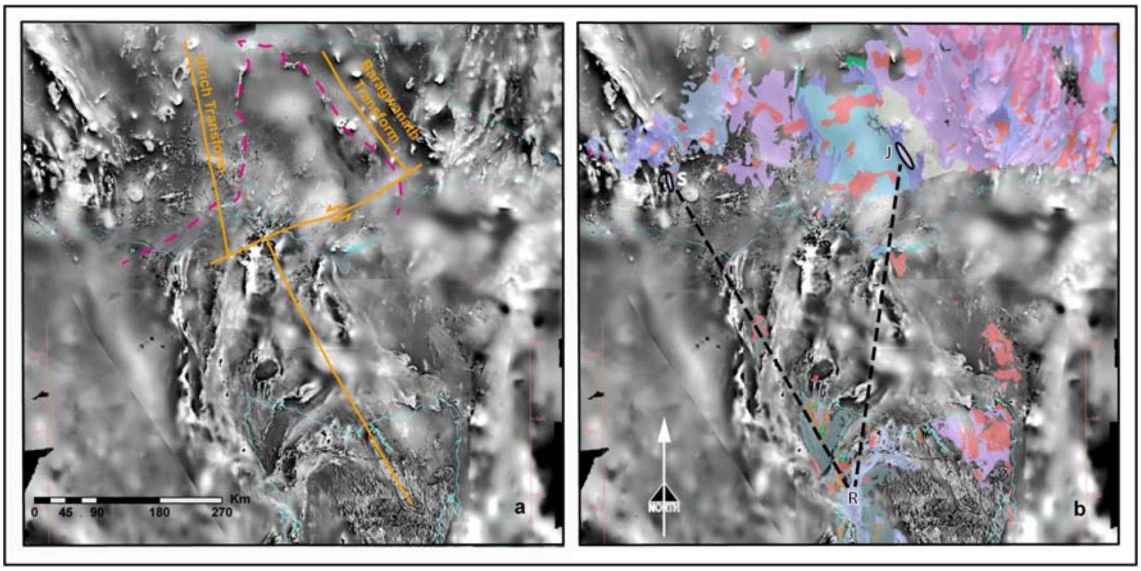


Figure 2-6. a. Rossiter and Grey's (2008) correlation between Victorian and Tasmanian granites (orange), shown with the total magnetic intensity as a background. The near-surface outline of the Selwyn Block as proposed by Cayley et al. (2002) is shown as a dashed magenta line. There is no evidence for their Ulrich Transform or the NE-striking sinistral fault in the magnetic data. b. Possible correlations of the Cambrian Mt Read Volcanics in Victoria. Pre-Permian geology shown with the same legend as for Figure 2-1. R, Mt Read Volcanics, S, Stavely Volcanic Group, J, Jamieson-Licola area. The Mt Read-Jamieson-Licola correlation is much more likely since they are along the same general magnetic trends, whereas correlation with the Stavely Volcanic Group crosses several distinct magnetic units.

Chappell et al. (1988) and Rossiter and Gray (2008) looked at the chemistry of the granites in the Bass Strait region. Whilst the granites of southernmost Victoria have very similar mineralogy and geochemistry to those of Tasmania and Bass Strait, those further north in the 'Melbourne Terrane' of Chappell et al. (1988) have significant differences, and so they argued that the basement must have changed. Rossiter and Gray (2008) noted that the granites in central Victoria (their broadly similar Central Victorian Superprovince) had a distinctly higher barium content, lower initial $^{87}Sr/^{86}Sr$ ratios and lower oxidation states than the granites further east or west and also inferred that this was in response to a different basement in central Victoria. The high barium contents implied that the source rocks were feldspar-rich and the initial $^{87}Sr/^{86}Sr$ ratios indicated that the source was probably no older than 550 Ma. The western boundary to their Central Victorian Superprovince broadly coincided with the Muckleford Fault in the Bendigo Zone. Rossiter and Gray (2008) placed the eastern boundary along the Wonnangatta Fault, slightly further east than the Selwyn Block edge. They hypothesized that the Muckleford Fault formed a major transform, the Ulrich Transform, during the ~420 Ma Bindian Orogeny. The Muckleford Fault was correlated with the Tamar-Tiers Lineament via a northeast-trending cross-fault with a displacement of about 50 km through southern Victoria. Rossiter and Gray (2008) concluded that the Selwyn Block was an exotic fragment emplaced late in the Bindian Orogeny.

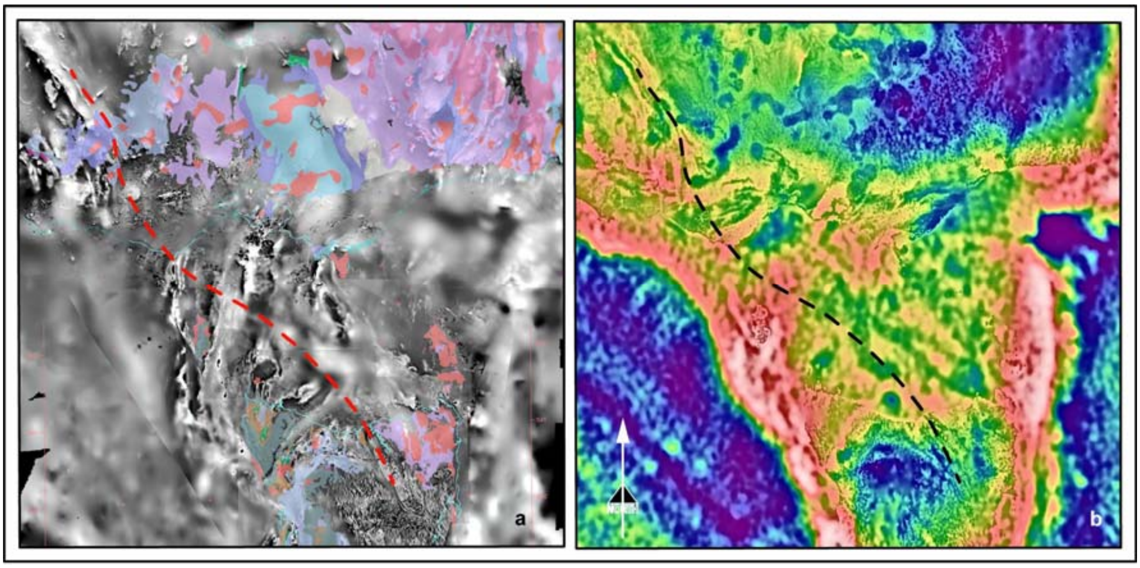


Figure 2-7. Teasdale et al.'s (2003) correlation of the Moyston Fault with the Tamar-Tiers Lineament. Correlation is shown as a red dashed line in a, an image of the total magnetic intensity overlain by the pre-Permian geology and a black dashed line in b, the Bouguer gravity onshore and free air gravity offshore. Neither the magnetic or gravity data offer support to this correlation.

The Tamar Fracture Zone, which is broadly the eastern boundary of outcropping Proterozoic rocks in Tasmania, has also been correlated with the Moyston Fault in western Victoria (Teasdale et al., 2003). This correlation principally relied on the assertion of deeply buried Proterozoic rocks under the Grampians-Stavely and Glenelg Zones in western Victoria and in western Tasmania, and that the Tamar Fracture Zone was generally along strike from the Moyston Fault. In this correlation, the Yarramyljup Fault in western Victoria was considered equivalent to the Tasman Fracture Zone, southwest of Tasmania. Unlike many other approaches, Teasdale et al. (2003) used many different sources of data, including magnetic, gravity and topographic data as well as the geology. They then attempted to derive a geological history of the basement framework and then used this to build a geological history of the overlying rocks.

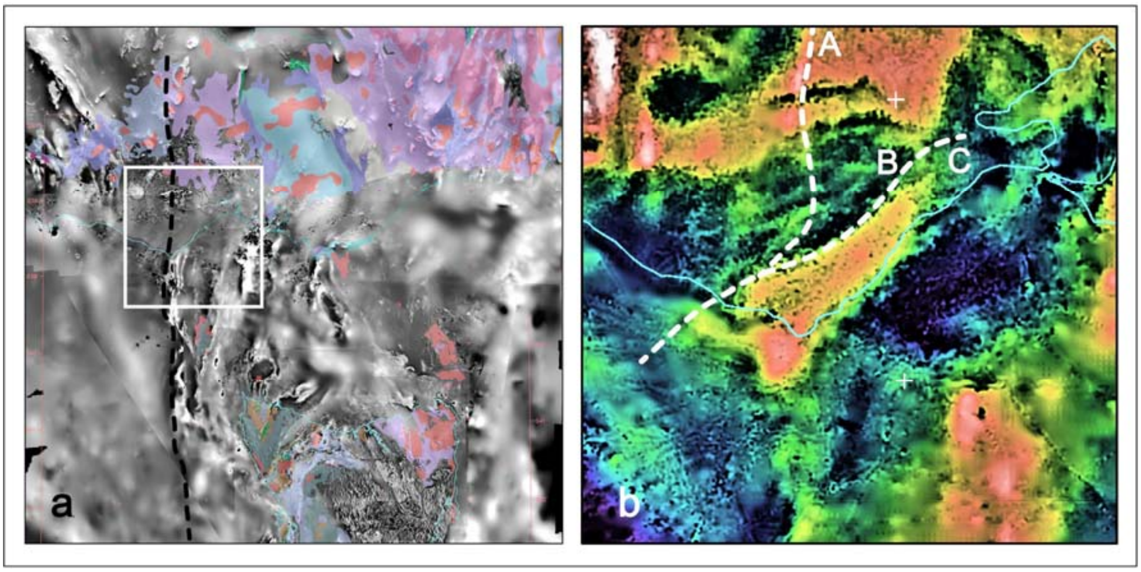


Figure 2-8. a. Total magnetic intensity overlain by the Paleozoic geology (semitransparent). Geological legend as for Figure 2-1. The black dashed line shows Gibson et al.'s (2011) suggested extension southward of the Avoca Fault and the white box shows the location of b. b. Gravity image of the area of the white box. The letter A marks the last mapped position of the Avoca Fault. Further south it is covered by the Cretaceous and Cenozoic sedimentary rocks of the Otway Basin. B marks the position of the Bambra Fault, which can be mapped on the surface (Tickell, 1990) and seen in seismic data (Finlayson et al., 1996). C marks the position of the Ceres Gabbro, an unusual metagabbro with no known correlatives in Victoria. The gravity image shows that the Avoca Fault does not cut the Bambra Fault, as proposed by Gibson et al. (2011) but instead terminates against it.

Morse et al. (2009) and Gibson et al. (2011) used the magnetic and bathymetric data to suggest that the southward extension of the Yarramyljup Fault in western Victoria controlled the position of breakup between Tasmania and Antarctica. They also interpreted the Yarramyljup Fault extended southwards to become the Tasman Fracture

Zone in the Southern Ocean. They correlated the Avoca Fault with the Sorell Fault, along the western margin of the pre-Ordovician Tasmanian crust. Like Stump et al.

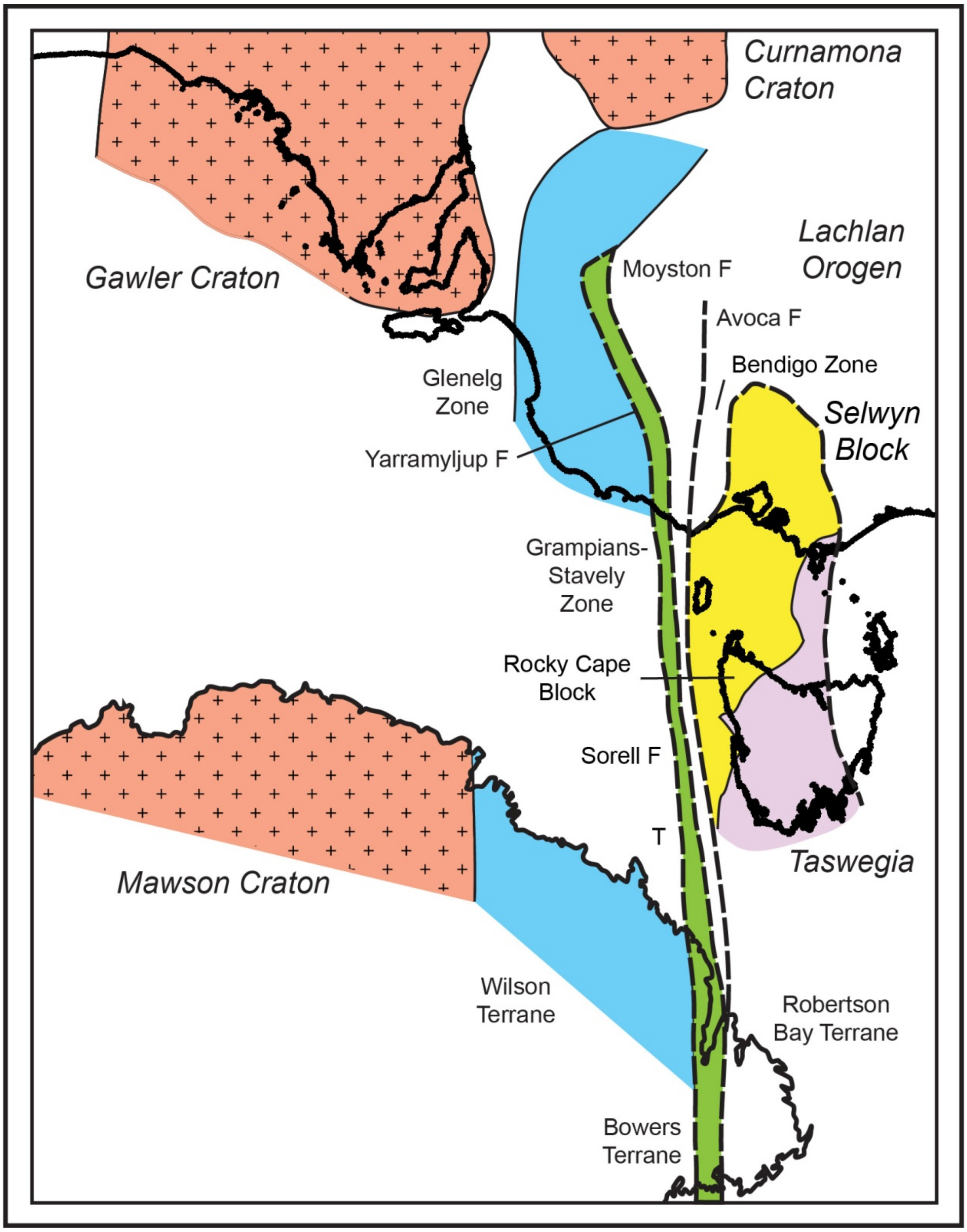


Figure 2-9. Australia-Antarctic connections, after Gibson et al (2011, 2013). Continent positions are for the end of the Jurassic. T marks the position where the Tasmana Fracture Zone would form.

(1986), they interpreted that the Glenelg, Grampians-Stavely and Stawell zones, all extended southwards, west of Tasmania, to link with the Wilson, Bowers and Robertson Bay terranes in Antarctica. The Bendigo Zone was faulted out between the Heathcote-

Bambra and Avoca Faults. Gibson et al. (2011) correlated the Selwyn Block with the Rocky Cape Block in Tasmania, which was displaced to the east of north Victoria Land by at least two Cretaceous or Cenozoic sinistral faults.

Gibson at al. (2011) interpreted that in the Early Cambrian, an island arc lay outboard of Gondwana, separated by an east-dipping subduction zone. At approximately 515 Ma, the island arc collided with the combined Tasmania-Selwyn Block, which was then part of Antarctica. Subsequently slab break-off caused the subduction zone to flip and re-establish outboard of Gondwana with a west dip. They considered that, at about 500 Ma, an east-west subduction zone began along a pre-existing plate boundary in what is now New South Wales. In the Early Ordovician, this drew the Tasmania-Selwyn Block northwards along a sinistral transform that in the Paleozoic was the Avoca Fault, but was reactivated in Tasmania to become the Sorell Fault system during the Australian-Antarctic break-up (Gibson et al., 2011).

Burrett and Berry (2001) and Berry et al. (2008) considered Tasmania and its relationships in the AUSWUS model that fitted Australia and North America in Rodinia (Karlstrom et al., 2001). They considered the possibility that in the Ediacaran, western Tasmania was close to the Musgrave Block in central Australia, but believed that a more likely position was near the central Transantarctic Mountains. This position better matched the metamorphic U/Pb monazite and detrital zircon ages that Burrett and Berry (2001) had found in Tasmania and the ages of detrital zircons in Tasmania given by Black et al. (2004). Burrett and Berry (2001) and Berry et al. (2008) placed Tasmania's Ordovician position in its present place with respect to mainland Australia, but gave no indication of the path it took after rifting from East Antarctica at 580 Ma.

2.4 CONFLICTS, STICKING POINTS AND RESOLUTIONS

Clearly not all of the models outlined in the previous section can be valid. Conflicts between the models are discussed in order. We test the hypotheses, particularly against the potential field data and other recently published data. Where resolution is not possible, we suggest possible future directions of study.

2.4.1 THE NORTHERN EXTENT OF TASMANIAN CRUST

Does Tasmanian crust extend across Bass Strait (Cayley et al., 2002; McLean et al., 2010; Cayley, 2011), or is it truncated by a fault under Bass Strait or the Otway Basin (Stump et al., 1986; Powell and Baillie, 1992; Elliott et al., 1993; Rossiter and Gray, 2008)? If there are faults that truncate the Tasmanian crust under Bass Strait, in which direction do they strike (e.g. Teasdale et al., 2003; Rossiter and Gray, 2008) and when were they active? Could the faults have aligned the gold deposits in eastern Tasmania with those in the Bendigo Zone in Victoria in the Devonian (Bierlein et al., 2005)? If Tasmania does continue north under Victoria, does the Tamar-Tiers Fracture Zone link with the Moyston Fault (Teasdale et al., 2003) or is it truncated by the Arthur Lineament (McLean et al., 2010)? If there is continental crust under the Melbourne Zone, is it an isolated ribbon, unconnected to Tasmania (Spaggiari et al., 2003b; Spaggiari et al., 2003c)?

These questions can be answered at least in part by the newer magnetic and gravity data, which show continuity of units across Bass Strait, notably the Neoproterozoic rift tholeiites on the eastern edge of King Island with metabasalts on Phillip Island (Figures 2-4 and 2-5). To connect with the Moyston Fault, the Tamar-Tiers Fracture Zone would have to cross these magnetic units. The correlation seems highly unlikely, since the magnetic units strike north-northeast rather than the northwest strike demanded by the model of Teasdale et al. (2003) (Figure 2-7). The model proposed by Rossiter and Gray (2008) seems equally unlikely, since it required that the Tamar-Tiers Lineament continue across Bass Strait and then be cut by a northeast-striking sinistral fault. The magnetic data (Figure 2-6a) demonstrate that neither of these faults is present. Furthermore, their proposed Ulrich Transform is at an angle of about 30° to the obvious western edge of the non-magnetic post-Tabberabberan granites above the Selwyn Block and the magnetic ~400 Ma Mt Cole Suite granites further west. It also subdivides the Bendigo Zone where there seems to be little stratigraphic evidence of a significant break (VandenBerg et al., 2000). Instead, the potential field data can be interpreted as showing that western Tasmania, and particularly the regions west of the Tyennan Block (Figure 2-1), continue northwards under central Victoria. This would explain the different granite geochemistry seen in central Victoria. If the granites were sourced from different material from those intruded elsewhere in Victoria, they would inevitably have a different geochemistry (Cayley et al., 2011).

The potential field data offers more support for the ribbon hypothesis of Spaggiari et al. (2003a, b), since the ribbon or ribbons could take any shape and be any size. There is no evidence of major sutures in the outcropping geology of the Melbourne Zone, which suggests that the ribbons were consolidated no later than the Early Silurian and more likely by the end of the Tyennan-Delamerian Orogeny. Thus this hypothesis is, in essence, similar to that by VandenBerg et al. (2000) and Cayley et al. (2002) that thin Tasmanian crust (the Selwyn Block) is also under the Melbourne Zone and nearby areas. This leaves several significant outstanding questions that remained to be addressed including: what were outlines and geology of the ribbon(s)? Where did they come from? When did they amalgamate?

2.4.2 CORRELATING THE MATHINNA SUPERGROUP

Does the Mathinna Supergroup most closely correlate with the Melbourne Zone (Powell and Baillie, 1992; Powell et al., 1993) or with the Tabberabbera Zone (Reed, 2001)?

Mapping by Seymour et al. (2011) in Tasmania and by Willman et al. (2005) and VandenBerg et al. (2006) in Victoria have significantly aided correlations across Bass Strait. The results, summarized in Table 2-1, show that the Mathinna Supergroup is more like the Tabberabbera Zone than the Melbourne Zone. Although fossil control is mostly poor, there is sufficient to discount a correlation of the Mathinna Supergroup with the Melbourne Zone in the Ordovician. The Melbourne Zone was sediment-starved, particularly in the Early Ordovician when the Tippogoree Group was being deposited (VandenBerg et al., 2006; Seymour et al., 2011). The correlation between the Melbourne Zone and the Mathinna Supergroup is more ambiguous in the Silurian and Devonian, since fossil controls are poor. There seems to be no equivalent to the lower part of the Panama Group, the younger Mathinna Supergroup. However Seymour et al. (2011) interpreted a faulted unconformity between the Tippogoree and Panama Groups, suggesting that the older part of the Mathinna Supergroup underwent Benambran deformation. Since such a deformation is not observed in the Melbourne Zone, correlation between the Melbourne Zone and the Mathinna Supergroup seems unlikely.

MATHINNA SUPERGROUP			MELBOURNE ZONE, MT EASTON PROVINCE			TABBERABBERA ZONE		
Unit	**Age**	**Lithologies**	**Unit**	**Age**	**Lithologies**	**Unit**	**Age**	**Lithologies**
Sideling Sandstone	E. Devonian	Fine-grained sandstone, lesser interbedded siltstone	Walhalla Group	E. Devonian	Deep marine mudstone with lesser shale and sandstone	Wentworth Group	E Devonian	Marginal marine to fluvial sandstone; lesser conglomerate & mudstone
Lone Star Siltstone	L. Silurian	Mostly thin-bedded siltstone, interbedded fine-grained sandstone increasing towards top	Jordan River Group	Silurian	Dominantly bioturbated mudstone or siltstone, with lesser sandstone and shale	UNCONFORMITY		
Retreat Formation		Interbedded turbiditic medium to very fine grained sandstone and lesser siltstone-mudstone				Dandongadale Siltstone	E. Silurian?	Siltstone-mudstone dominated, but with minor well sorted sandstone; deep marine turbidites
Yarrow Ck Mudstone		Mostly thin-bedded mudstone, with subordinate cross-laminated siltstone				Moornapa Sst	E. Silurian?	Thick bedded sandstone dominated, with lesser mudstone; marine turbidites
UNCONFORMITY								
			Mt Easton Shale	M.-L. Ordovician	Laminated black shale	Warbisco Shale	L. Ordovician	Black shale
Turquoise Bluff Slate	E.-M. Ordovician	Phyllitic dark grey-black slate; recumbent folded, cleaved				Pinnak Sst	E.-M. Ordovician	Lithic sandstone or granulestone grading upwards to sandstones, mudstones and rare chert; sandstone dominated
Industry Rd Member		Interbedded phyllitic slate and foliated very fine-grained sandstone; recumbent folded, cleaved						
Stony Head Sandstone		Graded thick-bedded fine-grained turbiditic sandstone, minor pelite; large-scale recumbent folded, cleaved	Phosphate Hill Fm	E. Ordovician	Black chert, siliceous shale, phosphorite			

MATHINNA SUPERGROUP			MELBOURNE ZONE, MT EASTON PROVINCE			TABBERABBERA ZONE		
Unit	**Age**	**Lithologies**	**Unit**	**Age**	**Lithologies**	**Unit**	**Age**	**Lithologies**
Sideling Sandstone	E. Devonian	Fine-grained sandstone, lesser interbedded siltstone	Walhalla Group	E. Devonian	Deep marine mudstone with lesser shale and sandstone	Wentworth Group	E Devonian	Marginal marine to fluvial sandstone; lesser conglomerate & mudstone
Lone Star Siltstone	L. Silurian	Mostly thin-bedded siltstone, interbedded fine-grained sandstone increasing towards top	Jordan River Group	Silurian	Dominantly bioturbated mudstone or siltstone, with lesser sandstone and shale	UNCONFORMITY		
Retreat Formation		Interbedded turbiditic medium to very fine grained sandstone and lesser siltstone-mudstone				Dandongadale Siltstone	E. Silurian?	Siltstone-mudstone dominated, but with minor well sorted sandstone; deep marine turbidites
Yarrow Ck Mudstone		Mostly thin-bedded mudstone, with subordinate cross-laminated siltstone				Moornapa Sst	E. Silurian?	Thick bedded sandstone dominated, with lesser mudstone; marine turbidites
UNCONFORMITY								
			Mt Easton Shale	M.-L. Ordovician	Laminated black shale	Warbisco Shale	L. Ordovician	Black shale
Turquoise Bluff Slate	E.-M. Ordovician	Phyllitic dark grey-black slate; recumbent folded, cleaved				Pinnak Sst	E.-M. Ordovician	Lithic sandstone or granulestone grading upwards to sandstones, mudstones and rare chert; sandstone dominated
Industry Rd Member		Interbedded phyllitic slate and foliated very fine-grained sandstone; recumbent folded, cleaved						
Stony Head Sandstone		Graded thick-bedded fine-grained turbiditic sandstone, minor pelite; large-scale recumbent folded, cleaved	Phosphate Hill Fm	E. Ordovician	Black chert, siliceous shale, phosphorite			

Table 2-1. Stratigraphic comparison between the Mathinna Supergroup and the Melbourne and Tabberabbera Zones. The age constraints given are from fossil control. Data from Vandenberg et al. (2000), **VandenBerg et al.** (2004)**, Willman et al.** (2005)**, Vandenberg et al.**(2006) **and Seymour et al.** (2011)**.**

Moreover, Powell and Baillie (1992) noted that the Melbourne Zone verged eastwards, whilst the Mathinna Supergroup verged to the west. Mapping by Willman et al. (2005) has confirmed earlier mapping by Fergusson (1987) that the western Tabberabbera Zone also verges westwards. This suggests that the Mathinna Supergroup correlates at least in part with the Tabberabbera Zone, although the correlation is less robust in the Late Silurian and Devonian.

2.4.3 THE AVOCA FAULT-SORELL FAULT CORRELATION

Does the Avoca Fault mark a major boundary that continues south to the Sorell Fault system (Gibson et al., 2011), or is it terminated by the Bambra Fault (Figures 2-1 and 2-8) (Cayley et al., 2002; Moore, 2002; Cayley, 2011)? Gibson et al. (2011) used the magnetic and bathymetric data to suggest this southward continuation of the Avoca Fault. However the magnetic data are difficult to interpret in this region, since the rocks either side of the Avoca Fault are at best weakly magnetic. As well, the Avoca Fault is covered by Cenozoic basalt and Cretaceous Otway Basin sedimentary rocks south of where it last crops out, about 70 km north of the area where the two faults intersect. However there have been vertical movements on the Avoca and Bambra Faults since the Early Cretaceous, and so the gravity data can be used to test Gibson et al.'s (2011) hypothesis. Isostatic Bouguer gravity data (Figure 2-8) shows that the Avoca Fault is most likely to have been truncated by the Bambra Fault with a significant apparent right-lateral offset, with no evidence of the Avoca Fault extending south of the Bambra Fault.

Furthermore, if the Bambra Fault is the southern extension of the Heathcote Fault Zone, then the timing of movements to the north must inevitably control movements on the Bambra Fault. The Heathcote Fault Zone displaces all of the Silurian and Early Devonian sedimentary rocks of the Melbourne Zone, implying that the Bambra Fault must have also moved then. The deep seismic data over the Heathcote Fault Zone shows that there was at least 20 km of east-verging movement in the Tabberabberan Orogeny, and illite crystallinity data from the Bendigo Zone suggests there may be as much as 50 km of overthrusting (Wilson et al., 2009; Cayley et al., 2011). Since the Bambra Fault strikes obliquely to this movement direction, it suggests between 30 and 70 km of Tabberabberan horizontal movement on the Bambra Fault. There is no evidence for post-Benambran strike-slip movement on the Avoca Fault, and so if the two faults intersect, the Bambra Fault should cut the Avoca Fault. We suggest that neither the observed geology nor the gravity data support an Avoca-Sorell Fault system correlation.

2.4.4 THE DISAPPEARING STAWELL AND BENDIGO ZONES

Cayley (2011) pointed out the paradox that the Stawell and Bendigo Zones strike towards King Island and are 150 km wide in southern Victoria, yet they are not present on King Island, less than 200 km to the south, or anywhere else in Tasmania. The regional Bouguer gravity data of southwestern Victoria shows the Avoca Fault is truncated by the Bambra Fault (Figure 2-8), and so the metaturbidites either side of the fault, in the Stawell and Bendigo Zones, must also have been truncated. Bernecker and Moore (2003) suggested that the Bambra Fault could be traced southwest onto the thin continental crust below the Otway Basin. This interpretation allows for the possibility that the Bendigo and Stawell Zones to be faulted out onto the (now) continental shelf. Bernecker and Moore (2003) agreed with Lister et al. (1991) that the continental crust outboard of King Island had been separated along a detachment fault, with the upper plate now forming the western South Tasman Rise. This could account for the northern part of the western South Tasman Rise including the Kanmantoo-like or Glenelg Complex-like rocks that Berry et al. (2007) described.

2.4.5 CAMBRIAN SUBDUCTION

A major contention in mainland Australian and Tasmanian geology is the polarity of Cambrian subduction. Crawford and Berry (1992) and Münker and Crawford (2000) interpreted the subduction in Tasmania as dominantly east-dipping, since the field evidence showed that the ophiolites obducted as a result of the subduction had come from the east. In contrast, Miller et al. (2005) and Foden et al. (2006) interpreted the subduction zone on the mainland and in Antarctica as west-dipping. The model of Miller et al. (2005) assumed that Tasmania was outboard of the west-dipping subduction zone that affected western Victoria, South Australia and north Victoria Land. As such, a Tasmanian subduction zone was irrelevant to their model. Miller et al. (2005) also interpreted the Moornambool Metamorphic Complex as a reworked zone at the edge of the Delamerian Orogen and the Glenelg Complex metamorphism to result from the dewatering of the west-dipping down-going plate. The east-verging Delamerian thrusts in the Stawell Zone also resulted from this westward shortening against the previously-formed backstop. In contrast, Foden et al. (2006) interpreted the apparently coeval boninite-related magmas in western Victoria, Tasmania and New Zealand as all forming in the same subduction system and that the subduction was clearly west-dipping in Victoria and Antarctica. Subduction resulted from far-field ridge push pressures and that strain was heterogeneously distributed across the Gondwana continental margin.

Crawford and Berry (1992) placed their east-dipping subduction zone close to the Motton Spilite, an interpreted Cambrian oceanic basalt sequence that can be traced at least 70 km offshore in the gravity data and aeromagnetic data. It crosses TASGO seismic line 5 (Figure 2-1) and coincides with a region where Barton (1999) noted east-dipping reflectors down to at least 8 s (about 24 km) in the hangingwall of the Arthur Lineament. Deeper imaging was hindered because the easterly dip could not be imaged near the eastern end of the line. The kinematics of high-temperature mylonites on the soles of ophiolite complexes led Stacey and Berry (2004) to interpret west-directed emplacement, consistent with an east-dipping subduction zone. We also note that seismic tomographic results also are consistent with a relict east-dipping subduction zone to depths of at least 100 km (Rawlinson et al., 2010). It seems likely that at least in northern Tasmania, Cambrian subduction was east dipping. If so, perhaps the most likely resolution is that there were two subduction zones, one east-dipping in Tasmania and the other west-dipping in Victoria and Antarctica, as suggested by Cayley (2011). Further studies are needed to resolve this paradox.

2.4.6 THE DELAMERIAN OROGENY; A COLLISIONAL EVENT?

Was there a Cambrian collision of proto-Tasmania (VanDieland) against Gondwana? If so, where? Was the Late Cambrian felsic volcanism a single line of igneous activity that extended through the Dimboola Igneous Complex-Stavely Volcanic Group through central Victoria to the Mt Read Volcanics in Tasmania and into the Wilson Terrane in Antarctica (Cayley, 2011)? Or was there more than one belt of igneous rocks with similar geochemistry, as implied by Cayley et al. (2002)? Was there no collision at all, as argued by Foden et al. (2006)?

Interpretations for collision tectonics during the Delamerian-Tyennan Orogeny still differ in the number of collisions, their nature, timing and location along the Gondwanan margin. In Tasmania, Stacey and Berry (2004) identified four events, west-directed ophiolite emplacement at about 520 to 515 Ma, south-directed convergence and collision at 510 Ma, east-west extension at about 500 Ma and finally east-west compression at about 495 Ma, implying a complex orogenic event. In southeastern South Australia, Flöttmann et al. (1998) interpreted the Delamerian Orogeny as resulting from west-northwest-directed compression. In far western Victoria, Moore (2006) used the magnetic data over the Dimboola-Stavely Igneous Complex and nearby areas to infer that it had collided from the east, although collision from the northeast was also possible. Further to the east, Cayley and Taylor (2001) focused on the geology east of the Moyston Fault and determined a northeast-southwest maximum compressive stress ($\sigma 1$), a horizontal and orthogonal $\sigma 2$, and an orthogonal and vertical $\sigma 3$ with respect to $\sigma 1$ for late in the Delamerian-Tyennan Orogeny, at about 495 Ma. The magnetic gravity and mapping data show no obvious earlier or later orientation in western Victoria. This seems at odds with the model by Cayley (2011), which places his VanDieland in collision with Gondwana south of western Victoria at about 495 Ma, and then moving northeastwards. If Cayley's (2011) model is valid, it requires a rapid switch in tectonic mode from accretion to extension. As well, there seems to be no obvious location for the present position of the hypothesized fault required to move VanDieland northeast to the Early Ordovician position required by Li et al.'s (1997) paleopole; this part of the model requires a definitive test. These conflicts remain to be reconciled, as does the consequent possibility that the felsic volcanic rocks in the Dimboola-Stavely, central Victoria, Mt Read and Wilson regions might have formed in a single igneous chain. However the magnetic and gravity data show that there is now no direct along-strike correlation of the Mt Read Volcanics with the Stavely-Dimboola Complex (Figure 2-6b).

Foden et al. (2006) interpreted that the Delamerian-Tyennan Orogeny in southeastern Australia was due to "ridge-push stresses to the previously attenuated continental margin at the beginning of subduction. Strain from this event was heterogeneously partitioned across the continental margin, becoming focused in thermally weakened, attenuated crust." In this model, the Delamerian Orogeny ended when slab rollback took place at 490 Ma, causing upper plate extension and anorogenic magmatism. However Cayley (2011) pointed out that this mechanism could not explain the presence of Early to Middle Cambrian oceanic crust in western and central Victoria or the continuing convergence in Antarctica throughout the Ordovician. As well, the magnetic data over the Dimboola-Stavely region strongly suggests that, at least locally, collision took place (Moore, 2006). Thus, at least some aspects of the late stages of the Delamerian-Tyennan Orogeny seem to require collision as a tectonic driver.

2.5 A POSSIBLE TECTONIC HISTORY

We now attempt to reconcile the data and propose a tectonic evolution of the region. Its early history is at best poorly known. The 1400 to 1450 Ma detrital zircon populations indicate that, in the Neoproterozoic and perhaps earlier, the western and central parts of Tasmania were south of north Victoria Land were receiving detritus from Laurentia or its equivalent in Antarctica (Black et al., 2004; Goodge et al., 2008). Although it shares the same detrital zircon population, there are no outcrops known elsewhere along the inland Rodinian margin for the King Island sequence with its 1290 Ma and 760 Ma metamorphic events (Turner et al., 1998; Black et al., 2004; Berry et al., 2005). However Goodge and Fanning (2010) found similar aged inherited zircons from the Ross Orogen off the eastern Wilkes Land coast in Cenozoic sediments and detrital metasedimentary and granite clasts. Parts of the Rocky Cape Block or adjacent fragments may have begun rifting early as 777±7 Ma if the sheared granite present in the Arthur Lineament formed in an extensional regime (Seymour et al., 2007). The extensional basin in which the Burnie Formation was deposited may also have formed at about this time, assuming the 711±16 Ma K/Ar age on the coeval Cooee Dolerite is a minimum depositional age (Black et al., 2004). On mainland Australia, the rifting of Rodinia and the formation of the Adelaide Rift Complex began at about 830 Ma (Preiss et al., 1993; Boger, 2011). Oceanic basin formation was completed after about 580 Ma, as indicated by the age dates on King Island and northwest Tasmania (Calver et al., 2004). As many as six fragments of continental crust could have been rifted during the Rodinian break up and are now incorporated into the Gondwanan margin of southeastern Australia—King Island, the Rocky Cape Block, the northeast Tyennan Block, the southwest Tyennan Block, perhaps small fragments under the Dundas Trough indicated by Leaman and Webster (2002) and also continental crust under the Dimboola-Stavely Igneous complex. Other fragments may now be elsewhere along the Terra Australis Orogen, including the Anakie Inlier in northern Queensland (Cawood, 2005 and references therein) and the Beardmore Micro-continent along the Antarctic margin of Gondwana (Stump et al., 2006). An implication of this is that several small oceanic basins formed, with continental ribbons separating them, perhaps as responses to episodes of roll back on a subduction zone outboard of Rodinia.

At 550 Ma, the west-dipping subduction driving the Ross Orogeny had commenced along the Gondwana margin from southern Victoria Land to South Australia (Stump et al., 2003; Turner et al., 2009). Sedimentation continued outboard of the Adelaide Rift Complex until at least 522 Ma, possibly in strike-slip zones adjacent to continental salients similar to those outlined by Williams et al. (2009) and Cawood (2005), where some parts of an irregular edge can be in compression while others are in extension (Jenkins et al., 2002). However the dominant deformation in southeastern Australia did not begin until about 515 Ma (Foden et al., 2006; Berry et al., 2007). In South Australia and western Victoria magmatic and metamorphic ages imply a break between about 513 and 505 Ma but continuing thereafter until about 490 Ma (Miller et al., 2005; Foden et al., 2006). The shortening direction was towards the west-northwest in South Australia but the later shortening in western Victoria was to the southwest and included sinistral strike slip movement (Cayley and Taylor, 2001; Foden et al., 2006). Deformation in Tasmania was firstly accompanied by west-directed ophiolite obduction (Crawford and Berry, 1992; Foster et al., 2005; Berry et al., 2007). In Tasmania, peak metamorphism occurred at c.a. 510 Ma, during north-south shortening (Stacey and Berry, 2004). At least some Tasmanian deformation resulted from collision of continental blocks, since metamorphism in the Franklin Metamorphic Complex reached 650°C and 1.9 GPa (Chmielowski, 2009). Later events included east-west extension and felsic volcanism in Tasmania and central and western Victoria, and finally east-west shortening and basin inversion (VandenBerg et al., 2000; Stacey and Berry, 2004).

By the Early Ordovician, northwest Tasmania was located outboard of, but relatively close to southeastern Australia (Li et al., 1997). It is likely that the Selwyn Block was due east of its present position; thereafter the only shortening direction in the Melbourne and Bendigo Zones has been east-west (VandenBerg et al., 2000). In the Ordovician, thick turbidite sedimentary packages eroding off the Tyennan-Delamerian highlands were deposited in the Bendigo and Tabberabbera zones in Victoria and formed the Tippogoree Group in eastern Tasmania (VandenBerg et al., 2000; Seymour et al., 2011). In western Tasmania and in the southeastern part of the Melbourne Zone shallow water marine carbonate deposits accumulated on a broad platform, but further north in the Melbourne Zone sedimentation was largely restricted to starved marine sequences that accumulated on a deeper part of the plateau (Seymour and Calver, 1998; Cayley et al., 2002; VandenBerg et al., 2006). Thus western and central Tasmania shed no detritus into the adjacent turbidite flows.

In west central Victoria, the Benambran Orogeny took place at the end of the Ordovician. The deformation was east-verging, giving simple upright folds and west-dipping faults (VandenBerg et al., 2000). In the Tabberabbera Zone, deformation began slightly later, in the Early Silurian and whilst the sediments were still unconsolidated, giving rise to tectonic mélange in the Wonnangatta Shear Zone (Fergusson, 1987; Willman et al., 2005). Here the shortening was northeast-southwest (Willman et al., 2005). The lack of detritus from the Tabberabbera Zone into the now-adjacent Melbourne Zone implies that the two were apart during the Benambran Orogeny and for 20 m.y. afterwards (Cayley et al., 2002). In eastern Tasmania, the deformation included southwest-verging recumbent folds within an overall northeast verging system in the Tippogoree Group (Reed, 2001; Seymour et al., 2011). The tectonic driver for the Benambran Orogeny is likely to have been the west-dipping subduction zone associated with the Narooma Accretionary Complex (Offler et al., 1998; Miller and Gray, 2007).

Subsequently sedimentation increased in the Melbourne Zone, where up to 10 km stratigraphic thickness is present in the west and at least 6 km in the east. Until the Late Early Devonian, sediments were only sourced from either the south or the west (VandenBerg et al., 2000). In the Tabberabbera Zone, local rifts that formed along faults accumulated marine sediments and felsic volcanic rocks, which were then deformed in the latest Silurian Bindian Orogeny (Willman et al., 2005). In eastern Tasmania, an extensional basin may have formed after the Benambran Orogeny (Seymour et al., 2011). This basin accumulated turbidites, and the presence of plant fossils suggests deposition proximal to a emerged landmass (Seymour et al., 2011). In western Tasmania, widespread Ordovician limestone deposition was followed by shelf clastic sequences, mostly sandstone and shale (Seymour and Calver, 1998; Seymour et al., 2007).

In the Tabberabbera Zone, much of the ~420 Ma Bindian deformation was concentrated in dextral strike-slip faults, suggesting that the zone moved southeastwards relative to the undeformed Melbourne Zone (Willman et al., 2005), perhaps driven by the formation of an orocline in the region (Moresi et al., 2014; see section 4.4.4). Miller and Gray (2007) interpreted that at this time, the west-dipping subduction zone associated with the Narooma Accretionary Complex was undergoing trench-perpendicular subduction. If so, it suggests that the east-west shortening could not be accommodated north of the Tabberabbera Zone, causing it to be driven south. This rigid block may either have been the Macquarie Arc or yet another thin, rigid plate fragment from the Rodinian breakup that was now embedded within the Lachlan Orogen. No Bindian deformation has been recognized in Tasmania, although it is coeval with a period of non-deposition in western regions and coincides with a southward-sourced sand pulse in the Melbourne Zone (Seymour and Calver, 1998; VandenBerg et al., 2000).

Tabberabberan deformation affected almost all of the Tasman Orogen (Champion et al., 2009). In Victoria, the Melbourne Zone underwent mostly east-verging east-west shortening (VandenBerg et al., 2006; Cayley et al., 2011). Two phases of deformation are recognized in western Tasmania, an early generally east-northeast-west-southwest compression followed by northeast-southwest compression. Deformation during the Tabberabberan Orogeny was strongly influenced by pre-existing basement structures (Stacey and Berry, 2004). In eastern Tasmania deformation is characterized by an early east to northeast vergence, and a late southwest-directed tectonic vergence onto the old craton (Reed et al., 2002; Worthing and Woolward, 2010; Seymour et al., 2011). This may indicate that the west dipping subduction zone associated with the Narooma Accretionary Complex was still active in driving convergence and was present outboard of Tasmania. In Victoria, the effect of this shortening was for the Bendigo Zone to override the western edge of the Selwyn Block. The sediments of the Melbourne Zone were trapped between the Bendigo Zone in the west and the basement high in the eastern Selwyn Block, where Cambrian volcanic rocks now crop out (Cayley et al., 2011). Because the Selwyn Block was rigid, the overlying sedimentary rocks shortened along a décollement. Further east, the Tabberabbera Zone was obducted onto the easternmost Selwyn Block (VandenBerg et al., 2000).

2.6 CONCLUSIONS

The new magnetic data confirm that the correlation between the geology of Tasmania and the Selwyn Block is most consistent with available geological and geophysical datasets. Models that cannot accommodate the continuity of major magnetic units across Bass Strait seem no longer viable. As well, future models need to take into account the presence of Kanmantoo- or westernmost Lachlan-like metasedimentary rocks on the western south Tasman Rise.

Whilst much of the later history of the region is becoming clearer, the early history is less well understood. Obvious questions include

1. The number of cratonic fragments that came together in the Tyennan Orogeny, and perhaps later, and their previous locations;
2. The basement to the marginal marine Rocky Cape Group. What is it and does it crop out?
3. The path that Tasmania took to arrive at its present position with respect to mainland Australia;
4. The relationship between the Mathinna Supergroup and the Tabberabbera and Melbourne Zones. Whilst the Mathinna-Tabberabbera correlation seems reasonable in the Ordovician and Silurian, the Devonian correlation seems better with the Melbourne Zone;
5. The reason for the granites appearing to young to the west, particularly in Tasmania; and
6. The relationship of the region with Antarctica.

As well, there is a need for mapping and magnetic interpretation that clarifies the detailed correlations across Bass Strait. This will be the subject of future work.

2.7 ACKNOWLEDGEMENTS

This paper stems from work done as part of a PhD at Monash University; their support is gratefully acknowledged. As well, many of the thoughts here result from long and fruitful discussions at GeoScience Victoria, and the project would not have been undertaken without these. The magnetic and gravity data supplied by GeoScience Victoria, Mineral Resources Tasmania and Geoscience Australia were another essential component. The AusIMM/GSA Bicentennial Gold 88 Fund also provided valuable support. Mineral Resources Tasmania has also generously provided other basic data used in this analysis. We also thank the reviewers for the time and effort they spent on this paper. They have greatly improved its quality.

PART B: Declaration for Thesis

Monash University

Declaration for Thesis Chapter 2

Declaration by candidate David H Moore

In the case of Chapter 3, the nature and extent of my contribution to the work was the following:

Nature of contribution	Extent of contribution (%)
Did the research, wrote almost all of the paper	90%

The following co-authors contributed to the work. If co-authors are students at Monash University, the extent of their contribution in percentage terms must be stated:

Name	Nature of contribution	Extent of contribution (%) for student co-authors only
Peter Betts	Supervision, guidance on text	
Mike Hall	Supervision	

The undersigned hereby certify that the above declaration correctly reflects the nature and extent of the candidate's and co-authors' contributions to this work*.

Candidate's Signature	[redacted]	28 August 2015
Main Supervisor's Signature	[redacted]	26 August 2015

*Note: Where the responsible author is not the candidate's main supervisor, the main supervisor should consult with the responsible author to agree on the respective contributions of the authors.

Chapter 3

Behold the turtle, it only makes progress when it sticks its neck out

J.B. Conant (in J.G. Hershberg, 1993, James B. Conant)

Fragmented Tasmania; the transition from Rodinia to Gondwana

ABSTRACT

The origin of the micro-continent VanDieland extends back to the late Paleoproterozoic, where it was positioned between East Antarctica and southwestern Laurentia within the supercontinent Nuna and Rodinia. Palaeo- to Mesoproterozoic events recorded in VanDieland have greater affinities with southwest Laurentia and East Antarctica, suggesting southern VanDieland was part of the Grenville Front and the central Tasmanian part was adjacent to the Miller Range in the central Transantarctic Mountains. Late in the Neoproterozoic Rodinia break-up, VanDieland separated from East Antarctica and southwestern Laurentia and moved north along the Terra Australis margin until its southern part was positioned next to the easternmost Robertson Bay Terrane of north Victoria Land. VanDieland comprises up to seven different crustal mega-boudins or micro-continental ribbon terranes that likely had amalgamated by the end of the Cambrian. These ribbon terranes are bounded by major faults and suture zones. Some boundaries, such as the Arthur Metamorphic Complex, are well known. However, other boundaries, like the eastern edge of the Tyennan Zone, and the boundary between King Island and northwestern Tasmania, are more cryptic as they are covered by younger geology or are under water. The boundaries are often defined by sedimentary and mafic volcanic infill that has been trapped between the crustal fragments. These rocks have previously been interpreted as allochthonous terranes but are more likely to represent inverted sections of attenuated transitional crust and back arc basin fill that formed along the eastern margin of the Gondwana plate during the Cambrian. This interpretation also provides an explanation for the previous tectonic analysis that suggests that Tasmania’s mafic-ultramafic complexes were obducted westward onto older sequences and were subsequently transported southwards as other ribbons collided along the northeastern and western edges of the growing micro-continent, which existed in the overriding plate of a west-dipping subduction zone at the convergent margin between Gondwana and the proto-Pacific plate.

3.1 INTRODUCTION

The period from 600 Ma to 500 Ma includes some of the most fundamental changes in Earth history. Not only did animals develop hard skeletons and the 'snowball earth' time finish, the Rodinian supercontinent had completely dispersed and the Gondwana supercontinent began to amalgamate. Facing the exterior ocean of Gondwana was the Terra Australis Orogen (Cawood, 2005), which records the accretion of dismembered fragments and oceanic material at the edge of Gondwana. This paper examines the geological evolution and crustal architecture of western Tasmania. The region contains excellent evidence of some of the last breakup events of the remnants of Rodinia, at about 580 Ma (Calver et al., 2004; Meffre et al., 2004). Thus western Tasmania provides an opportunity to examine a fundamental transition in Earth history, from almost 300 million years of continent destruction to 300 million years of continent building. Furthermore, the Cambrian orogenesis in Tasmania apparently took place isolated from the rest of Gondwana, and terranes of western Tasmanian were not truly accreted onto the Gondwanan margin until the Middle Devonian (Cayley et al., 2002; Cayley et al., 2011; Moresi et al., 2014). Hence the region provides opportunity to understand tectonic processes associated with Rodinia break-up and subsequent accretion on the Gondwanan margin during the Cambrian to Devonian stages of the Terra Australis Orogen.

Cayley (2011) coined the term 'VanDieland' for a Proterozoic micro-continent that included western Tasmania, the Selwyn Block (central Lachlan Fold Belt), the South Tasman Rise and the East Tasman Plateau. VanDieland extends over 1500 km from north to south (Figure 3-1) and has been considered to represent a single, coherent entity (Berry et al., 2007; Cayley, 2011; Berry and Bull, 2012). Western Tasmania (Figures 3-1 and 3-2) contains the most complete and accessible geological record of VanDieland. This region contains a wide range of pre-Ordovician rocks, including deep water turbidites, non-marine and shallow marine clastic deposits, carbonate rocks, and cratonic fragments. The oldest rocks were deformed and metamorphosed in the Mesoproterozoic (Chmielowski (2009). Extensive areas of VanDieland contain Neoproterozoic to Cambrian rift assemblages of deep-water metasedimentary rocks,

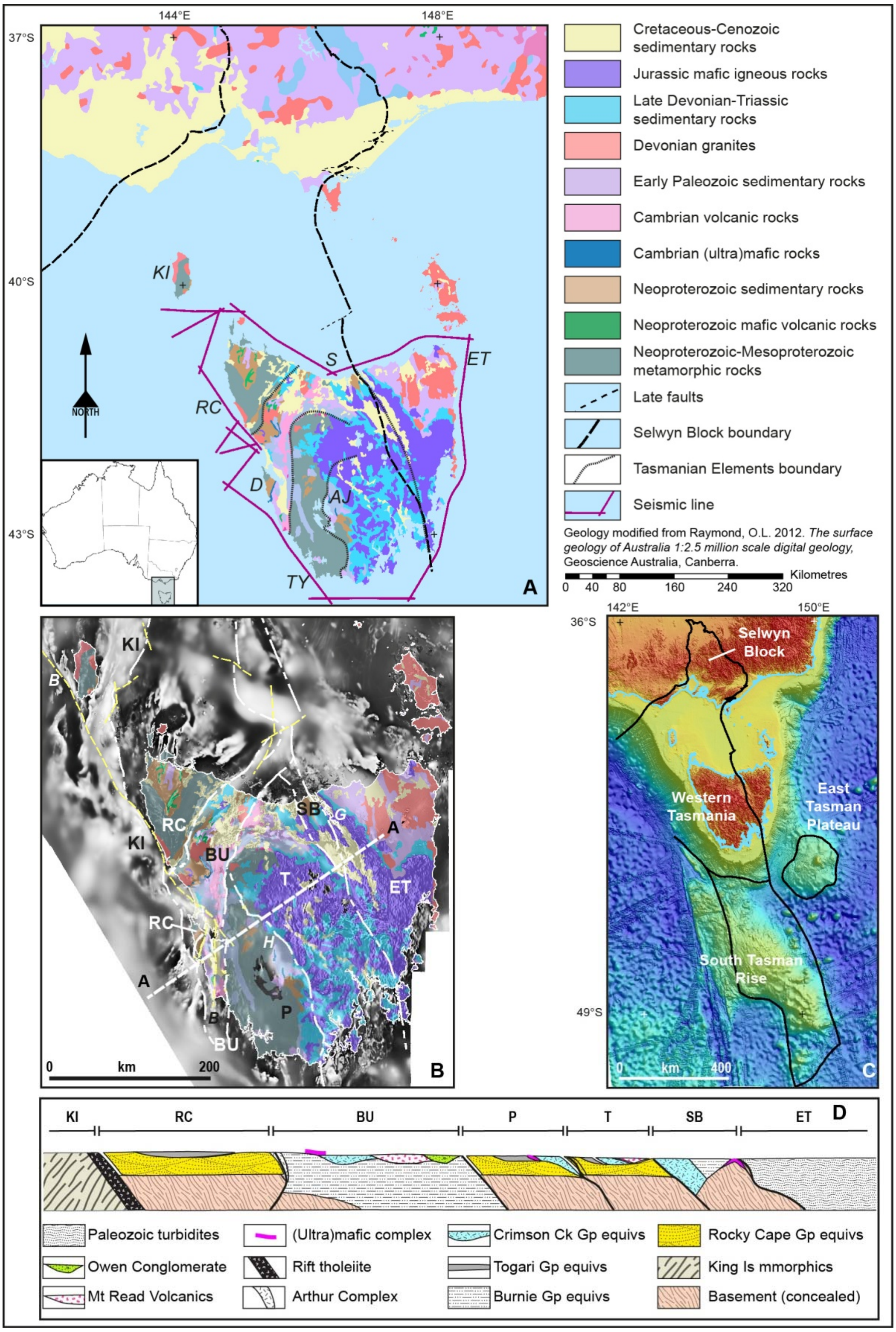


Figure 3-1. A. Southern Victorian and Tasmanian geology, showing the near-surface boundaries of the Selwyn Block and Tasmanian Elements (from Seymour and Calver, 1998). Elements are labelled as *AJ* Adamsfield–Jubilee, *D* Dundas, *ET* Eastern Tasmania, *KI* King Island, *S* Sheffield, *RC* Rocky Cape and *TY* Tyennan. Inset shows the location within Australia. B. Tasmanian geology as in A with the total magnetic intensity as an intensity layer. Suggested zones are BU Burnie Zone, ET Eastern Tasmania, KI King Island, SB Sorell–Badger Head, P Pedder (including the Adamsfield–Jubilee Element; see Sections 2.4 & 3.1.3), RC Rocky Cape, T Tyennan Zone. *B* marks the Braddon River Fault, *G* the Tamar Graben and *H* the Mt Hobhouse Fault. C. VanDieland subdivisions, modified from Cayley (2011) to conform with the usage here. D. Schematic section across Tasmania, generalised from the cross section A-A' in B.

granitic plutons, within-plate basalts or ocean floor basalts, Cambrian boninite-associated mafic-ultramafic complexes, eclogite, blueschist, greenschist and amphibolite facies metamorphic rocks (Berry and Crawford, 1988; Crawford and Berry, 1992; Black et al., 1997; Corbett et al., 2014). Low temperature-high pressure metamorphic rocks formed during the Middle to Late Cambrian Tyennan Orogeny (Meffre et al., 2000; Chmielowski and Berry, 2012). Rocks with oceanic affinities have been interpreted as allochthonous sheets emplaced at one or more collision zones associated with the Tyennan Orogeny (e.g. Stacey and Berry, 2004; Berry and Bull, 2012) (Crawford and Berry, 1992; Turner et al., 1998; Meffre et al., 2000; Berry et al., 2005; Chmielowski and Berry, 2012; Seymour et al., 2013). The presence of these obducted mafic-ultramafic rocks suggests that discrete continental crustal blocks may have been separated by oceanic or transitional crust (Moore et al., 2013). The Tyennan Orogeny is one example of the earliest phases of Gondwanan orogenesis (Crawford and Berry, 1992; Everard et al., 2007; Berry and Bull, 2012; Seymour et al., 2013) and appears to be part of a much larger orogenic system that includes the Ross Orogeny in Antarctica and the Delamerian Orogeny on mainland Australia (Boger and Miller, 2004; Cawood, 2005; Champion et al., 2009; Boger, 2011). By the Ordovician, the extensive limestone sheet that stitches the many parts of western Tasmania indicates that the collage of cratonic and oceanic fragments had become a single continental ribbon (Seymour and Calver, 1998; Moore et al., 2013), with final consolidation into Gondwana in the Middle Devonian (Cayley et al., 2011).

In constructing a new tectonic model for VanDieland, we used detailed mapping by Mineral Resources Tasmania (including Jennings, 1963; Gee and Legge, 1979; Brown, 1986; Brown et al., 1989; Corbett, 2003; McClenaghan and Vicary, 2005; Reed and Vicary, 2005; Calver et al., 2006; Everard et al., 2007; Calver, 2008; Vicary et al., 2008; and summarised in Mineral Resources of Tasmania, 2011; Brown et al., 2012; and Corbett et al., 2014) and others (e.g. Powell and Baillie, 1992; Exon et al., 1997b; Berry and Gray, 2001; Hall, 2001; Holm and Berry, 2002; Noll and Hall, 2005), and combined these data with airborne magnetic data and the ground gravity data collated by Geoscience Australia.

Where appropriate, we supplemented the data with interpretations of the TASGO seismic survey that circumnavigated Tasmania (Figure 3.1A) (Hill and Webber, 1995;

Drummond et al., 2000). This approach allowed a reliable connection of the many geological observations along the coast with the sparse inland data. The geophysical interpretation was grounded in geological observations, while the geology could more confidently be extrapolated beyond the immediate outcrop areas.

We propose a tectonic model that defines each of the end-Proterozoic 'micro-plate' boundaries and provides a geological history for their amalgamation. We resolve major Cambrian structural and stratigraphic breaks and define their spatial extent, geometry and role in the evolution of an amalgamated VanDieland. We submit that, prior to the Tyennan Orogeny, western Tasmania was not a single cratonic fragment, but rather several microplates that had previously been extended to form mega-boudins (ribbons) linked by transitional oceanic crust, which lay in an oceanic setting similar to the present western Pacific (Crawford et al., 2003b; Metcalfe, 2011). In modern settings, these fragments can be full- or part-thickness cratonic fragments, as well as parts of fore-arc, back-arc, oceanic or turbidite packages (Cawood et al., 2009). This contrasts with the Ordovician to Devonian evolution of the Tasmanides, which was dominated by accretionary processes and resulted in turbidite-dominated successions that are thrust-repeated with trivial amounts of oceanic crust (Gray and Foster, 1998), a setting that typifies parts of the modern northwest Pacific (Cawood et al., 2009). Our proposed model offers tectonic solutions for long-standing paradoxes such as the apparent change in Cambrian subduction polarity along the Gondwanan margin (c.f. Crawford and Berry, 1992; Miller et al., 2005; Foden et al., 2006), as well as an opportunity to assess and test different tectonic models, including those involving rapid changes of subduction polarity (Crawford and Berry, 1992; Crawford et al., 2003b), oblique subduction on an outboard fragment (Cawood, 2005) and complex tectonic scenarios where contemporaneous east- and west-dipping subduction zones were present in the Early Cambrian and VanDieland was sucked obliquely into Gondwana by the west-dipping subduction zone in the Late Cambrian before rotating and moving away from Gondwana during the Early Ordovician (Cayley, 2011). Finally, we place VanDieland in the context of Precambrian continental reconstructions.

3.2 REGIONAL GEOLOGY

Tasmania is typically separated into two regions. Western Tasmania comprises Mesoproterozoic and Neoproterozoic cratonic crust, whereas eastern Tasmania has oceanic affinities and was cratonised in the Middle Devonian Tabberabberan Orogeny (Figures 3-1 and 3-3) (e.g. Reed et al., 2002; Black et al., 2004; Berry and Bull, 2012; Seymour et al., 2013). The boundary between eastern and western Tasmania is believed to be present beneath the Tamar Graben (Figure 3-1B) (Seymour et al., 2013), just west

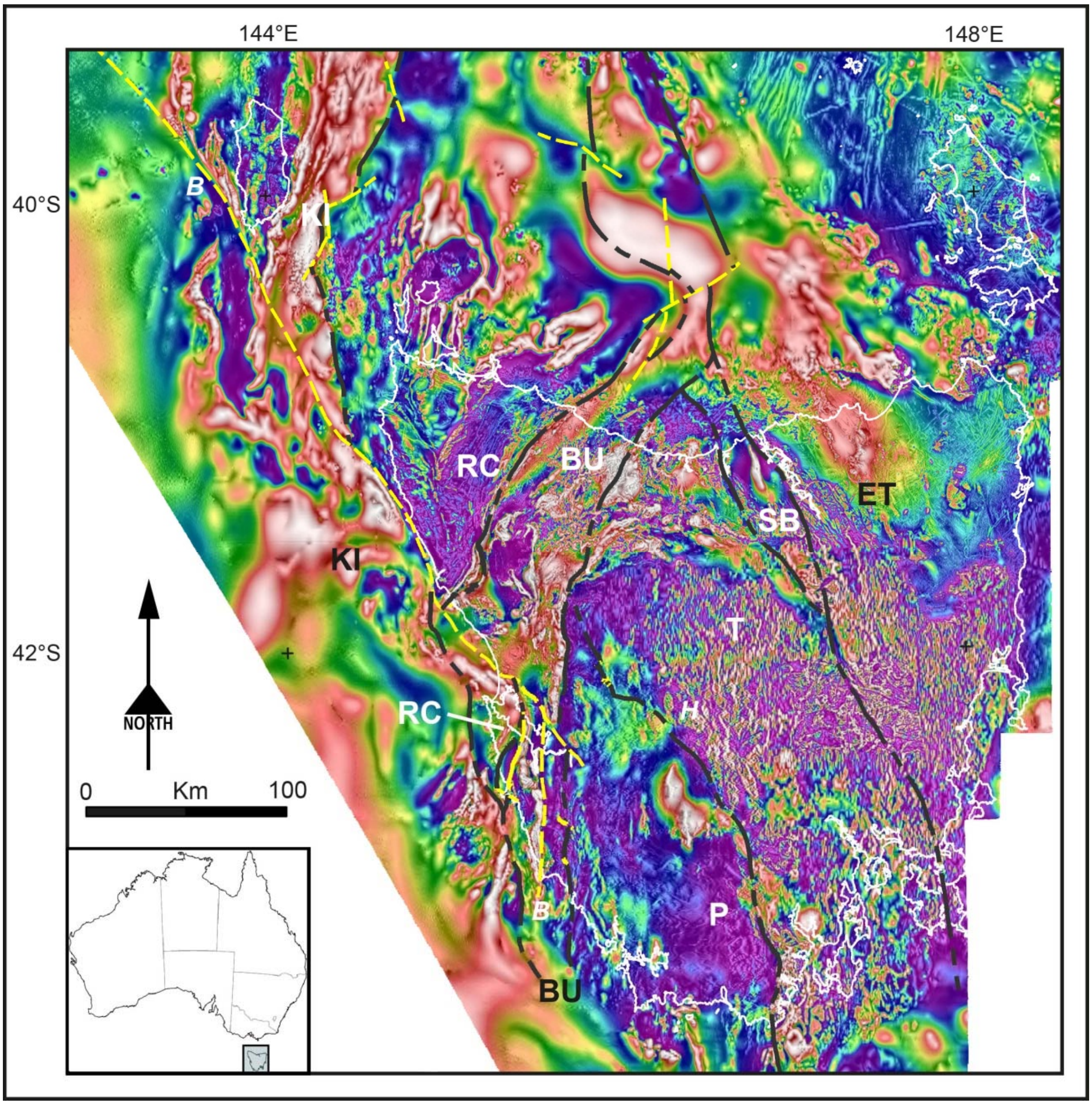


Figure 3-2. Tilt-derivative filtered magnetic image of Tasmania and adjacent areas, together with the zone boundaries (black), later faults (yellow) and letter symbols from Figure 3-1B.

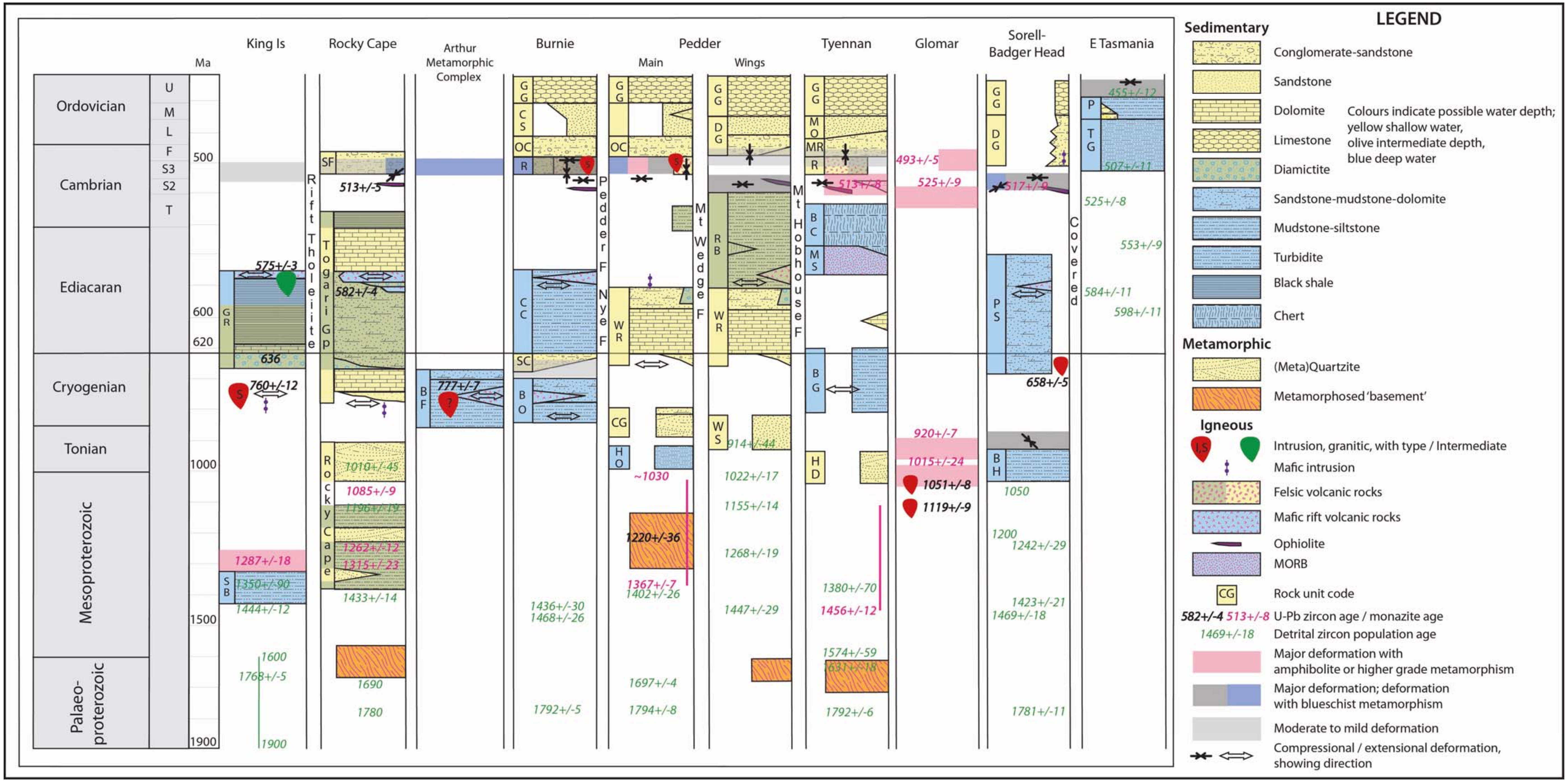


Figure 3-3. Tasmanian Pre-Silurian time-space plot for zones outlined in Figure 3-1B and C. Because of internal differences, the Wings Subzone, largely the Adamsfield–Jubilee Element of Seymour and Calver (1995), has been subdivided from the main Pedder Zone. The Arthur Metamorphic Complex has also been shown as a separate column. Rock unit codes; BC Barrington Chert, BF Bowry Formation, BH Badger Head Group, BO Burnie and Oonah Formations, CC Crimson Creek Formation and equivalents, CG Clark Group, CS Caroline Creek Sandstone, DG Denison Group, GG Gordon Group, GR Grassy Group, HD Howell, Fisher and Dove Groups, HO Harrisons Opening Formation, MO Moina Sandstone, MR Mt Roland Conglomerate, MS Motton Spilite, OC Owen Conglomerate, P Panama Group, PS Port Sorell Group, R Mount Read Volcanics, RB Ragged Basin Complex, SB Surprise Bay Formation, SC Success Creek Formation, SF Scopus Formation, TG Tippogoree Group, WR Weld River Group, WS Wings Sandstone. References in the text.

of which is seen a Lower Devonian sequence that has affinities with both eastern and western Tasmanian rocks (Rickards et al., 2002). Further south, the boundary is covered by Permian and younger rocks. Offshore seismic reflection data from eastern Tasmania indicated the possible presence of Proterozoic crust at depth along the southern half of the Tasmanian east coast (Drummond et al., 2000). In contrast, Seymour et al. (2013) suggested that eastern Tasmania is not floored by Proterozoic crust and placed the boundary further west. Eastern Tasmania has always been considered as a single block or 'element' (Burrett and Martin, 1989; Seymour and Calver, 1998; Corbett et al., 2014). Western Tasmania records a more complex history including Neoproterozoic rifting, which involved break-up and eastward drift of the micro-continent from East Antarctica (Berry et al., 2008), followed by Middle to Late Cambrian crustal shortening and subsequent extensional tectonism, the Tyennan Orogeny (Berry and Bull, 2012). Late Ordovician to Devonian reworking associated with the accretion of VanDieland into the eastern Gondwanan margin (Cayley et al., 2011) affected both eastern and western Tasmania as eastern Tasmania was either partly (Patison et al., 2001) or completely (Powell and Baillie, 1992; Reed et al., 2002) transported westward during the Tabberabberan Orogeny. During the Jurassic, a Large Igneous Province resulted in voluminous emplacement of dolerite (Hergt and Brauns, 2001). Cretaceous and Paleogene extension and crustal thinning led to the break-up of Australian and Antarctica (Norvick and Smith, 2001; Veevers, 2012), which was followed by Cenozoic basaltic volcanism and sedimentation (Stacey and Berry, 2004; Mineral et al., 2011; Corbett et al., 2014).

Seymour and Calver (1995) divided western Tasmania into six 'elements'; King Island, Rocky Cape, Dundas, Sheffield, Tyennan and Adamsfield–Jubilee (Figure 3-1A). Some boundaries between the elements are well established. For example, the Arthur Metamorphic Complex is a high-strain zone containing blueschists and rift-related rocks that separate the Rocky Cape and Dundas Elements (Turner and Bottrill, 2001; Holm and Berry, 2002). Other boundaries are covered. The boundary separating eastern Tasmania from elements to the west is beneath younger sedimentary successions, Jurassic dolerite and Cenozoic basalt. The boundary between King Island and the Rocky Cape Element is covered by water and Cretaceous and Cenozoic sedimentary rocks. The boundaries have also been modified by later events, notably the Devonian Tabberabberan Orogeny (Holm and Berry, 2002; Reed et al., 2002). The multiple deformations resulted in significant metamorphic grade variations and styles across the boundaries, which range from greenschist to amphibolite and blueschist facies (Meffre et al., 2000; Chmielowski and Berry, 2012). Age dating criteria are unable to distinguish the various phases of deformation as most record ages that cluster at ca. 510 Ma (e.g. Figure 10 in Berry et al., 2007).

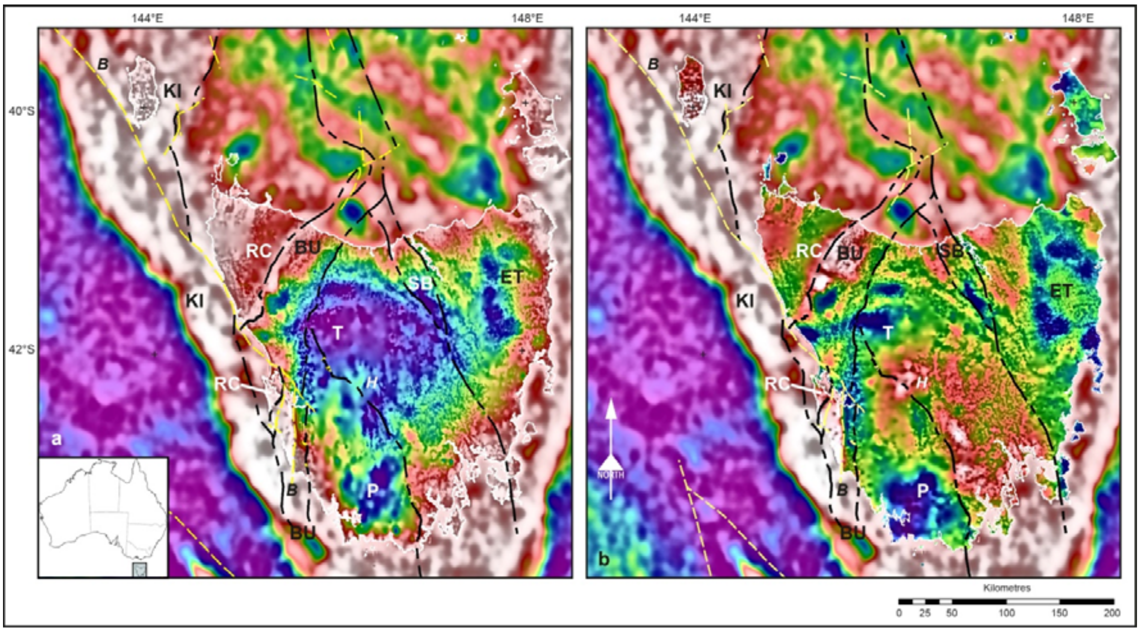


Figure 3-4. Gravity image of Tasmania and adjacent areas, a Bouguer onshore, free air offshore, b, isostatic onshore, free air offshore, together with the zone boundaries (black), later faults (yellow) and letter symbols from Figure 3-1B.

The following section first outlines each of the zones that comprise western Tasmania and the South Tasman Rise. Figure 3-3 shows the pre-Silurian stratigraphic columns for each zone. Some zones replicate the "Elements" of Seymour and Calver (1995, 1998), but we have chosen to vary or subdivide others. In order to avoid confusion, we have called our divisions "Zones" and only used the term "Element" where it refers to areas defined by them.

3.2.1 KING ISLAND ZONE

King Island is largely comprised of Mesoproterozoic metasedimentary rocks with detrital zircon populations older than 1350±90 Ma (Black et al., 2004). In the west, the rocks underwent regional metamorphism to greenschist or amphibolite facies at 1287±18 Ma (U/Th/Pb on monazite, Berry et al., 2005). On the west and north coasts, later metamorphism, deformation and granite intrusion at 760±12 Ma is coeval with ca 750-780 Ma granites (Black et al., 1997; Turner et al., 1998; Calver et al., 2013b). On the east

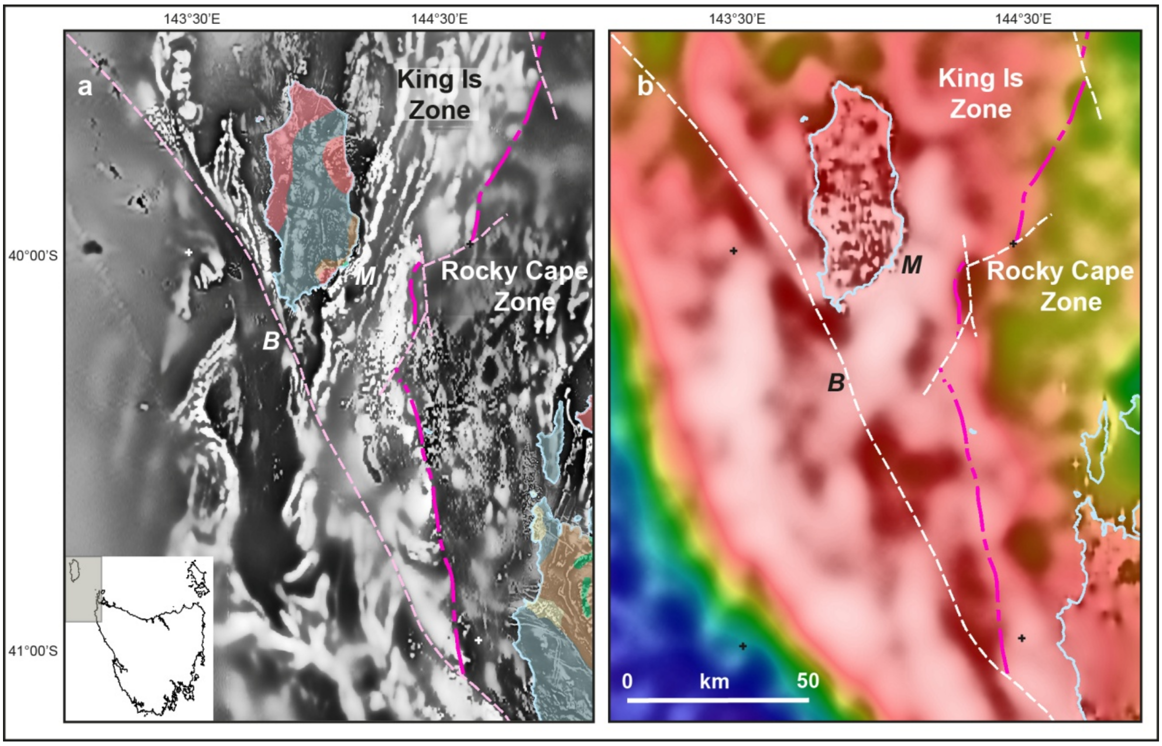


Figure 3-5. King Is Zone. a. Geology overlaid on the total magnetic intensity plus 40% of its first vertical derivative. b. Gravity response, Bouguer onshore, Free Air offshore. Location given by the shaded area of the map of Tasmania inset in a. Thicker dashed magenta lines mark zone boundaries; paler thinner dashed lines mark faults cutting zone boundaries. *B* marks the Braddon River Fault and *M* the location of outcropping rift tholeiite. The high magnetic and gravity responses associated with the tholeiite continue north to southern Victoria, where a MORB outcrop is present (Henry and Birch, 1992). The eastern edge of this high is taken as the boundary of the King Island Zone. Magnetic modelling suggests that the sequence has an east dip of 45° to 70° (Meffre et al., 2004). Geology from Raymond et al. (2012); grey and brown mark Proterozoic metasediments, green, Proterozoic mafic volcanic rocks, red granite and yellow Cenozoic sediments.

coast, metasedimentary successions are unconformably overlain by Marinoan glaciomarine deposits with detrital zircon dates of ca 636 Ma (chemically abraded TIMS $^{238}U/^{206}Pb$ zircon, Calver et al., 2013a). This sequence is successively overlain by a cap carbonate, laminated black shale and the mafic volcanic (rift tholeiites) and intrusive rocks that form the eastern edge of the zone. The mafic rocks are strongly magnetic and relatively dense and are evident in regional geophysical data as a 35 km zone wide zone (Figures 3-2, 3-4 and 3-5) that extends between outcrops on the east coast of King Island to Phillip Island on the southern Victorian coast (Henry and Birch, 1992; Moore et al., 2013). The magnetic package is truncated by the Braddon River Fault, approximately 40 kilometres to the south of King Island. The package dips moderately to steeply to the east between 45° and 70° (Meffre et al., 2004). Meffre et al. (2004) suggested the sequence was tilted and fault repeated, and proposed a pre-repetition width of several tens to hundreds of kilometres of rift tholeiites. Based on stratigraphic and geochemical considerations, a volcanic passive margin sequence setting was interpreted. Sm/Nd analyses suggested the tholeiites were emplaced at ca 580 Ma (Meffre et al., 2004). Mafic dykes in the sequence yielded a U-Pb SHRIMP date of 575±3 Ma (Calver et al., 2004).

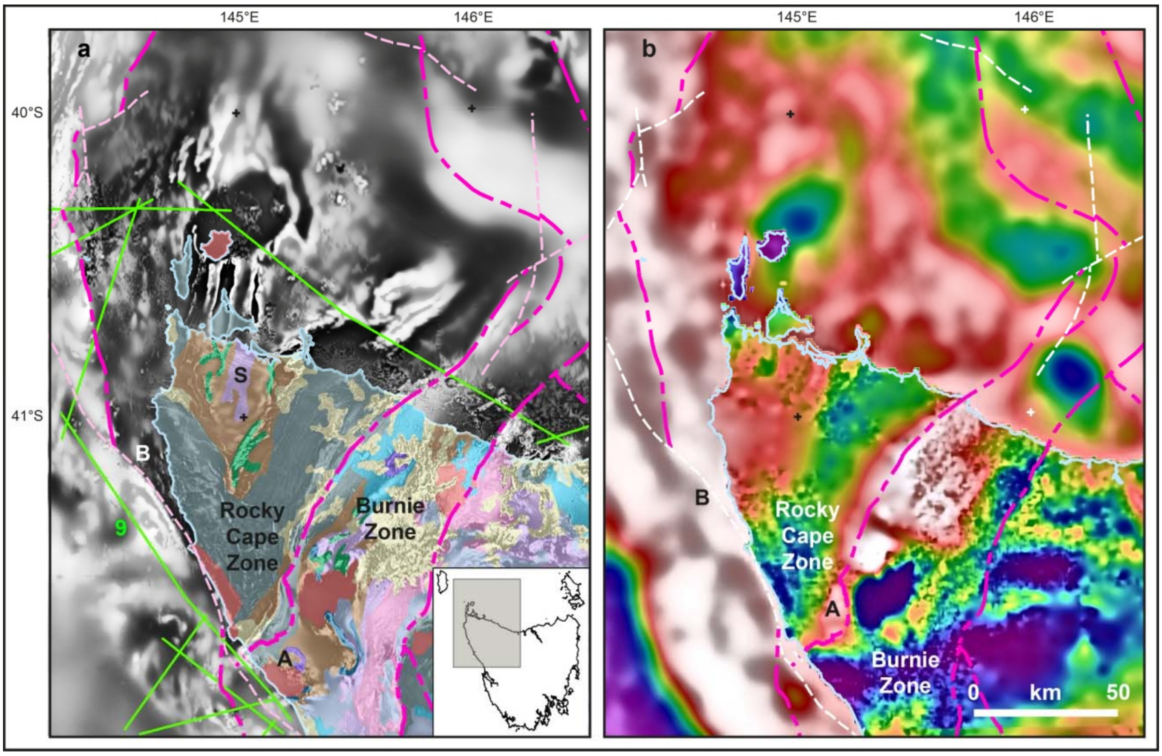


Figure 3-6. Northern Rocky Cape Zone. a. Geology on total magnetic intensity. Location given by the shaded area of the map of Tasmania. Grey marks Mesoproterozoic metasediments, brown Neoproterozoic metasediments, Irish green Cambrian basalt, pink and lavender Cambrian felsic rocks, red granite, bright blue Permian sedimentary rocks, and yellow Cenozoic sediments. Geology from Raymond et al. (2012). A marks the Arthur Metamorphic Complex, the eastern boundary of the Rocky Cape Zone, B, the Braddon River Fault and S, the Smithton Basin. Apple green lines, seismic lines; 9, line 148/9 (Figure 3-7). b. Isostatic gravity onshore, free air offshore. The magnetic and gravity responses suggest an eastwards dip of the Arthur Metamorphic Complex.

The Braddon River Fault (Figures 3-1, 3-2, 3-4 to 3-6) is prominent in regional magnetic images of offshore western Tasmania. It is exposed onshore at the Braddon River head waters, where late movement has displaced Early Paleozoic rocks (Baillie and Corbett, 1985). Further south, it is defined by a 2 km wide mylonite at the eastern edge of a ?Neoproterozoic mafic volcanic rock package (Brown (2011). The Braddon River Fault is imaged in seismic reflection data as steeply west-dipping and has a significant change in seismic character across it (Figure 3-7). South of King Island, it appears to have a displacement of approximately 70 km. However, further south much of this movement appears to be taken up in internal faulting within the Burnie Zone (Figure 3-8).

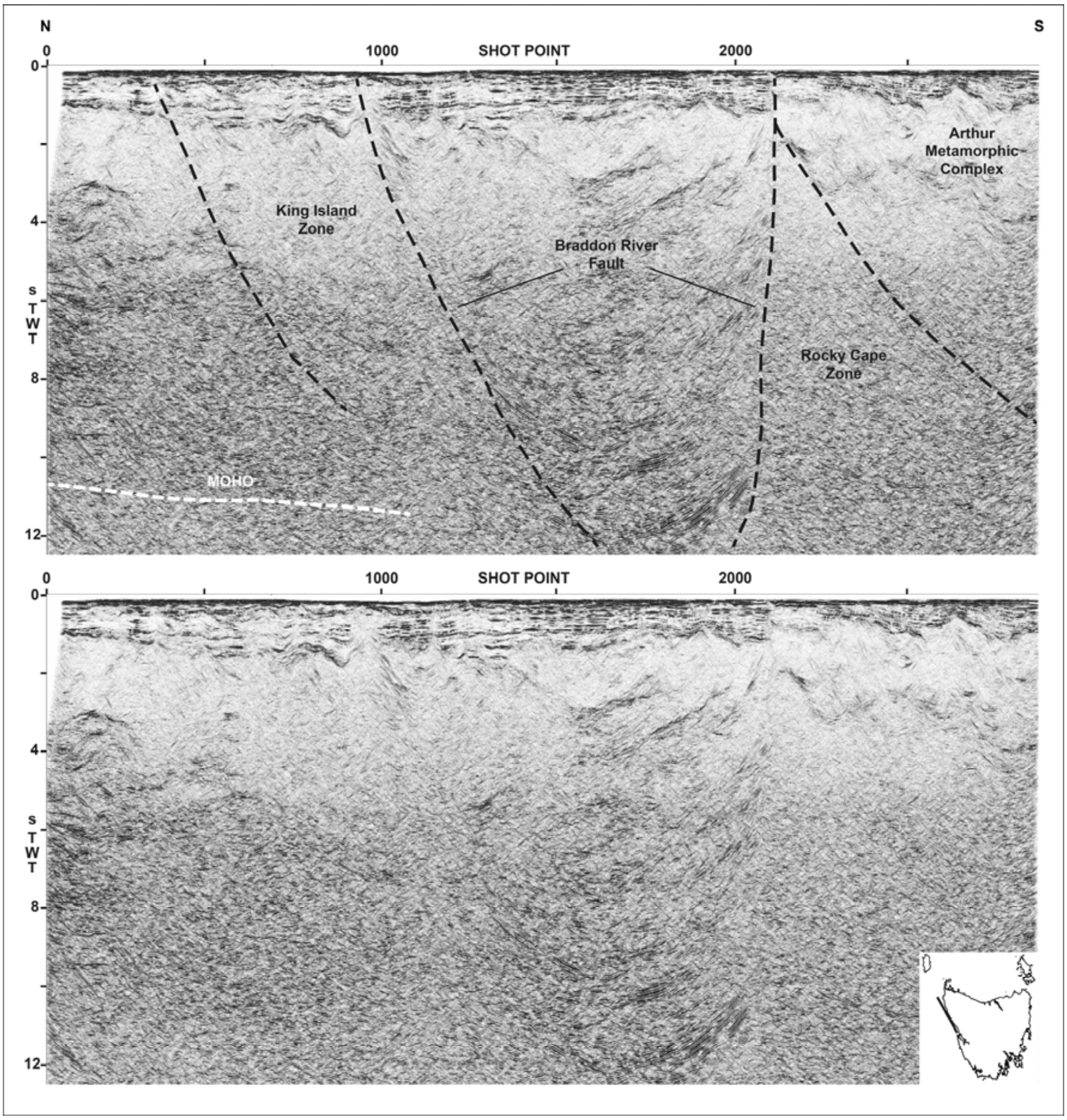


Figure 3-7. Seismic line, AGSO 148/9, upper panel shows the interpretation, lower panel uninterpreted line, inset gives the line location. The Braddon River Fault is sub-parallel to the seismic line causing significant off-line effects. Shot points are approximately 50 m apart.

3.2.2 ROCKY CAPE ZONE

The Rocky Cape Zone occurs in the northwest corner of Tasmania and the northern Sorell Peninsula (Figures 3-1 to 3-4, 3-6, 3-8 and 3-9). The oldest rocks are the Mesoproterozoic Rocky Cape Group, which comprises alternating sequences of marginal marine quartzite and shelf siltstone (Figure 3-9, Table 3-1) (Everard et al., 2007; Halpin et al., 2014). The presence of the fossil *Horodyskia williamsii* suggests sedimentation at the basal part of the package occurred between 1400 and 1100 Ma (Calver et al., 2010). Authigenic monazite from quartzites yield dates that cluster in three populations, between ca 1360-1290 Ma, ca 1280-1240 Ma and ca 1090 Ma (Halpin et al., 2014). Cambrian and Devonian granites that intrude the Rocky Cape Group contain slightly different zircon inheritance patterns, notably an excess population of 1650-1600 Ma ages that are poorly represented in the Rocky Cape Group (Black et al., 2010). These are likely to be sourced from the cratonic basement on which the Rocky Cape Group was deposited (Black et al., 2010).

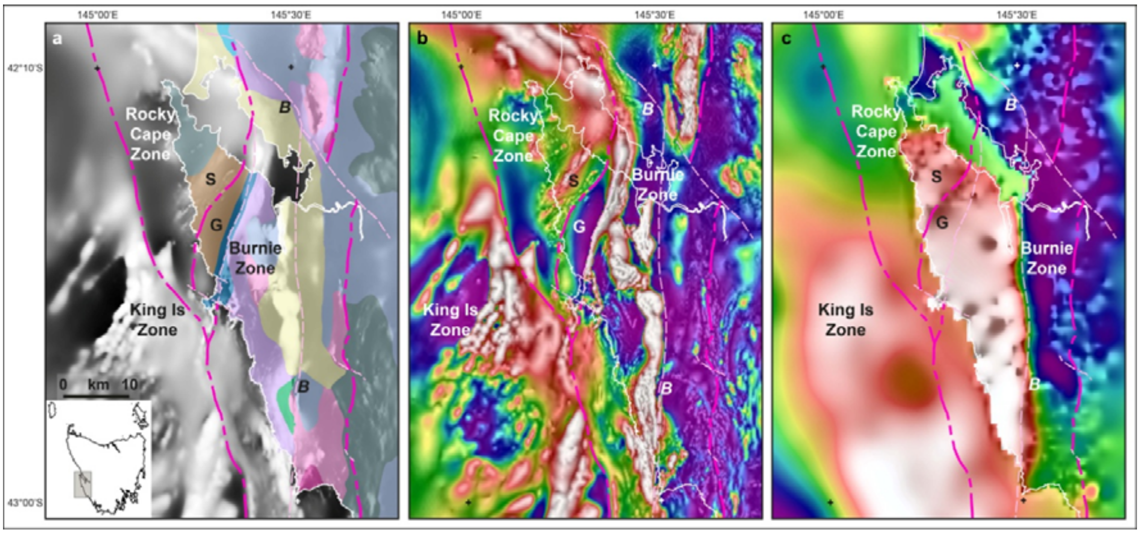


Figure 3-8. Sorell Peninsula area. a. Geology on background grey scale total magnetic intensity. Location given by the shaded area of the map of Tasmania. Colour legend as for Figure 3-1A, with dark blue Cambrian ultramafic rocks. B marks the Braddon River Fault, G, the Burnie (Oonah) Formation and S, the interpreted southern extension of the Smithton Basin. b. Tilt filtered total magnetic intensity. c. Gravity, isostatic onshore, free air offshore. The Rocky Cape-Burnie and Burnie-Pedder zone boundaries have been largely determined by the outcropping and inferred Precambrian boundaries. The former is interpreted as an east-dipping thrust fault, which we interpret to lie in front (west) of a much larger fault marked by a mafic-ultramafic-boninite package at the western edge of a Cambrian felsic volcanic sequence. We interpret this fault system as an east-dipping, leading imbricate fan (Boyer and Elliott, 1982)**. In the magnetic and seismic images, these major boundary systems can be traced south from the Braddon River Fault until they are truncated by the southern extension of the eastern boundary of the King Island Zone (Figures 3-2 and 3-10). Although the mafic-ultramafic-boninite package has been subsequently disrupted by the Devonian Tabberabberan deformation, we consider that this western boundary of the Burnie Zone on the Sorell Peninsula is more or less in place. In the north, the Braddon River Fault has a sinistral displacement of up to 70 km. However, the eastern boundary of the Burnie Zone is only displaced by approximately 5 km, and we suggest that most of the missing movement has been taken up along the sheared eastern boundary of a Neoproterozoic mafic volcanic package within the Burnie Zone. These volcanic rocks can be seen as both magnetic and gravity highs.**

Within the Smithton Basin, the Togari Group unconformably overlies the Rocky Cape Group (Everard et al., 2007). It includes conglomerate, dolomite, chert, diamictite, volcaniclastic rocks, siliceous metasedimentary rocks and tholeiitic basalt, which were deposited in an extensional setting (Everard et al., 2007). $\delta^{13}C$ values and $^{87}Sr/^{86}Sr$ ratios (Calver, 1998) from the Togari Group and correlations with similar succession of the Adelaide Fold Belt suggests the lower Togari Group may be mid-Cryogenian (Everard et

Figure 3-9. Rocky Cape Group (Detention Metaquartzite) with herringbone crossbeds just above the scale. Location 40°51'41"S, 145°30'44"E. Scale is 16 cm long.

al., 2007). A rhyodacite in a rift tholeiite package in the upper Togari Group yielded an U-Pb zircon SHRIMP age of 582±4 Ma (Calver et al., 2004), while the uppermost siltstone unit contains Early to Middle Cambrian fossils (Everard et al., 2007). A Late Cambrian marginal marine to terrestrial clastic sequence overlies the Togari Group. This package contains serpentinite-rich detritus from mafic-ultramafic rocks, similar to those that were obducted onto the adjacent Burnie Zone during the Cambrian Tyennan Orogeny (Everard et al., 2007).

3.2.3 BURNIE ZONE

The oldest rocks of the Burnie Zone are the clastic metaturbidites of the Burnie (Oonah) Formation (Figures 3-10 to 3-13). Similar detrital zircon populations to the shallow water Rocky Cape Group can be interpreted as the two being coeval (Black et al., 2004). More likely, the Burnie Formation represents a younger package derived from the Rocky

Youngest sample	Jacob Quartzite	Detention Subgroup	Data sources
Detrital zircon populations (Ma)	1009±43, 1250, 1440, 1850, 2650	1433±14, 1690-1740, 1780	Black *et al.* 1997, 2004
Authigenic monazite age (Ma)		1085±9	Halpin *et al.* 2014
Oldest sample	**Pedder River Siltstone**		
Detrital zircon populations	1442±28, 1630, 1720, 1790, 1850		Halpin *et al.* 2014
Authigenic monazite (Ma)	1315±23, 1350		Halpin *et al.* 2014

Table 3-1. Rocky Cape Group age constraints

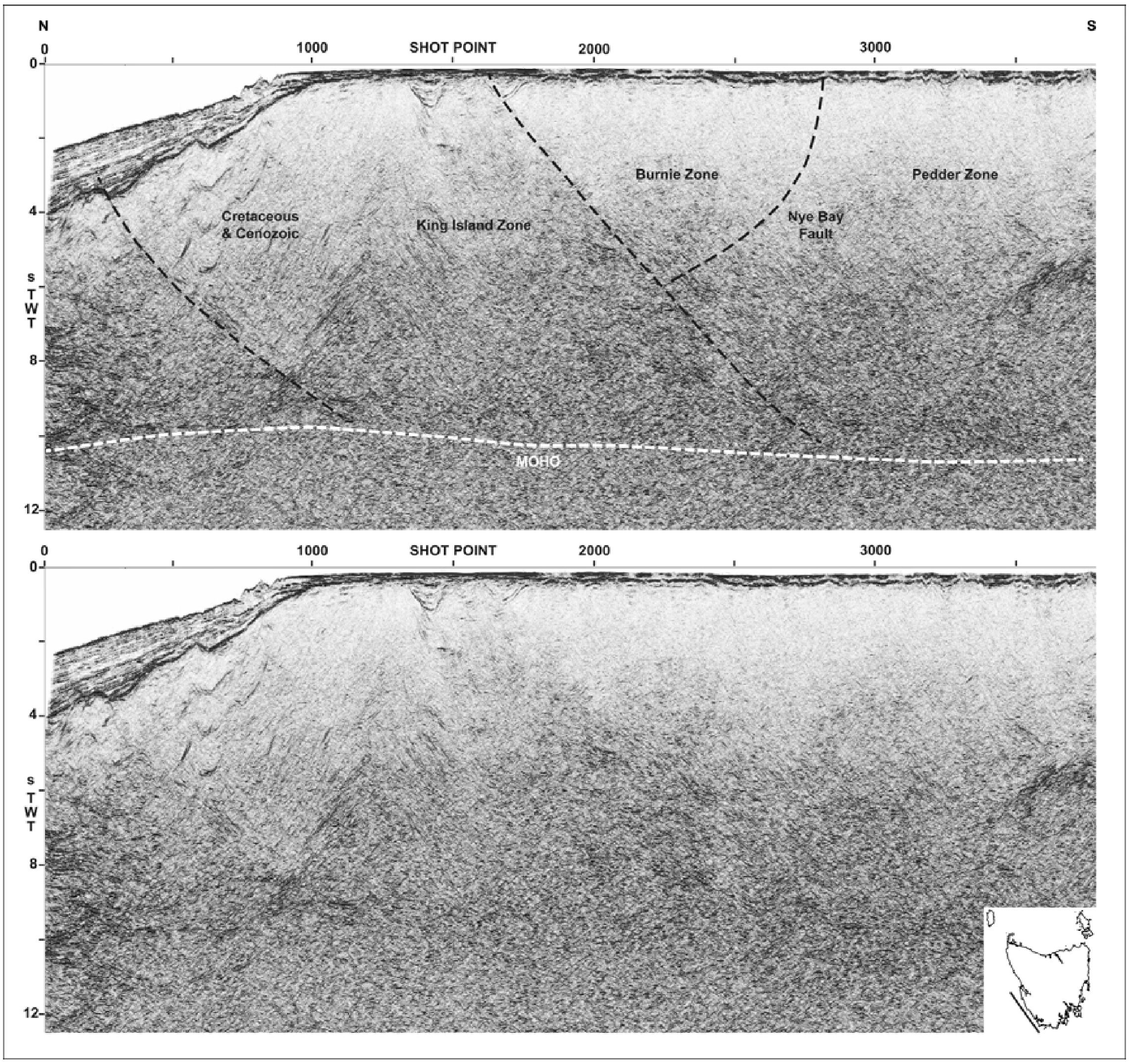


Figure 3-10. Seismic line AGSO 148/14, upper panel shows the interpretation, lower panel uninterpreted line, inset gives the line location. Shot points are approximately 50 m apart. The Nye Bay Fault has a gross normal sense, and an unpublished 40Ar/39Ar age determination on muscovite from the northern side of Nye Bay gave an age of 490.2±1.3 Ma (J.McL. Miller, written comm.), giving a possible time of late movement, approximately 15 m.y. later than the 505±2 Ma metamorphism of the rocks to the east (Chmielowski and Berry, 2012). The King Island Zone boundary terminates the Burnie and Pedder zones, implying that the King Island Zone was emplaced later.

Cape Group or similarly sourced rocks. The minimum age for the Burnie Formation is ca 710 Ma based on a K-Ar biotite date in an alkaline dolerite that was intruded into wet sediments (McDougall and Leggo, 1965, recalculated in Black et al., 2004), suggesting the Burnie Formation correlates with the basal successions of the Togari Group in the Rocky Cape Zone (Calver, 1998).

Further south, the Burnie Formation is overlain by a shallow water rift sequence of the Success Creek Group, and the deep water turbidites and tholeiitic basalt, chert, and mudstone of the Crimson Creek Formation (Brown, 1986). The Success Creek Group has been correlated with ca 700 Ma succession in the Smithton Basin based on similarities in the stromatolites present (Brown, 1986). The Crimson Creek Formation is considered to have been deposited on a rifting margin (Brown, 1986), with the basalts interpreted as seaward-dipping reflectors (Direen and Crawford, 2003a).

Boninitic mafic-ultramafic complexes within the Burnie Zone include dunite-harzburgite, layered pyroxenite-dunite or layered peridotite-pyroxenite-gabbro cumulates that are interpreted to have formed in a forearc settings (Crawford and Berry, 1992; Stern et al., 2012) or west-directed obducted slices from supra-subduction zones during the early phase of the Tyennan Orogeny (Stacey and Berry, 2004). These rocks represent a small component of the Burnie Zone (Berry and Crawford, 1988). The ages of these boninitic complexes are mostly poorly constrained, although Mortensen et al. (2015) gave a U/Pb zircon ca 516 Ma date for an associated gabbro in the Burnie Zone. The packages mark a gross change of tectonic environment from crustal extension to crustal shortening.

During the Middle Cambrian, in the second phase of the Tyennan Orogeny, the marine succession of the Mount Read Volcanics was deposited in north-south graben associated with east-west extension (Berry and Bull, 2012). The Mount Read Volcanics vary in composition from basalt to rhyolite and include intrusions, lavas and volcaniclastic metasedimentary rocks, shelf limestone and mudstone (Burrett and Martin, 1989; Corbett et al., 2014) deposited between ca 507 and ca 495 Ma (Seymour et al., 2013; Mortensen et al., 2015). Berry et al. (2008) gave a depleted mantle model age (T_{DM}) of approximately 1670 Ma from two felsic porphyries, suggesting a late Paleoproterozoic mantle extraction age. In the Late Cambrian, the Owen Conglomerate and associated rocks were deposited in half graben (Noll and Hall, 2005). By the Middle Ordovician, sedimentation had evolved from coarse clastics to fine clastics and then to the micritic dolomitic limestone of the Gordon Group (Seymour et al., 2013).

3.2.4 PEDDER ZONE

The Pedder Zone and the Tyennan Zone form the Tyennan Element of Seymour and Calver (1998) (Figures 3-1, 3-3 and 3-14 to 3-16). The Pedder Zone is differentiated

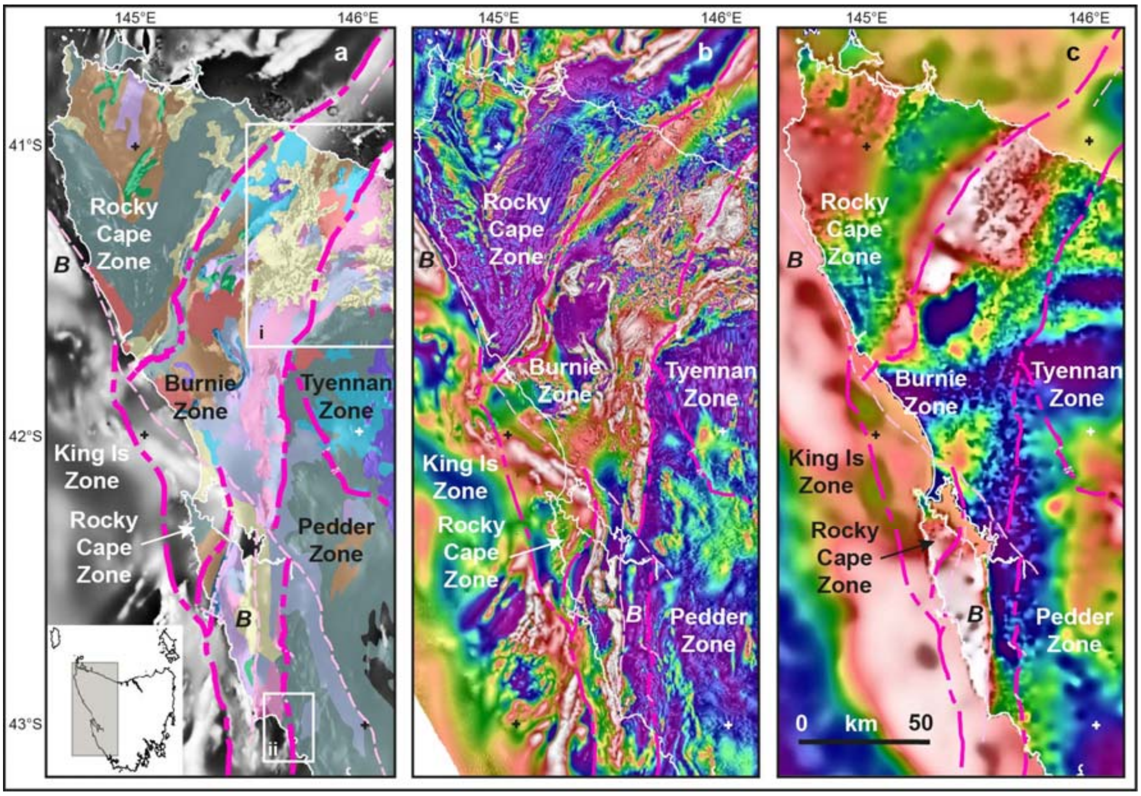


Figure 3-11. Burnie Zone. a. Geology on background grey scale total magnetic intensity. Location given by the shaded area of the map of Tasmania. Colour legend as for Figure 3-1, with dark blue Cambrian ultramafic rocks. B marks the Braddon River Fault. b. Tilt filtered total magnetic intensity. c. Gravity, isostatic onshore, free air offshore. The western boundary of the Burnie Zone is the Arthur Complex, indicated by the east-dipping magnetic and gravity highs. The eastern boundary is the edge of the Tyennan (in the north) or Pedder zones (in the south), in the central and southern parts typically with gravity responses of the order of 100 µm s-2 higher than the adjacent parts of the Burnie Zone. Further north the gravity low bifurcates, with the eastern arm marking the Cambrian dial Range Trough. The northern arm of the gravity low continues to the Tasmanian north coast. This coastal area marks the change from weakly to moderately deformed Neoproterozoic Burnie Formation metaturbidites in the west to strongly deformed metaturbidites and metaconglomerates, altered ?Cambrian MORB or back-arc basalt (Motton Spilite), tectonic mélange, sedimentary megabreccia, deep marine and volcanic-sourced chert (Figure 3-13) (Crawford and Berry, 1992; Seymour and Vicary, 2010); the boundary is also located at a possible east-dipping remnant subduction zone (Rawlinson et al., 2010). We have taken an outcropping fault on the western edge of this region as the eastern boundary of the Burnie Zone and have named it the Penguin Fault (Figure 3-13d). Over the 10 km east from the town of Penguin, the major faults generally dip to the west-northwest at about 70°. We infer that the many west-dipping faults to the east of are antithetic back-thrusts to the Penguin master fault. The deformation is older than the Lower to Middle Ordovician Moina Sandstone, which unconformably overlies many of the above lithologies. The eastern arm of the gravity low marks the Fossey Mountain Trough, a Cambrian feature largely comprised of Mount Read Volcanics. Boxes i and ii show the locations of Figure 3-12 a and b, where the boundary is detailed further.

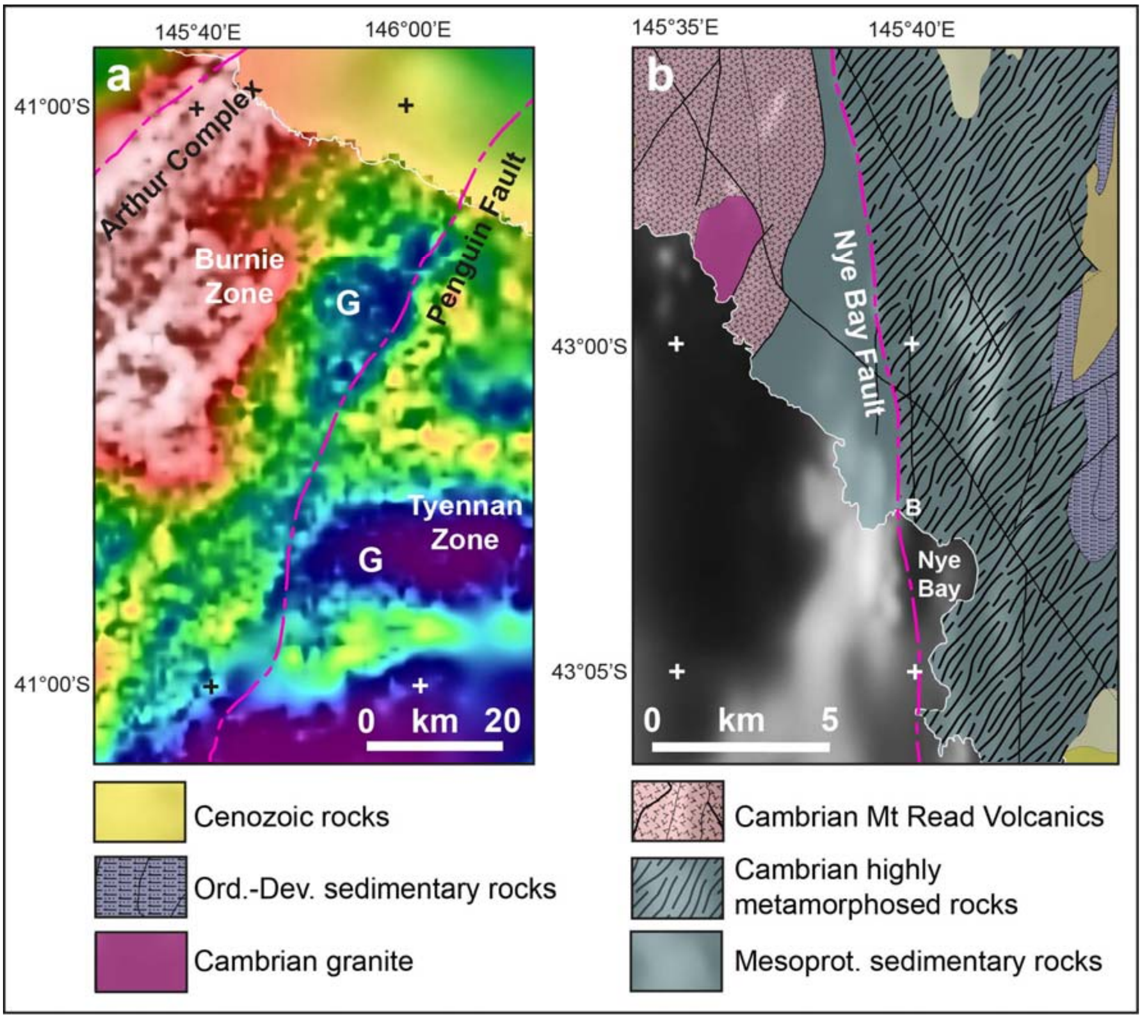


Figure 3-12. a. Northern Burnie Zone, isostatic gravity onshore free air offshore. Gravity lows labelled G are from granites. The gravity low defining the eastern edge of the Burnie Zone can be seen just west of the Penguin Fault. b. Southern end of the Burnie Zone, Nye Bay region. 1:250 000 scale geology from Mineral Resources Tasmania (2011). The dashed line marks the position of the interpreted boundary between the Burnie and Pedder zones, the Nye Bay Fault. This has juxtaposed rocks variously containing garnet, plagioclase ± kyanite and possibly glaucophane, at B, (C. Venn, pers. comm.) to the east of phyllites. Monazite ages from the higher grade rocks indicate metamorphism at 505±2 Ma (Chmielowski and Berry, 2012). The boundary places the Port Davey Metamorphic Complex in the Pedder Zone, with the lower grade Precambrian quartzites, phyllites and mafic volcanic rocks to the northwest in the Burnie Zone. It also retains the Mount Read Volcanics as a cover sequence that generally overlies the Burnie Zone. An unpublished 40Ar/39Ar age determination on muscovite from the northern side of Nye Bay gave an age of 490.2±1.3 Ma (J.McL. Miller, written comm.), giving a possible time of late movement. Locations of images indicated on Figure 3-11.

from the Tyennan Zone by the presence of the eclogite facies rocks of the Franklin Metamorphic Complex (Figure 3-15a). Metamorphic analysis of these eclogites yielded pressures between 1400 to 1960 MPa (depths of~40 to 60 km) and temperatures of ~550° to 650°C (Chmielowski and Berry, 2012), and so are likely to have formed at a boundary between cratonic blocks rather than within the interior of a single block.

Calver et al. (2006) interpreted the oldest rocks in the Pedder Zone as clast-bearing proximal turbidites. A mylonite within this package yielded a typical 1400 to 1900 Ma detrital zircon population, but some had metamorphic overgrowths that yielded a date of 1220±36 Ma (Chmielowski, 2009), suggesting the metaturbidites were either metamorphosed at 1220 Ma, or derived from rocks that had been metamorphosed then.

Figure 3-13. Traverse west along the northern Tasmanian coast showing the trasition across the western edge of the Burnie Zone (a) Simple anticline in the Burnie Formation, outlined by a dark shale bed. Location 41°4'9"S, 145° 56'22"E, on the eastern edge of Burnie. (b) Lighter coloured Burnie Formation intruded by dark dolerite; the Burnie Formation has been bleached by the interaction with water heated by the dolerite intrusion. Location 41°2'35"S, 145°53'E, east of Cooee. K/Ar dating on biotite from similar mafic rocks gave an age of 711±16 Ma. (c) Complexly deformed Burnie Formation just west of the Tyennan-Burnie boundary. White scale is 16 cm long. Location 41°6'16"S, 146°3'51"E, west of Penguin. (d) Multiple episodes of quartz veining in the Penguin Fault (at the Tyennan-Burnie boundary) some of which show a sinistral movement sense. Brecciation and a network of quartz veins is present over approximately 270 m across the dominant strike of veining. Location 41°6'19"S, 146°3'56"E, approximately 150 m southeast of (c). (e) Folded conglomerate and sandstone east of the Tyennan–Burnie boundary; view is to the south. Location 41°8'12"S, 146°8'12"E, south of Goat Island, approximately 3 km west of Ulverstone.

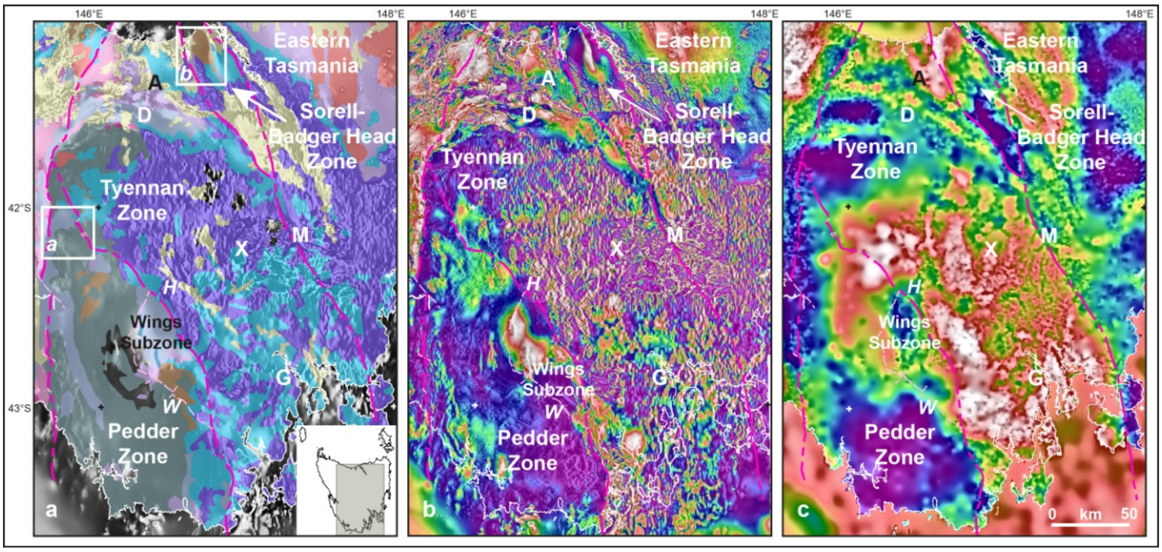


Figure 3-14. Central Tasmania, showing the Pedder, Tyennan and Sorell–Badger Head zones. a. Geology with a greyscale background of the total magnetic intensity; colours as for Figure 3-1A. Inset gives the location of the images. b. Tilt filtered total magnetic intensity. c. Isostatic gravity onshore, free air gravity offshore. *H* marks the Mt Hobhouse Fault, *W* the Mt Wedge Fault, the western boundary of the Wings Subzone. A marks the Avenue Road Inlier and D the Fossey Mountain Trough. The boundary between eastern and western Tasmania is constrained by drill holes at G that intersected andesite similar to that in the Mount Read Volcanics (Crawford and Berry, 1992), at M that intersected sheared Mathinna Group (Clarke and Farmer, 1983), and at X that intersected probable Proterozoic dolomite (Reid et al., 2003). The western boundary of eastern Tasmania is also indicated by the rise of approximately 60 µm s^{-2} in the isostatic gravity response in western Tasmania. The boundary is also visible in the offshore seismic data as a feature that dips east at approximately 30° from 2 to 8 s TWT (approximately 6 km to 24 km depth, Figure 3-17) (Barton, 1999). The western boundary of the Sorell–Badger Head Zone is adjacent to an outcrop of blueschist facies rocks (Figures 3-14 box b, Figure 3-15b) (Calver and Reed, 2001). The boundary between the Pedder and Tyennan zones, the east-dipping Mt Hobhouse Fault (Moore et al., 2012b) is seen in the offshore seismic data in southern Tasmania (Figure 3-16) (Drummond et al., 2000) and in the long wavelength magnetic data. This fault has generally been placed along mapped faults; it also lies east of all of the ultramafic rocks in the Pedder Zone, placing it east of the Wings Subzone that has been inferred to have been emplaced westwards (Crawford and Berry, 1992). It is also interpreted to lie adjacent to the only eclogite facies metamorphic rocks of western Tasmania (Figures 3-14 box a, Figure 3-15a).

The shallow marine to sub-aerial succession of the Clark Group overlies these metaturbidites, and the similarities in lithologies and detrital zircon populations suggest correlation with the Rocky Cape Group (Black et al., 2004; Calver et al., 2006). Several samples contained a metamorphic monazite population of 1367±7 Ma (Chmielowski, 2009). Metaquartzite, dolomite and diamictite are in fault contact with, and are interpreted to overlie the Clark Group. These rocks are correlated with the Togari Group (Calver et al. (2006).

The oldest Tyennan ages of metamorphic monazites in the Franklin Metamorphic Complex are 529±10 Ma from a muscovite-quartz-garnet-plagioclase schist, and the youngest 505±7 Ma from a quartz-muscovite-plagioclase-garnet-biotite schist; most zircon and monazite ages cluster at approximately 510 Ma (Chmielowski and Berry, 2012; Fergusson et al., 2013). On the southwest coast near Nye Bay (Figure 3-12b), the metamorphism is slightly younger, with monazite ages clustering at 505±2 Ma (Chmielowski and Berry, 2012).

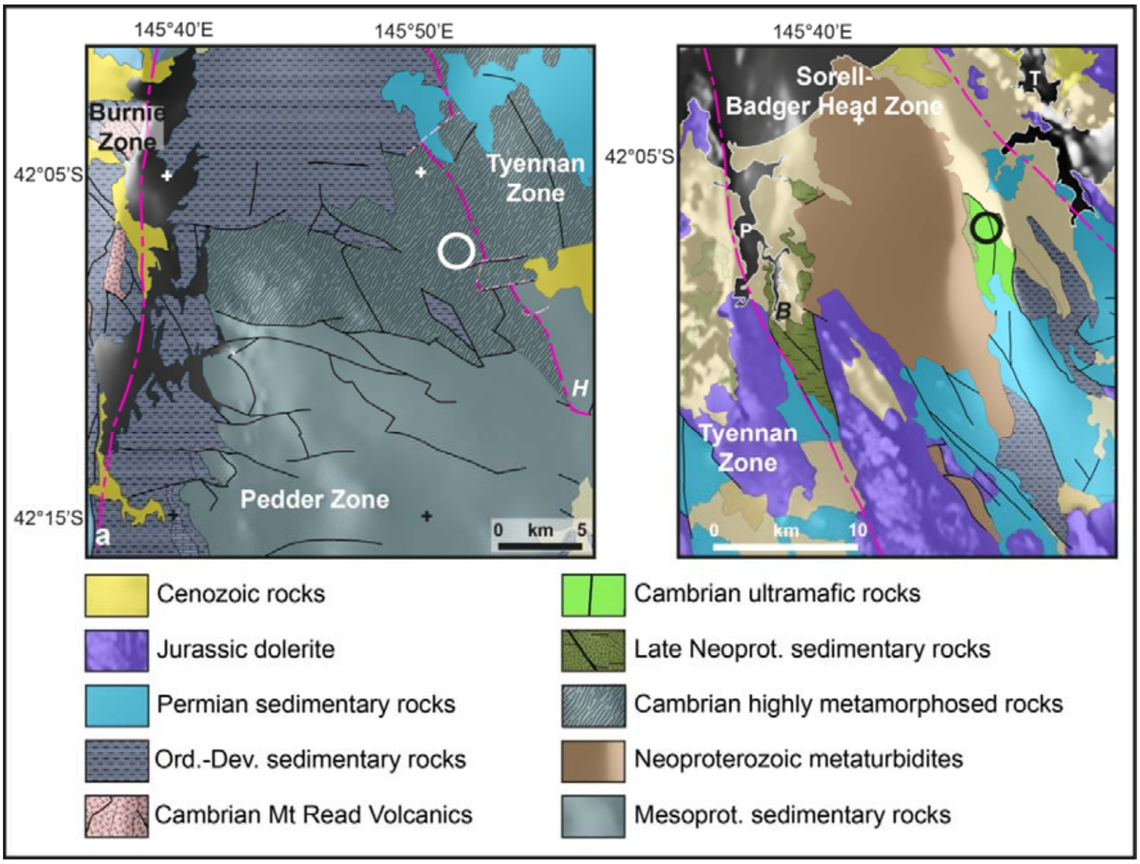


Figure 3-15. a. Northern Pedder Zone and adjacent areas, with the location circled of the eclogite in the Franklin Metamorphic Complex described by Chmeilowski and Berry (2012), approximately 20 km away from the boundary of the Burnie Zone. H marks the Mt Hobhouse Fault. b. Detailed geology of the northern Sorell–Badger Head Zone overlaid on the total magnetic intensity. The circle indicates the location of a small outcrop of ~650 Ma granite. B marks the location of blueschist facies rocks described by Calver and Reed (2001), P marks the Port Sorell area and T the Tamar River. The magnetic data indicate that the Andersons Creek Ultramafic Complex (green) extends at depth both north and south of the outcrop area. Locations of (a) and (b) in Figure 3-14a. The geology was modified from Mineral Resources Tasmania (2011).

Wings Subzone

The southeastern part of the Wings subzone is comprised mostly of the Clark and the Weld River groups, broadly equivalent to the Rocky Cape Group and the Smithton Basin rocks (Calver et al. 2006). Further north and west lies the Wings Sandstone, characterised by an unusual detrital zircon population with dominant populations between ca 900 and ca 1400 Ma, with the youngest population 914±44 Ma (Black et al. (2004). Ediacaran to Cambrian lithic volcanic metasediments, chert and minor basaltic tuff of the Ragged Basin, and slices of ultramafic rocks occur adjacent to the Wings Sandstone (Calver et al. (2006). There are no direct age controls on these east-dipping ultramafic slices; however, Crawford and Berry (1992) suggested they were obducted at the same time as other Tasmanian mafic-ultramafic complexes at ca 515 to 520 Ma. Unconformably overlying all the older sequences are stratigraphic equivalents to the Owen Conglomerate, which are conformably overlain by Ordovician carbonate successions. The subzone has been tentatively placed in the Pedder Zone as it lies west ofthe Mt Hobhouse Fault, although it could easily be placed in the Tyennan Zone.

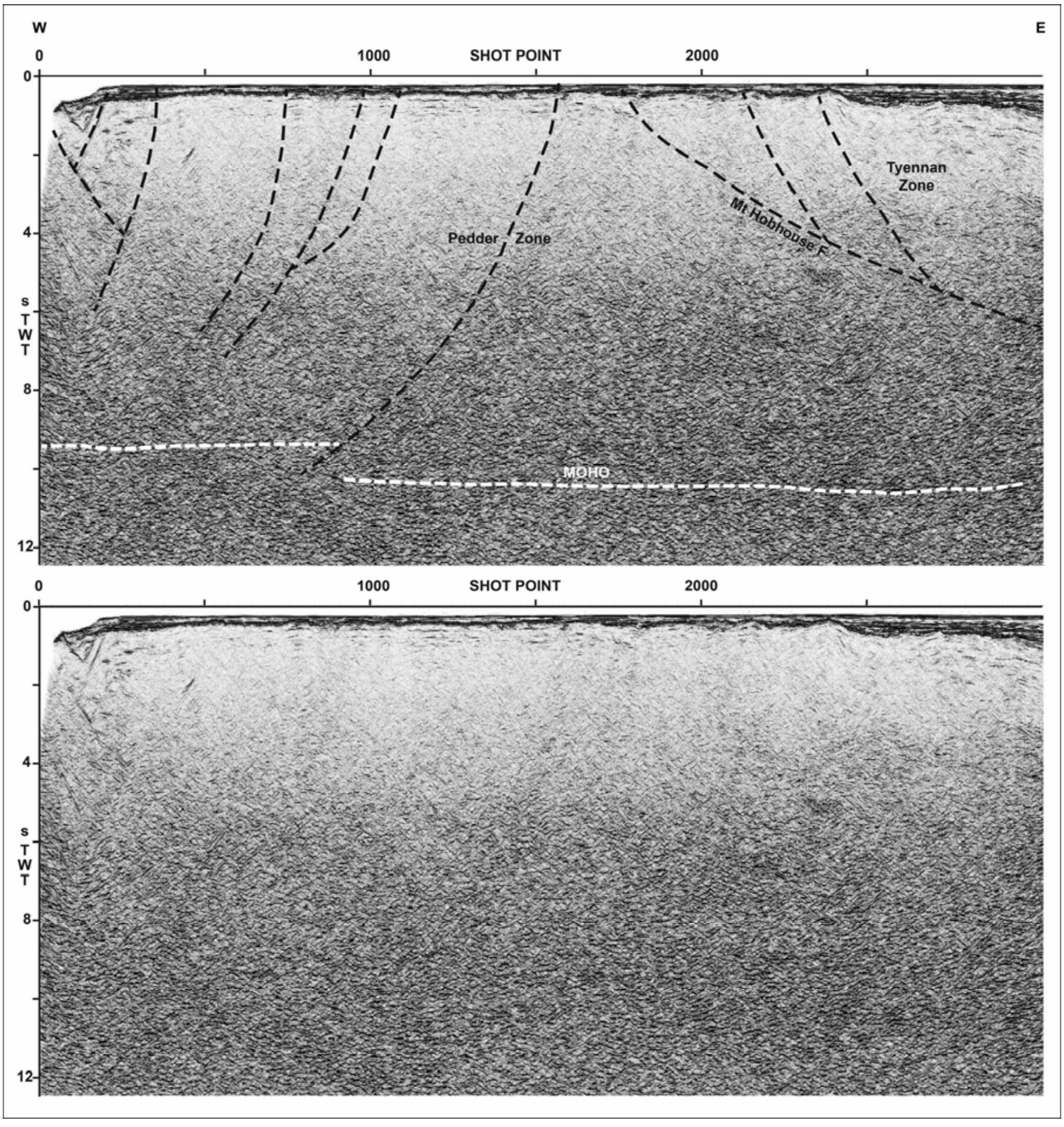


Figure 3-16. Seismic line AGSO 148/15, upper panel shows interpretation based mostly on that by Drummond et al. (2000), lower panel uninterpreted line, inset gives the line location. Shot points are approximately 50 m apart. The Mt Hobhouse Fault forms the boundary between the Pedder and Tyennan zones.

3.2.5 TYENNAN ZONE

The oldest mapped rocks in the Tyennan Zone are marginal marine metaquartzite and garnet schist, with a maximum deposition age of 1574±59 Ma based on the youngest detrital zircon population (Black et al., 2004). Like the Rocky Cape Block, the basement to this package is unknown. Metamorphic conditions for these rocks reached 1960 MPa and 545°C at 508±9 Ma (Chmielowski, 2009; U/Th/Pb monazite age) in a region near our interpreted boundary with the Pedder Zone. Drilling in central Tasmania intersected dolomite that was correlated with dolomite in the Togari Group of the Rocky Cape Zone (Figure 3-14) (Reid et al., 2003).

The metaquartzites are overlain by metaturbidites that have previously been interpreted as Burnie Formation or its equivalents (Vicary et al., 2008). Along the north coast, the poly-deformed Goat Island Conglomerate contains a large proportion of quartzite cobbles to boulders (Figure 3-13e) (Berry and Gray, 2001), suggesting, at least locally, significant paleo-topography and erosion from the quartzites now exposed to the south. The timing of deposition of this package pre-dates metamorphism and deformation at ca 510 Ma (Chmielowski and Berry, 2012). Also present on the north coast of Tasmania are tectonic mélange, mafic-ultramafic complex rocks, and MORB basalt (Motton Spilite) that overlies Cambrian chert (Seymour and Vicary, 2010; Everard and Calver, 2014). It is suggested that at least some of these rocks were deposited either prior to or in the early stages of the Tyennan Orogeny.

The Tyennan Zone was extensively deformed and metamorphosed during the Tyennan Orogeny, and formed the Ulverstone/Forth Metamorphic Complex. The Forth Metamorphic Complex was metamorphosed to peak temperatures of 700°C and pressures of 1690 MPa (Meffre et al., 2000; Chmielowski and Berry, 2012). Metamorphism has been dated at 512±5 Ma (SHRIMP, Black et al., 1997 recalculated by Foster et al., 2005) and 508±2 Ma ($^{40}Ar/^{39}Ar$, Foster et al., 2005). The metamorphic grade increases in the footwalls of west-dipping faults and Berry and Gray (2001) suggested the entire sequence was inverted and allochthonous. We are not persuaded by this interpretation. Their mapping and the cross section in Foster et al. (2005) show the metamorphic rocks in faulted contact with weakly metamorphosed packages, implying that the metamorphism took place before the faulting. If there was inversion, the sequence most likely came from the west. The Arthur Metamorphic Complex, at the western edge of the Burnie Zone, is characterised by temperatures and pressures of 350°C and 700 MPa (Turner and Bottrill, 2001), which are significantly lower than those of the Forth Metamorphic Complex, suggesting they are not equivalents; other alternatives seem equally unlikely.

While the southern part of the boundary between the Burnie and Tyennan zones is well defined, the northern end is less obvious, as it is mostly overlain by the cover sequences of the Cambrian-Ordovician Fossey Mountain Trough and intruded by Devonian plutons (Mineral Resources of Tasmania, 2011). However, images of the isostatic gravity (Figures 3-12a, 3-14c) are consistent with the boundary continuing northwards to a 200 m-wide breccia zone on the coast that is named the Penguin Fault (Figure 3-13d). This breccia zone also appears to separate the relatively uniform Neoproterozoic Burnie Group in the west from the complex ?Cambrian suite of conglomerate, mélange, mafic volcanic rocks and chert in the east. The proposed boundary also places an isolated outcrop of polydeformed metaquatzite schist and dolomite that lies north of the Fossey Mountain Trough (McClenaghan and Vicary, 2005) in the Tyennan Zone. Elsewhere, drilling near Hobart intersected high-Al basalt that has similar geochemistry and petrology to basalt in the Mount Read Volcanics (Crawford and Berry, 1992).

3.2.6 SORELL–BADGER HEAD ZONE

Metaturbidites at Badger Head, in the eastern part of this zone (Figures 3-14, 3-15b and 3-17), contain a youngest detrital zircon population of 1242±29 Ma (Black et al., 2004), implying syn- or post-Grenville Orogeny deposition. The rocks have been correlated with the Burnie Formation turbidites of the Burnie and Tyennan zones (Gee and Legge (1979). They are metamorphosed to lower greenschist facies, and locally contain retrogressed garnet (Reed et al. (2002).

To the west of these metaturbidites are mudstone, dolostone, volcaniclastic sandstone, conglomerate, chert, dolerite, rift tholeiite and rare rhyolite. These rocks are also associated with bedding-parallel broken formation (Calver and Reed, 2001). Microfossil-bearing chert indicates deposition during the Cryogenian, which is supported by late Neoproterozoic δC^{13} values from these rocks (Calver and Reed, 2001). The sequence may correlate with the Crimson Creek Formation of the Burnie Zone (Reed et al., 2002).

The Andersons Creek Ultramafic Complex lies immediately east of the metaturbidites at Badger Head (Figure 3-15b). Magnetic and gravity modelling by Zengerer (1999) suggested the complex lay in the hangingwalls of east and west-vergent thrusts. The west-dipping thrust shallows beneath the Badger Head rocks, whereas the east-dipping thrust extends beneath Cambrian to Devonian platform sequences. Highly deformed granite in the apex of the ultramafic complex yielded U-Pb SHRIMP dates of 658±5 Ma and 661±8 Ma (Black, 2007, unpublished data, OZCHRON database). The intrusion was metamorphosed to amphibolite facies at ca 510 to 520 Ma (U/Th/Pb on monazite, Berry et al., 2007).

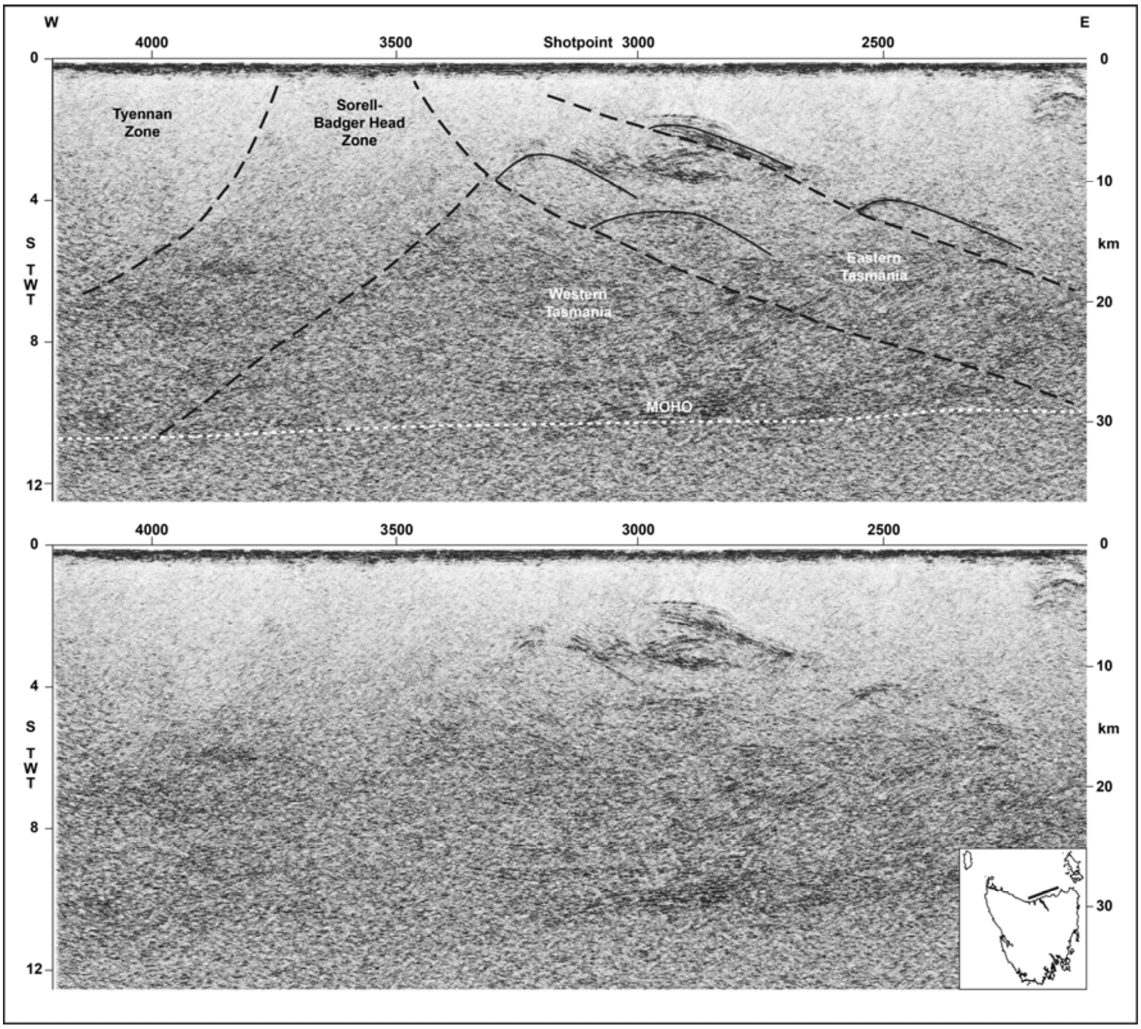


Figure 3-17. **West end of TASGO Seismic Line 4, northern Tasmania, showing the boundaries of the Tyennan, Sorell–Badger Head and Eastern Tasmanian zones. The ramp anticlines in the Eastern Tasmanian Zone may reflect mafic volcanic packages. Older west dipping faults have been truncated by younger (Tabberabberan) faults. Dashed lines mark interpreted faults, solid lines marker horizons. No vertical exaggeration. Insert shows the location of imaged data. Shot points are approximately 50 m apart.**

Reed et al. (2002) proposed a complex tectonic model involving two intersecting subduction zones during the Tyennan Orogeny. The model assumes that the rocks of the Sorell–Badger Head Zone were allochthonous and not connected to western Tasmania prior to the Tyennan Orogeny. In this model, Reed et al. (2002) suggested the change between eastern and western Tasmania is a simple facies change from shallow-water succession in the west to deep-water succession in the east. This model does not account for the presence of the 658±5 Ma granite in the footwalls of both thrusts nor is it easily reconciled with the geology of the other tectonic zones throughout Tasmania and the regional west-dipping subduction system outboard of Tasmania (Stump et al., 2003; Cawood, 2005; Squire and Wilson, 2005) . We accept the interpretation that the Badger Head metaturbidite package is allochthonous (Reed et al., 2002) and has been thrust eastwards above the ultramafic rocks. However, we prefer the interpretations by Powell and Baillie (1992) or Patison et al. (2001) that infer the presence of concealed Precambrian cratonic crust east of the Andersons Creek Ultramafic Complex as a basement to the Cambrian and Ordovician shelf sequences there, and consider that the eastern boundary of the cratonic crust lies under the Tamar Graben (Rawlinson et al., 2010; Young et al., 2011).

3.2.7 GLOMAR ZONE

Royer and Rollet (1997) and Berry et al. (1997) first outlined the boundaries, nature and positions of the thinned continental crust that lies to the south of Tasmania. Based on the submarine topography, Royer and Rollet (1997) subdivided the region into three areas, the west South Tasman Rise, the east South Tasman Rise and the East Tasman Plateau (Figure 3-18). Reconstructions by Exon et al. (1997a), Royer and Rollet (1997) and Norvick and Smith (2001) indicated that the East Tasman Plateau was rifted from the east South Tasman Rise in the Late Cretaceous. In order to avoid confusion with their terminology, we have grouped a slightly redefined east South Tasman Rise and the East Tasman Plateau into the Glomar Zone, after the ship that recovered the first pre-Mesozoic rocks in the region.

Exon et al. (1995) briefly described the results from four cruises in the region, three of which had recovered pre-Mesozoic samples from 42 dredge sites or drill holes. Subsequent age dating on 14 samples has allowed initial conclusions to be drawn about the South Tasman Rise. The east South Tasman Rise is topographically smoother and the bathymetry is ~1000 m. Seismic interpretations indicate the presence of up to 2 s TWT (~3 km) thickness of Cretaceous and Cenozoic sedimentary rocks (Exon et al. (1997b). Basement is interpreted to have been attenuated during Gondwana break-up; however, the present-day expression is consistent with separation during the Tasman rifting event. The free air gravity response has a strong linear pattern that trends at about 120°, although along its western edge the gravity anomalies trend parallel to the boundary with the west South Tasman Rise.

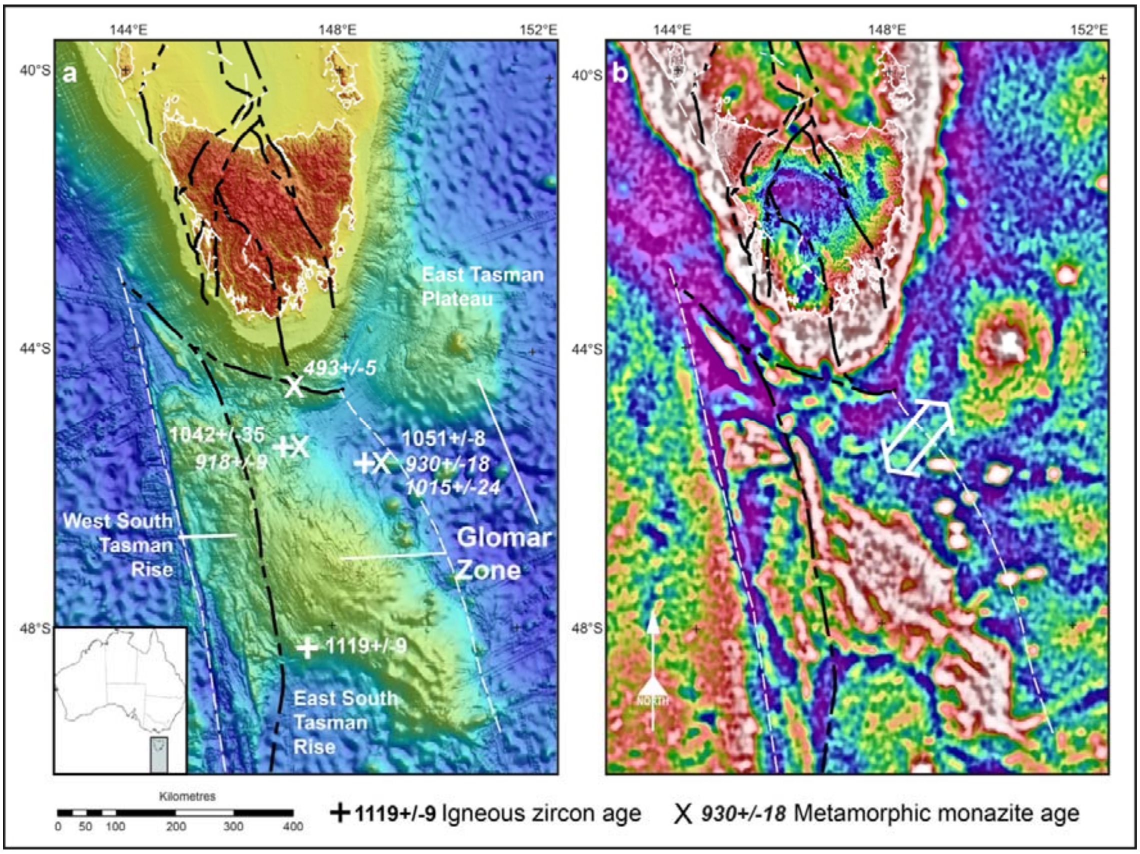


Figure 3-18. Glomar Zone and adjacent areas, with zone boundaries indicated by dashed black lines. a. Bathymetry, topography and locations of selected dredge samples with age determinations. The west South Tasman Rise has only yielded detrital zircons and metamorphic age determinations most like those from the Delamerian Orogen (Berry et al., 2007) and so has been excluded from the Glomar Zone. b. Gravity (Bouguer onshore, Free Air offshore). In comparison to the west South Tasman Rise, the east South Tasman Rise is topographically higher, its gravity responses are dominated by a northwest-trending fabric and dredge samples have yielded several Proterozoic age determinations (Fioretti et al., 2005a; Berry et al., 2008). The shape of the boundary with the Pedder Zone, as determined from dredge samples, bathymetry and the gravity, is consistent with a north dip. Monazite in a mylonitic gneiss on the boundary gave a metamorphic age of 493±5 Ma (Berry et al., 2008). The open arrow indicates the approximate direction of rifting of the East Tasman Plateau from the South Tasman Rise during the opening of the Tasman Sea. The northwest gravity trends also result from this rifting.

Cores in zircons from deformed granites yielded dates of 1051±8 Ma and 1042±35 Ma, and metamorphic monazites gave dates between 1015±24 Ma and 920±7 Ma, (Berry et al., 2008). An undeformed syenite in the southwestern Glomar Zone yields a U-Pb SHRIMP date of 1119±9 Ma and a T^{Nd}_{DM} age of 1270 Ma (Fioretti et al., 2005a). Originally this location had been considered as part of the west South Tasman Rise, but the bathymetric, gravity and geological data (Figure 3-18) from the area have more in common with other parts of the east South Tasman Rise. All of these ages are unlike any from mainland Tasmania. As a result we have chosen to regard the Glomar Zone as a separate geological entity. Samples from the East Tasman Rise include granite gneiss, quartzite and marble with blue amphibole, all rocks unknown in eastern Tasmania but the first two are also present in western Tasmania. No age date data are available.

The west South Tasman Rise is bathymetrically rough and typically lies in deeper water, mostly between 2000 and 4000 m. It also has a 200 μm^{-2} lower gravity response. In the southern part, linear gravity anomalies generally trend approximately 010°, but in the north, anomaly trends vary between 070° and 140°. Detrital zircons in a dredged metasedimentary sample suggests that at least part of the west South Tasman Rise can be correlated with the Delamerian Orogen of southeastern mainland Australia (Berry et al., 2007), suggesting that the west South Tasman Rise may be better excluded from VanDieland. An alternative interpretation is that the west South Tasman Rise is similar to the Kanmantoo Group in South Australia or the Robertson Bay Terrane in north Victoria Land, both of which have Proterozoic deep crust or mantle at depth (Handler et al., 1997; Fioretti et al., 2005b; Rawlinson et al., 2014).

3.3 DISCUSSION

3.3.1 TASMANIAN ISSUES

Despite the apparent structural, metamorphic and stratigraphic complexity outlined above, there are many geological features that are shared across many of the tectonic zones (Figure 3-3). For example, the oldest exposed rocks are generally quartzites or quartzite-rich sequences with a minor but persistent detrital zircon population at 1380 to 1450 Ma. These are generally marginal marine facies, suggesting they were deposited on an older cryptic basement. Neoproterozoic extension occurred between ca 780 Ma and 580 Ma and is characterised by rifting, sedimentation and intermittent granite production. During this period western Tasmania was disaggregated into smaller crustal blocks separated by MORB and mafic-ultramafic complexes. The rifted blocks were deformed in the Tyennan Orogeny. Early west- or south-directed movement was accompanied by regional metamorphism, including blueschist facies in the Arthur Metamorphic Complex and eclogite facies along the extension to the Mt Hobhouse Fault (Figures 3-14, 3-15a and 3-16). The Mount Read Volcanics and the Owen Conglomerate were deposited during the waning stages of the Tyennan Orogeny across several tectonic zones, suggesting western Tasmania had reconfigured into a single crustal block. Subsequent deformation in the Tabberabberan Orogeny modified the boundaries and internal structures but pre-existing structures can still be recognised.

Despite these similarities there remain other observations that need to be addressed in order to develop a comprehensive tectonic model of VanDieland. These are discussed below.

What do the boundaries reflect?

Most of the boundaries between the tectonic zones are defined by fault zones. However, several boundaries, such as the Tamar ‘Lineament’ and the Arthur Metamorphic Complex, are more distributed fault networks. Several of the boundaries may separate tectonic zones that are completely allochthonous with respect to their neighbouring zones (e.g. King Island Zone). Based on comparison with other regions (e.g. the north Atlantic, Péron-Pinvidic and Manatschal, 2010; the Southern Ocean, Direen et al., 2012), some boundaries may represent the inverted remnants of variably thinned crust reflecting that, at the end of the Proterozoic, western Tasmania was a series of crustal scale megaboudins.

The diversity in character and length of these boundaries suggests that different processes were active at the same time along the boundaries. For example, the high pressure (to 1.9 GPa) metamorphism seen in the northern part of the Pedder-Tyennan boundary is consistent with craton-craton collision. By contrast, the southeastern segment of this boundary is characterised by lower metamorphic grade Proterozoic rocks (Calver et al., 2006) that do not indicate collisional tectonic processes. We suggest the Neoproterozoic to Early Cambrian rifting of the Pedder and Tyennan Zones resulted in ocean crust formation in the northern segment of the fault zone, while the southern zone was characterised by thin continental crust. We suggest a similar setting along the Rocky Cape-Burnie Zone boundary, where the Arthur Complex has attained blueschist facies metamorphic grades in the northern part (Turner and Bottrill, 2001), but the equivalent boundary to the south is defined by mafic-ultramafic units that underwent lower grade metamorphism (Corbett, 2003).

Arthur Metamorphic Complex

The Arthur Metamorphic Complex (Arthur Lineament) separates mostly shallow marine Proterozoic rocks of the Rocky Cape Zone to the west from largely deep water Proterozoic metasediments of the Burnie Zone to the east (Seymour and Calver, 1995). Holm and Berry (2002) mapped the Arthur Metamorphic Complex on both the western and northern Tasmanian coasts, and we interpret the boundary northward almost to the Victorian coast, and southward across the Sorell Peninsula (Figures 3-1, 3-2 and 3-4). Holm et al. (2003) divided the rocks in the zone into three packages. The westernmost package was similar to rocks in the adjacent Smithton Basin. The central package comprised the rift-related mafic schist, amphibolite and metagabbro of the Bowry Formation, which was intruded by ca 777±7 Ma granite (Turner et al., 1998) and metamorphosed to blueschist facies (350°C and 700 MPa, Turner and Bottrill, 2001). The eastern package comprised rocks similar to the older parts of the Burnie Zone.

Geophysical modelling of the Arthur Metamorphic Complex suggests that none of the outcropping magnetic units persist to depths of more than 2 km (Leaman and Webster, 2002). An east-dipping listric fault truncates the western edge of the package (Leaman and Webster, 2002). A body 10 to 15 km wide of weakly magnetic (0.01×10^{-3} SI), dense rocks (ρ=2080 to 2860 kgm^{-3}) is present under the Arthur Metamorphic Complex, suggesting the presence of either mafic igneous or high grade metamorphic rocks. Further south, deep seismic data are consistent with the eastern boundary of the Arthur Metamorphic Complex having a dip to the east of approximately 30° and continuing to at least 5 s TWT, about 15 km depth (Leaman and Webster, 2002).

Approximately 7 km offshore from the west Tasmanian coast, the Arthur Metamorphic Complex is displaced almost 50 km south (sinistrally) by the Braddon River Fault (Figures 3-1 and 3-8). On the Sorell Peninsula the boundary between the Rocky Cape and Burnie zones is an east-dipping thrust fault (Figure 3-8).

The dominant Cambrian movement sense in the Arthur Metamorphic Complex is of southwest-directed (sinistral) movement, with a later west-verging thrusting event (Holm and Berry, 2002). Both events took place after the obduction of the mafic-ultramafic complexes at about 516 Ma. We interpret the Arthur Complex as crust that was extended at ca 770 Ma and then shortened in the Tyennan Orogeny and further modified in the Tabberabberan Orogeny. An implication of this is that the western part of the Burnie Zone may be underlain by equivalents to the older rocks in the Rocky Cape Zone.

Origin of the mafic-ultramafic complexes

The understanding that VanDieland was a completely separate crustal fragment in the Cambrian (Cayley et al., 2011) has resolved the paradox of west-dipping subduction zones in Victoria (Miller et al., 2005) and Antarctica (Federico et al., 2006) lying either side of an east-dipping subduction zone in Tasmania (Crawford and Berry, 1992; Moore et al., 2012a). However, the question remains as to where the mafic-ultramafic complexes were obducted from.

Crawford and Berry (1992) considered that they came from an unseen but speculated Cambrian arc system that lay to the east of VanDieland. In this interpretation, mafic-ultramafic units were obducted at least 130 km westwards over the Tyennan and Pedder zones while leaving no evidence for their occurrence to the east, closer to their interpreted origin. We pose an alternative model, in which the mafic and ultramafic rocks with oceanic and transitional crust affinities were formed between micro-continental fragments that separated during 250 m.y. of Neoproterozoic extension. In the Tyennan Orogeny, these basins were inverted and the individual ribbons re-amalgamated to produce the geometries and lithological distributions in western Tasmania. The model does not require an arc system as proposed by Crawford and Berry (1992). Obducted mafic-ultramafic units in the Burnie Zone were derived from the inversion of thinned continental crust (perhaps with some oceanic crust) along the deep, east-dipping Penguin Fault (Rawlinson and Urvoy, 2006; Rawlinson et al., 2010) that forms part of the western boundary of the Pedder and Tyennan zones (Figures 3-1, 3-11 and 3-12a). This model requires obducted slices to travel ~30 km, not 130 km. Inversion of small transitional crust to oceanic back arc basins may not have initiated subduction, as inversion may have been accommodated by obduction and thrust slicing of the transitional and ocean crust substrate in a manner similar to that imaged in Bendigo Zone of the Lachlan Fold Belt (Cayley et al., 2011). The driver for this inversion may have been a convergent margin located outboard of Tasmania. A similar scenario might have seen the mafic-ultramafic complex rocks in the Tyennan Zone obducted west from the western boundary of the Sorell–Badger Head Zone.

An east-dipping mafic-ultramafic complex lies close to the western edge of Seymour & Calver's (1995) Adamsfield-Jubilee Element (Figure 3-1a). Mapping implies that another significant boundary lies west of the mafic-ultramafic complex (Calver et al., 2006). We interpret the mafic-ultramafic rocks and the rocks either side to have been within a leading-edge thrust system, with the master fault being the Mt Hobhouse Fault (Moore et al., 2012b). While it is somewhat arbitrary as to whether one considers the Adamsfield–Jubilee area part of the Tyennan Zone or the Pedder Zone, we have chosen the latter in order to emphasise the significance of the Mt Hobhouse Fault.

Mount Read Volcanics

The Mount Read Volcanics have been interpreted as a volcanic arc that formed above a west-dipping subduction zone during the late Cambrian (Crawford and Berry, 1992). In this model, a transient, west-dipping subduction zone formed to the east of the 'Tamar Lineament'. However, there seems to be little independent evidence for this subduction zone. Recent geochronology results by McNeill et al. (2012) from the Mount Read Volcanics demonstrate that they were formed over 12 m.y., apparently longer than the "short-lived" event envisioned by Crawford and Berry (1992), and potentially questioning the validity of their interpretation. Meffre et al. (2004) described an east-dipping sequence at least 15 km thick of a rift-related tholeiitic volcanic passive margin sequence that crops out on King Island. This magnetic sequence can be traced southwards approximately 50 km to the west of the Mount Read Volcanics (Figures 3-1 and 3-2). Inversion of this passive margin is likely to have resulted in the amalgamation of the King Island Zone with the rest of western Tasmania and, as this took place, the Mount Read Volcanics formed as the east-dipping subducting slab delaminated. We consider that this is a more satisfactory solution to the origin of the Mount Read Volcanics than that proposed by Crawford and Berry (1992).

Fossey Mountain Trough

The Fossey Mountain Trough comprises volcanic and sedimentary rocks of the Mount Read Volcanics and the overlying Tyndall Group, and Ordovician to Devonian sedimentary rocks. It forms the northern boundary of Seymour and Calver's (1995) Tyennan Element (Figures 3-1a). All of these rocks, including the Mount Read Volcanics (Corbett, 1992), were deposited in shallow to moderately deep water, implying that there must have been older crust below the Trough.

We suggest that the 'Tyennan Element' rocks extend north, under the Fossey Mountain Trough, to include the Avenue Road Inlier (Figure 3-14), forming the Tyennan Zone. This Inlier is comprised of polydeformed quartz mica schist, metaquartzite and dolomite (Mineral Resources Tasmania, 2011), similar to shallow-water lithologies found in many areas of Seymour and Calver's (1995) Mesoproterozoic Tyennan Element. We consider that the magnetic and gravity images of the region are dominated by Paleozoic features, and so do not give a clear representation of the Proterozoic basement geology on which the subdivisions here are based. The strong magnetic responses in the region (to 1000 nT) are from the Cambrian Mount Read Volcanics, the overlying Tyndall Group and Cenozoic basalt. The dominant linear gravity lows of approximately 100 μms^{-2} are from the Fossey Mountain Trough, and these are further overprinted by deeper lows of up to another 200 μms^{-2} from the ca 370 Ma Dalcoath, Housetop and Granite Tor granites. Hence, gravity responses of at most 20 μms^{-2} that are related to the northern part of the Penguin Fault (e.g. Figure 3-12a) are, at best, second order effects.

In this interpretation, the Fossey Mountain Trough may be considered as a generally east-west-striking late Tabberabberan syncline, unconformably overlying the Proterozoic outcrops to the north and south that mark the adjacent anticlines. These fold structures terminate against southern extension of the Penguin Fault, which continues south to form the western boundary of Seymour and Calver's (1995) Tyennan Element.

Wickham 'Orogeny'

The Tasmanian Cryogenian and Ediacaran geological record is dominated by events that can be interpreted as extensional (Figure 3-3) (Li, 2001). The Wickham 'Orogeny' was identified on the northern and western parts of King Island, where granite intrusion and associated widespread contact metamorphism and tholeiitic dyke emplacement took place at 760±12 Ma (Black et al., 1997; Calver, 2004; Calver and Everard, 2014), consistent with extension. These rocks are coeval with a 777±7 Ma granite and rift-related tholeiites in the Arthur Metamorphic Complex (Turner et al., 1998; Holm et al., 2003). Everard et al. (2007) showed deposition occurred into the extensional Smithton Basin from the middle Cryogenian (ca 750 Ma). We therefore consider that the Wickham 'Orogeny' is likely to represent a period of crustal extension rather than one of crustal shortening, and is likely to record Rodinia break-up in western Tasmania (Holm et al. (2003).

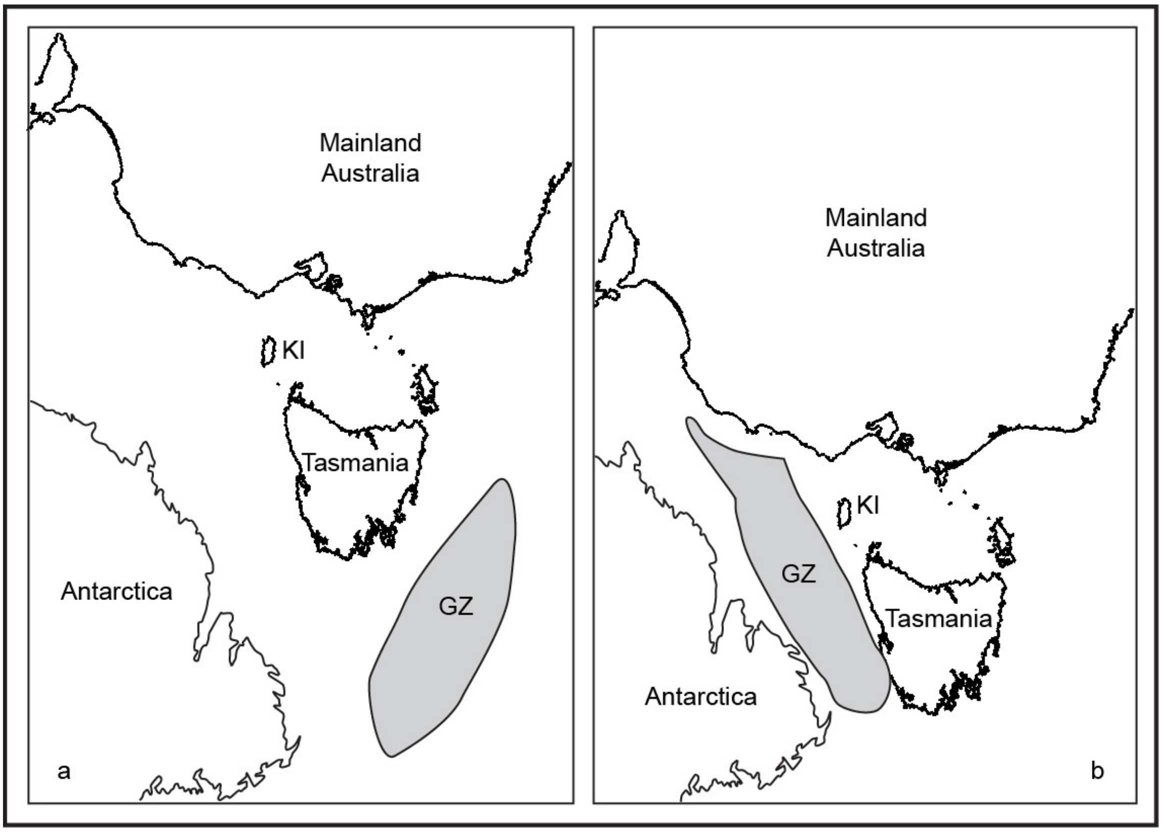


Figure 3-19. Gondwanan reconstructions by Norvick and Smith (2001), a, and Williams et al. (2012), b, showing theiir locations of the Glomar Zone. In both texts, the region is referred to as the South Tasman Rise.

Pre-Gondwana breakup location of the Glomar Zone

There is no consensus to the precise pre-breakup alignment of Australia and Antarctica. Several authors have favoured the alignment of the Darling Fault in Western Australia with the Denman Glacier in Antarctica (Fitzsimons, 2003; Finn et al., 2006; Goodge et al., 2010), whereas others have preferred a closer alignment of the eastern edge of the buried Gawler Craton with the western edge of Ross Orogen (Fioretti et al., 2005b; Boger, 2011; Lisker and Läufer, 2013), or intermediate locations (Williams et al., 2012;

Aitken et al., 2014). Placing smaller crustal fragments in the context of Rodinia and Gondwana is even more subjective because their origins and original geometries are poorly constrained, and so there is no consensus about the original position of the Glomar Zone. Norvick and Smith (2001) placed it to the southeast of Tasmania, while Williams et al. (2011) showed it to the west at about 160 Ma, and Royer and Rollet (1997) had it in its present location at 95 Ma. The reconstruction of Norvick and Smith (2001) (Figure 3-19a) seems unlikely, as it places +1050 Ma crust east of our interpreted position of the eastern margin of VanDieland Proterozoic crust. The reconstruction by Williams et al. (2012) (Figure 3 19b) appears to be equally unlikely, since it places the Glomar Zone, with granites with ages of 1050 Ma and metamorphic rocks at 920 Ma, adjacent to King Island, with metamorphism at 1290 Ma and granites at approximately 760 Ma, and there are no known thermal events between 1050 to 920 Ma on King Island. We therefore consider the most likely location for the Glomar Zone before Gondwana breakup is close to its present position south of Tasmania, as originally suggested by Royer and Rollet (1997). This interpretation may be supported by the presence of Cretaceous northwest-trending basins on the east South Tasman Rise (Exon et al. (1997b), which are parallel with the dominant trends in the free-air gravity anomalies from with the Glomar Zone (Figure 3-18). Furthermore, K/Ar ages determined on biotite in the Glomar Zone gave ages of 451±4 Ma and 511±4 Ma (Berry et al., 1997), consistent with Tyennan metamorphism. These data suggest there was not sufficient crustal extension or increase in the geothermal gradient to exceed the closure temperature of biotite, approximately 300°C, during Gondwana break-up.

3.3.2 POSSIBLE CORRELATIONS WITH VANDIELAND

A fundamental part of deriving a coherent geological history for VanDieland is to place it within the wider context of earth history. The Proterozoic events seen in VanDieland took place within the assembly and dispersal of Rodinia and the formation of Gondwana (Calver et al., 2014) and so events elsewhere should be reflected in the history of VanDieland. Because there is only one Proterozoic paleomagnetic data point of doubtful reliability from VanDieland (McWilliams and Schmidt, 2003), other data assume greater importance in determining the most likely relationships with other cratons. We accept the reservation outlined by Andersen (2014) that identical detrital zircon populations can form at the same time on cratons distant from each other. Nevertheless, even though these data are non-unique, they can exclude some possibilities. As well, where other data are available, we have attempted to integrate them as closely as possible into the discussion and Table 3-2.

Elsewhere in Australia

The provenance of the detrital zircons in Tasmania provides important constraints on where any Tasmanian crustal fragments may have come from. The dominant population, from approximately 1650 to 1900 Ma, is widespread in Australia and may have been derived from any of the Nuna assembly events (Evans, 2013). In contrast, the 1380 to 1460 Ma zircons seen in VanDieland are less common elsewhere (Condie et al., 2009) and so provide a tighter constraint. In Australia, the only possibilities appear to be the Western Australian Madura Complex (near the South Australian border and covered by Cenozoic sedimentary rocks), which is known to include meta-igneous rocks containing zircons with the appropriate ages (Nelson, 2006a, b) or the Musgrave Province in central Australia (Kirkland et al., 2013). However, if VanDieland has always been in generally the same position relative to the rest of Australia, as suggested in most Rodinian assemblies, e.g. Li et al. (2008b), the drainage required to transport the 1400 Ma zircons must have crossed the Mawson Craton, which includes the 1500 to 1600 Ma Hiltaba, Spilsby and St Peter Suites but there is little evidence of these older zircons in the VanDieland meta-sedimentary rocks. In contrast, 1500 to 1600 Ma zircons are abundant in the Neoproterozoic Adelaide Geosyncline metasedimentary rocks, which were also marginal to the Mawson Craton (Ireland et al., 1998; Preiss, 2000).

An alternative interpretation is that VanDieland was once adjacent to the Madura Province in southeasternmost Western Australia. However, the paleomagnetic evidence suggests that the Northern Australian, Western Australian and Mawson cratons have either maintained their present relative positions since 1500 Ma (Wingate and Evans, 2003), or at most rotated approximately 40° between 650 and 550 Ma (Li and Evans, 2011). Although this interpretation can be used to correlate the 1290 Ma metamorphism seen on King Island and the Rocky Cape Zone (Berry et al., 2005; Halpin et al., 2014) with that seen in the Albany-Fraser Orogen (Bodorkos and Clark, 2004), it raises other problems such as extracting VanDieland and then positioning the other cratons back into their previous locations. Detrital zircons found in the Anakie Inlier in central Queensland include a significant 1100 to 1300 Ma population (Fergusson et al., 2001), similar to those seen in the Sorell–Badger Head Zone and are coeval with the 1290 Ma metamorphism on King Island, but other populations typical of those found on King Island or in the Sorell–Badger Head Zone are not present, suggesting the two regions have little in common. The presence of detrital 1370 Ma monazite in VanDieland, an age absent from Australian terranes, adds yet another layer of difficulty in locating a possible Australian source area.

	Southwest Laurentia	**East Antarctica**	**Cathaysia**	**Western Baltica**
Mesoproterozoic or older quartzite source	Hondo & Vadito Groups (1)	Source of Cotton Plateau 'Goldie Formation' (5)	Shihuiding Formation and protoliths (10)	Same source as that of widespread quartzites (15)
Appropriate detrital zircon source	Granite-Rhyolite Province, Mazatzal Province (2)	1440 Ma A-type granite (6)	Baoban Complex (10)	Danopolonian Orogeny (15)
Appropriate detrital monazite source	Granite-Rhyolite Province (2)	1440 Ma A-type granite (6)	**Baoban Complex (10)**	Late Danopolonian Orogeny A-type granites? (15)
Sm–Nd model age basement approx. 1700 Ma	Yavapai & Mazatzal Provinces (2)	Central Transantarctic Mts outboard of Miller Range (7)	No; 2000 to 2300 Ma (10)	Mostly to 2100 Ma, with parts to 3500 Ma (16)
Midcrust with 1600 Ma zircon	Yes (2)	Terre Adélie Craton (8)	Unlikely; midcrust probably pre-1850 Ma (12)	Possibly in SW Fennoscandia (16)
Grenville-age boundary	Yes (2)	Near Beardmore Glacier? (7)	Most likely boundary deformed at 830 Ma (11)	Yes (16)
Rifting started at approx. 760 Ma	Yes, Chuar Group (3)	Probably, Cobham Formation base (5)	Initial rifting from 820 to 730 Ma (11)	Approximately 800 Ma (15)
Rifting completed at approx. 580 Ma?	No; probably 600 Ma (4)	Yes; compressional Ross Orogeny starts 560 Ma (9)	Continued to 540 Ma (13)	Approximately 620 to 550 Ma (17)
Other issues or problems	Places VanDieland on the far side of the Panthalassic Ocean in the Ediacaran	Quartzite source from under the icecap? (5)	Located off Western Australia? (13)	Most likely correlative, SW Fennoscandia, had semi-continuous zircon production from 1730 to 1480 Ma (16)
Reconstruction favored for Late Cambrian location of VanDieland	Goodge et al. (2010), a modification of Dalziel (1997)	Goodge et al. (2010), a modification of Dalziel (1997)	Li and Evans (2011)	Evans (2009)

Table 3-2. Possible correlatives of VanDieland. Data sources; 1 Jones et al. (2011), 2 Whitmeyer and Karlstrom (2007), 3 Timmons et al. (2001), 4 Macdonald et al. (2013), 5 Goodge et al. (2004), 6 Goodge et al. (2008), 7 Borg and DePaolo (1994), 8 Peucat et al. (2002), 9 Goodge et al. (2012), 10 Li et al. (2008a), 11 Zhao et al. (2011), 12 (Li et al., 2014), 13 Zhao and Cawood (2012), 14 Cawood et al. (2013), 15 Kheraskova et al. (2002), 16 Bogdanova et al. (2008), 17 (Andréasson et al., 1998)

A further difficulty is that the Mesoproterozoic and Neoproterozoic history of VanDieland seems a poor fit with other regions in Australia. There are many areas along the eastern Rodinian breakup margin where 1600 to 1900 Ma basement is present. However, none seem capable of generating the quartzites derived from what appears to have been a Grenville-aged source; all possibilities are either in Western Australia or have been previously eliminated by the problem of the 1400 Ma zircons (Betts et al., 2002).

Laurentia

We consider that the southwestern margin of Laurentia is a potential source area for much of the detritus in western Tasmania. The clean quartzites of both southwestern Laurentia and western Tasmania and equivalents contain a dominant detrital zircon population of 1700 to 1800 Ma, and traces of >2400 Ma zircons (Figure 3-20) (Black et al., 2004; Jones et al., 2009). As well, metasedimentary rocks from both regions host minor detrital zircon populations between 1380 and 1450 Ma (Black et al., 2004; Jones et al., 2011). Southwestern Laurentia was also a potential source of 1100 Ma igneous zircons (Bright et al., 2014) that are present in many of the western Tasmanian detrital zircon populations (Black et al., 2004). Because of the multiple sources suggested, it is not possible to do meaningful statistical analysis such as that outlined by Sircombe and Hazelton (2004). Metasedimentary rocks from central Tasmania also include 1370 Ma detrital monazites (Chmielowski, 2009) that could reasonably have been derived from the western Laurentian Granite-Rhyolite Province and other A-type granites in the Yavapai and Mazatzal orogens (e.g. Jones et al., 2011).

The excess of ca 1600 Ma inherited zircon populations in the Tasmanian Devonian granites relative to the host Rocky Cape Group (Black et al., 2010) suggests the zircon source from an unexposed VanDieland Mesoproterozoic basement that lies beneath the Rocky Cape Group. We suggest that this basement may correlate with the Yavapai-Mazatzal terranes of southwest Laurentia. Limited T_{DM} model age data of 1700 Ma or younger is consistent with data from the Cochise Block in the Mazatzal Orogen (Whitmeyer and Karlstrom, 2007).

VanDieland also shows evidence of Grenville-age events. Detrital zircon populations with ages ranging from 914±44 Ma to 1268±19 Ma (Black et al., 2004) are common, and the Glomar Zone includes granites with ages ranging from 1119±9 Ma (Fioretti et al., 2005a) to 1042±35 Ma, which were metamorphosed at 918±9 Ma (Berry et al., 2008). We suggest that the Glomar Zone correlates with the Grenville Front in southwestern USA, and that the Grenville-Glomar Zone zircons were subsequently eroded to be deposited in the upper Rocky Cape Group and the Badger Head Group. The ~770 Ma Wickham 'Orogeny' in VanDieland appears to be coincident with widespread rifting in Laurentia, in which the ca 800 to 740 Ma Chuar Group was deposited in Arizona (Timmons et al., 2001), and ca 780 Ma mafic intrusions were emplaced in a belt that extends from Montana to the Great Slave Lake (Harlan et al., 2003). The final breakup of the southwest Laurentian margin occurred at about 580 Ma (Yonkee et al., 2014), coeval with the 580 Ma breakup in VanDieland.

Halpin et al. (2014) suggested that the Rocky Cape Group in Tasmania could be correlated with the Belt-Purcell Supergroup in Laurentia. However, Medig et al. (2014) correlated the 1500 Ma Mt Isa granite event with detrital zircon populations found in the Coal Creek Inlier in the Yukon, and hence the Mawson Craton granites with input into the Belt-Purcell Supergroup. This correlation places potential Laurentia-East Antarctica correlates of VanDieland, much further to the south than the Belt-Purcell rocks. As well, the hypothesis of Halpin et al. (2014) seems to take insufficient account of the 1120 to 1040 Ma zircon igneous age dates (Fioretti et al., 2005a; Berry et al., 2008) and approximately 920 Ma metamorphic monazite dates (Berry et al., 2008) from granites from the Glomar Zone. Like Fioretti et al. (2005a), we prefer correlating the Glomar Zone with the Grenville Orogen of Laurentia.

East Antarctica

There are few clear links between VanDieland and the pre-1300 Ma rocks of East Antarctica, including the Mawson Craton (Payne et al., 2009). However, the prominent 1650 to 1900 Ma detrital zircon peak in the Tasmanian quartzites overlaps with the 1720 to 1730 Ma Nimrod Orogeny in East Antarctica (Figure 3-20); (Goodge et al., 2001). Clean metaquartzites in the area of the Princess Anne Glacier in the Queen Elizabeth Range also contain first cycle zircon populations with peaks between 1700 and 1800 Ma and lesser peaks from 1400 to 1450 Ma (Goodge et al., 2004), consistent with similar sources to the Tasmanian quartzites (Black et al., 2004). Although Goodge et al. (2004)

did not record paleocurrent directions from the Antarctic quartzites, other units indicated sources from under the icecap, and they considered that the quartzites had come from the same direction.

Goodge et al. (2008) recorded the presence of a 1441±6 Ma (U-Pb zircon SHRIMP date) A-type granite clast in a moraine a few tens of kilometres north of the quartzites and indicated that it came from under the icecap. This clast was correlated with the Granite-Rhyolite Province in Laurentia, suggesting southwestern Laurentia abutted the central Transantarctic Mountains during Rodinia assembly. Similar granites may have sourced both the 1380 to 1450 Ma detrital zircon and the 1370 Ma detrital monazite in the Tasmanian quartzites. Goodge et al. (2010) strengthened this Laurentia-East Antarctica correlation by finding meta-igneous clasts in moraines with protolith ages of 1000 to 1100 Ma. These ages are older than the 1290 Ma metamorphic ages seen on King Island (Berry et al., 2005), but are consistent with the 1000 to 1250 Ma detrital zircons seen in the Rocky Cape and Sorell–Badger Head Zones (Black et al., 2004). They also accord with the 1040 to 1120 Ma ages of dredged granite and orthogneiss from the Glomar Zone (Fioretti et al., 2005a; Berry et al., 2008).

Borg and DePaolo (1991) and Borg and DePaolo (1994) used Sm-Nd isotopic data to propose a strip of Nimrod Group basement that extended outboard of southern Victoria Land through to the Central Transantarctic Mountains and as far south as the Beardmore Glacier. This outboard terrane is characterised by T_{DM} model ages of 1600 to 1900 Ma (ε_{Nd} -15.4 to -10.0 at 500 Ma), which are significantly younger than the >2000 Ma model ages for the inboard Antarctic terranes. Recent data supports this division (Peucat et al., 2002; Goodge et al., 2012). Circa 1600 Ma granites are present in the Mawson Craton under the Antarctic ice (Peucat et al., 2002; Betts et al., 2008), but it is unlikely that the middle and lower VanDieland crust can be correlated with the deep Antarctic crust, because the Antarctic model ages are older than 2000 Ma where these intrusions are present.

Other correlations can also be made. Goodge et al. (2002) recorded the presence of minor ca 765 and 780 Ma detrital zircon populations in the central Transantarctic Mountains, which may be derived from the same igneous event as ~770 Ma rift-related granites in VanDieland. Rhyolitic clasts in the Skelton Group in southern Victoria Land yielded a

650 Ma age (Cooper et al., 2011), within error limits of a 660 Ma deformed granite near Badger Head in northern Tasmania (Black, 2007, unpublished data). A similarly aged (668±1 Ma, U-Pb zircon) gabbro was intruded into the middle of the Beardmore Group

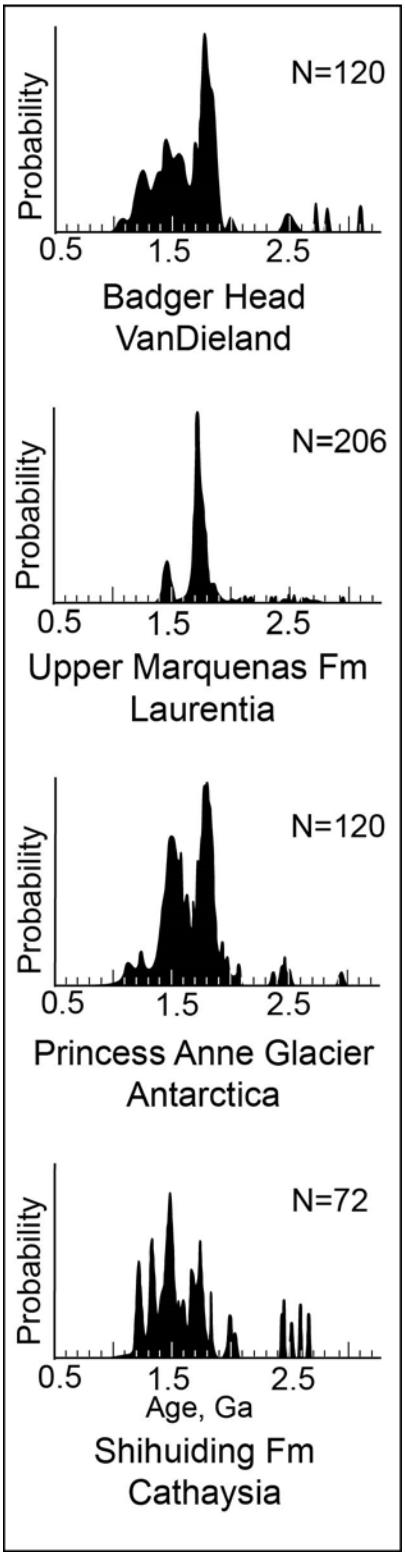


Figure 3-20. Detrital zircons from possible correlatives and/or source areas to VanDieland. Data sources; Badger Head (Black et al. 2004), Marquenas Formation (approximately 1450 Ma, Jones et al. 2011), Princess Anne Glacier (Goodge et al. 2004) and Shihuiding Formation (Li et al., 2008a).

(Goodge et al., 2002), implying that rift-related sedimentation must have commenced earlier. The upper age of the Beardmore Group is poorly constrained, but an unconformably overlying package contained detrital zircons with ages as old as 589±6 Ma, which Goodge et al. (2012) interpreted as indicating that arc volcanism had commenced by then. However, the age is indistinguishable from that of 582±4 Ma zircons in a rift-related rhyolite within a basaltic sequence in the upper Togari Group in Tasmania (Calver et al., 2004), suggesting that similar rift-related rhyolites may have been the source to the 589±6 Ma zircons. Lastly, it may be no coincidence that both the early Beardmore Group and the possibly coeval Oonah Formation in Tasmania are quartz rich at the base, but become more carbonate rich near the top (Brown, 1986; Goodge et al., 2002).

Cathaysia

Li et al. (2008b) and Li and Evans (2011) suggested that the Cathaysia Block lay between Laurentia and the Northern Australian Craton, which could suggest that VanDieland once lay adjacent to Cathaysia. This correlation is strengthened by the presence of similar zircon populations in the Sorell–Badger Head Zone to those from the Shihuiding Formation on Hainan Island at the southern end of the Cathaysia Block (Li et al., 2008a) (Figure 3-20); both regions have detrital zircons with age peaks at approximately 1000 Ma, and 820 to 720 Ma igneous zircons present in the southeastern-most Cathaysia Block cover the age of the 770 to 740 Ma Wickham 'Orogeny' (Black et al., 1997; Black et al., 2004; Li et al., 2014).

Zheng et al. (2011) considered that most, if not all, of the Cathaysia Block was underlain by Archean crust. However, Li et al. (2014) suggested that these Archean ages were derived from far-travelled zircons, and that a more appropriate basement age is approximately 1850 Ma or younger. Regardless of which is correct, neither age supports a correlation with VanDieland, where the data suggests a T_{DM} model age of approximately 1700 Ma or younger. Furthermore, Cawood et al. (2013) considered that the presence of 1000 to 900 Ma arc and back-arc rocks along the western margin of the Cathaysia Block indicated it must have been located on the edge of Rodinia, perhaps off the Western Australian Craton, while Li et al. (2014) considered that these ages reflected plume-related breakup. Both the nature of the southernmost Cathaysia Block on Hainan

Island and whether these volcanic rocks are arc-related or break-up plume-related are beyond the scope of this paper. Thus, VanDieland and southeastern-most Cathaysia might be able to be correlated, but there are too many uncertainties to be more definite.

Baltica

In his ANACONDA model of Rodinia, Evans (2009) placed Baltica between Laurentia and the Southern Australia-Mawsonland craton. Zircons of the appropriate 1380 to 1450 Ma age were produced in the Danopolonian Orogeny of southwestern Baltica, suggesting that a correlation with VanDieland might be possible. As well, widespread remnants of older quartzites are present throughout much of Baltica (Bogdanova et al., 2008), suggesting a possible source for the quartzites in VanDieland. However, other factors seem incompatible with this interpretation. Bogdanova et al. (2008) described a semi-continuous sequence of zircon-producing igneous and metamorphic events that lasted from 1730 to 1480 Ma. This contrasts with the detrital zircon patterns in VanDieland, which typically show a population gap between approximately 1450 and 1650 Ma (Black et al., 2004), and seems less likely to provide the contrast in inheritance between the Paleozoic granites and the Proterozoic metasediments (Section 2.1.2) (Black et al., 2010). Finally, we note the slightly different breakup ages between Fennoscandia, 800 Ma to 550 Ma (Andréasson et al., 1998), and VanDieland, 770 Ma to 575 Ma. Whilst not of themselves critical differences, they too suggest that other hypotheses may be stronger.

3.4 SYNTHESIS, CORRELATIONS AND CONSTRAINTS; A REGIONAL TECTONIC HISTORY

This section connects the previous interpretations and links them with others along the Australian and Antarctic Rodinian to Gondwanan margin.

The xenocrystic zircons in the Paleozoic granites in western Tasmania indicate the presence of an older, unexposed substrate with a strong 1600 to 1650 Ma igneous component. We suggest, first, that the source of these zircons is the Mazatzal Orogen of the Laurentian craton (Section 3.2.2), which Goodge et al. (2010) indicated was adjacent to East Antarctica from at least 1400 Ma to approximately 800 Ma, and second, that VanDieland lay between the two (Figures 3-21 and 3-22). An implication is that we favour a broadly SWEAT-like reconstruction of Rodinia for Australia, Antarctica and Laurentia similar to that suggested by Goodge et al. (2010), Cawood et al. (2013) or Li et al. (2013).

The oldest rocks outcropping in the Tasmanian region are at the base of the Rocky Cape Group and on King Island, and were metamorphosed at ca 1300 Ma (Berry et al., 2005; Halpin et al., 2014). If a Rodinian fit close to those suggested by Goodge et al. (2010) or Li and Evans (2011) is valid, then it appears likely that the 1300 Ma metamorphism is connected to the Grenville events in southwestern USA or East Antarctica. We also suggest that the Grenville Front can be mapped at the northern edge of the Glomar Zone (Figure 3-18), since Grenville-age rocks have been recorded on the Glomar Zone but not in Tasmania.

As Rodinia began to break up, at approximately 830 Ma (Wingate et al., 1998), extension took place along the proto-Terra Australis margin. In VanDieland, it was marked by the 780 to 750 Ma Wickham 'Orogeny' and sedimentation beginning in the Smithton Basin, coeval with the 777±7 Ma bimodal Boucaut Volcanics in the Adelaide Geosyncline (C.M. Fanning personal communication, 1994, in Preiss, 2000). Continued extension is indicated by the intrusion of ca ?710 Ma rift tholeiites in the Burnie Formation and also by the beginning of sedimentation in the Beardmore Group in the Transantarctic Mountains, which pre-dates 668±1 Ma pillow basalts and gabbro there (Goodge et al., 2002; Goodge et al., 2004). Sedimentation also commenced at about 770 Ma in southwestern North America (Timmons et al., 2001; Yonkee et al., 2014). The 660 Ma intrusive age of the deformed granite in the Sorell–Badger Head Zone is also close to the age of the gabbro in the Beardmore Group, suggesting that they may have formed during the same extensional event. We interpret this rifting to have at least partly separated the Pedder, Tyennan, and Sorell–Badger Head zones into extensional allochthons or microcontinental ribbons (Figure 3-21).

After approximately 650 Ma, strain became more distributed, allowing continued deposition in the Smithton Basin and equivalents, but perhaps also the Success Creek Group, which unconformably overlies the Oonah Formation (Brown, 1986). Rifting also took place elsewhere along the Rodinian margin. In western Victoria, ?Neoproterozoic metasedimentary rocks were intruded by gabbro dykes that yielded primary zircons with an age of 643±4 Ma (Morand and Fanning, 2006). In the Transantarctic Mountains, deposition continued in the Beardmore Group (Goodge et al., 2004). A seaway had developed along the entire western USA (Yonkee et al., 2014). The final rifting event in VanDieland is marked by the 575±3 Ma MORB tholeiitic basalts of the Grassy Group on King Island (Calver et al., 2004; Meffre et al., 2004) and the 582±4 Ma rift tholeiitic basalt of the Spinks Creek Volcanics in the Smithton Basin (Calver et al., 2004). The Crimson Creek Formation of the Burnie Zone and other mafic volcaniclastic packages in western Tasmania were probably deposited at this time or shortly before (Seymour and Calver, 1998). These events left the King Island Zone isolated from both East Gondwana and VanDieland, while the Rocky Cape and Burnie zones remained contiguous. The Tyennan and Pedder zones were separated at their (now) northern ends, but connected by subcontinental lithosphere further south, similar to the present western North Atlantic Ocean, where the Hatton, Rockall and Porcupine banks are partly separated by V-shaped basins (Péron-Pinvidic and Manatschal, 2010). The Tyennan and Sorell–Badger Head zones were also contiguous. This event slightly post-dates the last rifting event in the western Lachlan Fold Belt (586±3 Ma, Greenfield et al., 2011), which is evidenced by ca 580 Ma detrital zircon population in the Delamerian Orogen (Ireland et al., 1998; Fanning and Morand, 2002; Morand and Fanning, 2009). In East Antarctica, 589±6 Ma zircon cores surrounded by 559±6 Ma rims in a clast of muscovite-biotite granite may record the same event (Goodge et al., 2012). The rims and the younger of the locally derived 580 to 540 Ma detrital zircon populations seen in the Cambrian Byrd Group in the Transantarctic Mountains (Goodge et al., 2002) may mark the transition to a convergent margin. Breakup was also completed in Laurentia (Yonkee et al., 2014), suggesting that VanDieland was left as an isolated micro-continental ribbon between the larger Gondwanan and Laurentian cratons.

After this, VanDieland drifted as at least two, and perhaps as many as seven fragments, on the proto-Pacific Ocean. In Tasmania, Late Neoproterozoic sedimentation was restricted to carbonate deposition. Early Cambrian laminated siliceous siltstone and shale deposits suggest VanDieland was isolated from clastic sources (Burns, 1964; Everard et al., 2007). The rifted fragments of VanDieland migrated northwards along the earliest Gondwanan margin and were not affected by the ca 560-530 Ma early phases of the Ross

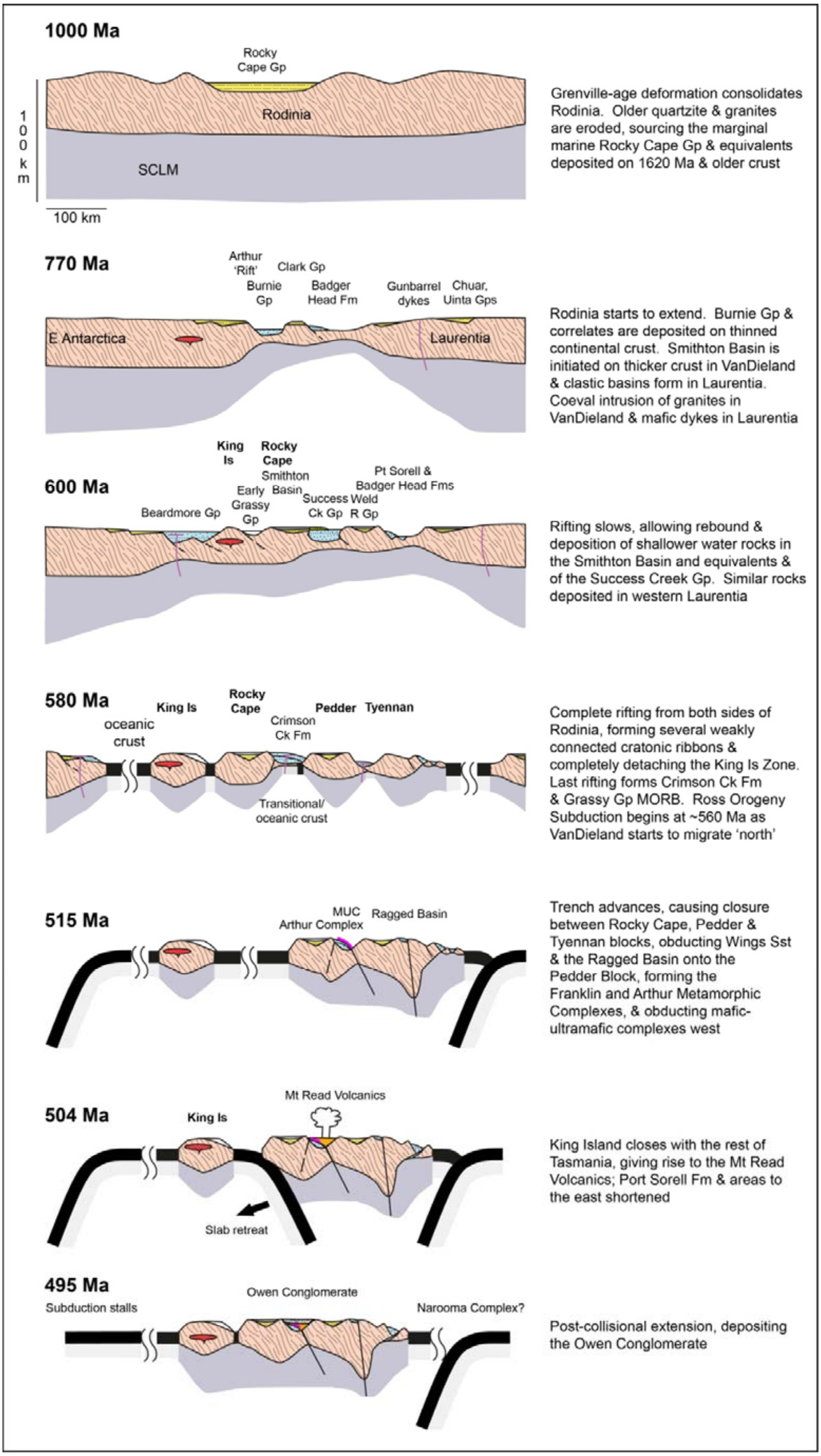


Figure 3-21. Geological history of VanDieland.

Orogeny (Allibone and Wysoczanski, 2002; Stump et al., 2003; Goodge et al., 2012). Following a period of relative tectonic quiescence, the separated fragments of VanDieland began to converge at ca 530 Ma (Figures 3-21 and 3-23), perhaps beginning to generate the high-pressure Franklin Metamorphic Complex at 529±10 Ma, before they underwent decompression at 512±4 Ma (Chmielowski, 2009; Chmielowski and Berry, 2012). Convergence between the Pedder and Tyennan zones occurred along the Mt Hobhouse Fault and may have coincided with the westward obduction of the mafic-ultramafic complexes in the region at 516±1 Ma (Mortensen et al., 2015). Final docking of these zones post-dated movement along the Mt Hobhouse Fault because the eastern edge of the Burnie Zone truncates the fault. During this event, the Wings Subzone, west of the Mt Hobhouse Fault (Figure 3-14) was trapped and obducted westwards onto the Pedder Zone. At a regional scale, these events are coeval with the 514±4 Ma U/Pb zircon age for the syn-tectonic Rathjen Gneiss in South Australia (Foden et al., 1999). Deformation associated with the Ross Orogeny also continued in Antarctica (Goodge et al., 1993), and was associated with felsic magmatism (Encarnación and Grunow, 1996), associated with west-dipping subduction in the region (Boger and Miller, 2004).

By 510 Ma, collisional events were close to peaking in Tasmania. They drove the Tyennan metamorphism and deformation, including in the basement in the Sorell–Badger Head Zone in the east (monazite, 513±8 Ma, Chmielowski, 2009), in the Forth Metamorphics in the central north of the Tyennan Zone (zircon, 512±5 Ma Black et al., 1997 recalculated by Foster et al., 2005; monazite, 510±11, Chmielowski, 2009), and in the Pedder Zone in the southwest (e.g. monazite, 511±5 Ma, Chmielowski, 2009). Inversion and cooling started along the Mt Hobhouse Fault (monazite, 512±4 Ma, Chmielowski and Berry, 2012). Holm and Berry (2002) described north-south shortening along the Arthur Metamorphic Complex in the west. In the east, northeast-southwest shortening took place, thrusting the Badger Head Block and a west-dipping mafic-ultramafic slice over the Port Sorell Formation (D2 in Reed et al., 2002). If the two shortening events were coeval, they would be likely to have driven some mafic-ultramafic complexes south, as described by Berry and Bull (2012). More regionally, Preiss (2001) interpreted an initial northwest-southeast plate convergence at about 510 Ma. This was followed by northward propagating transpression. In the Koonenberry Block in western New South Wales, northeast-southwest extension was accompanied by basaltic to rhyolitic volcanism (Greenfield et al., 2011). Deformation and magmatic activity also continued along the East Gondwanan Antarctic margin (Cawood and Buchan, 2007 and references therein).

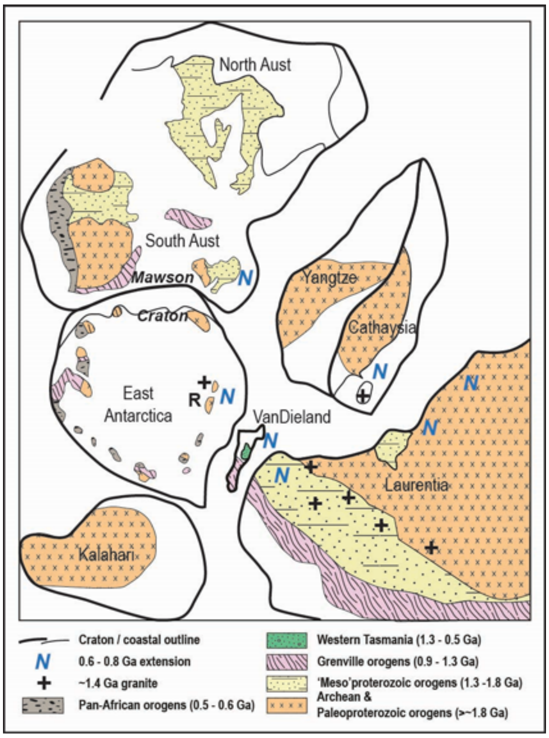


Figure 3-22. Likely position of VanDieland at approximately 780 Ma. Rodinian fit after Li et al. (2013), geology after Goodge et al. (2010). R marks the location of the Miller Range and the Princess Anne Glacier.

The Mount Read Volcanics began erupting at 506.8±1.0 Ma (McNeill et al., 2012; Mortensen et al., 2015). We attribute this to subduction associated with the arrival and collision of the King Island Zone. We note that probable blueschist metamorphism in southwestern Tasmania continued until 505±2 Ma and the white-schist metamorphism along the Mt Hobhouse Fault at 504±5 Ma (U/Th/Pb, monazite, Chmielowski and Berry, 2012), suggesting shortening was maintained until this time. The Mount Read Volcanics continued to erupt until 496±0.9 Ma (Mortensen et al., 2015), coeval with the last thermal event in north central Tasmania at 497±3 Ma (U/Th/Pb, monazite, Chmielowski, 2009). We suggest that the volcanism and shortening were due to an east-dipping subduction zone between King Island and Tasmania that rolled back after about 505 Ma.

The Mount Read Volcanics are contemporaneous with the geochemically similar Stavely Volcanic Belt in western Victoria (Crawford et al., 2003a). Volcanic rocks from the Stavely Volcanic Belt yield U-Pb SHRIMP dates of 501±9 Ma and 495±5 Ma (Stuart-Smith and Black (1999a), similar to Delamerian magmatism in South Australia (Foden et al., 2006) and western New South Wales (Greenfield et al., 2011). This magmatism occurred in the overriding plate of a west-dipping subduction zone along the main East

Gondwana margin, inboard of Tasmania (Foden et al., 2006; Greenfield et al., 2011). The cessation of Tasmanian volcanism is coincident with movement on the Moyston Fault in western Victoria at about 495 Ma ($^{40}Ar/^{39}Ar$ on mica and hornblende, Miller et al., 2005). This fault has a major dip-slip component, with the hangingwall of amphibolite facies rocks and footwall of prehnite-pumpellyite facies rocks (Cayley and Taylor, 2001) and is interpreted to represent the relicts of a west-dipping subduction zone (Miller et al., 2005). However, the fault also has a significant component of sinistral strike-slip movement, suggesting that oblique convergence in the region would have been likely (Cayley and Taylor, 2001). The movement sense is consistent with the regional kinematics in the back-arc region in South Australia (Preiss, 2001), suggesting northwest oblique convergence outboard of the Cambrian East Gondwana margin continued for at least 15 m.y. It is also coeval with subduction associated with the Narooma Complex on the south coast of New South Wales (Prendergast and Offler, 2012). In the same period, our model suggests that the individual tectonic elements that comprise VanDieland were accreting to the west of the Narooma Complex (Moore et al., 2012a) but outboard of the Gondwanan margin.

After these accretion events, at the end of the Cambrian and into the Early Ordovician, western Tasmania underwent extension, perhaps caused by further rollback and/or by retreat of the eastern subduction zone. As a result, the locally-derived Owen Conglomerate was deposited in half-graben on the west coast and in the north as far east as the eastern part of the Sorell–Badger Head Zone (Noll and Hall, 2005; Reed and Vicary, 2005). Similar aged, locally derived conglomerate is also present in western New South Wales (Greenfield et al., 2011) and probably also in north Victoria Land (Tessensohn and Henjes-Kunst, 2005). Near VanDieland, subduction stalled, perhaps as a result of the collision of the Dimboola Complex into the early Gondwanan margin (Moore, 2006). In Antarctica, granite intrusion and deformation continued into the Ordovician, until at least 480 Ma (Goodge et al., 2004; Rossetti et al., 2011). At least one granite, the 512±3 Ma Surgeon Island Granite, shows strong affinity in its inherited zircon population with granites in western Tasmania (Fioretti et al., 2005b). As well, the eastern Robertson Bay Terrane appears to be underlain by a different basement from the rest of north Victoria Land (Fioretti et al., 2005b). This suggests that VanDieland and the eastern Robertson Bay Terrane, separated in the Gondwanan breakup, were once part of the same micro-continental fragment. Thus the terrestrial source of the Cambrian sedimentary rocks of the Robertson Bay Terrane (Henjes-Kunst and Schüssler, 2003; Tessensohn and Henjes-Kunst, 2005) may have been the erosional products of the Tyennan Orogeny in VanDieland.

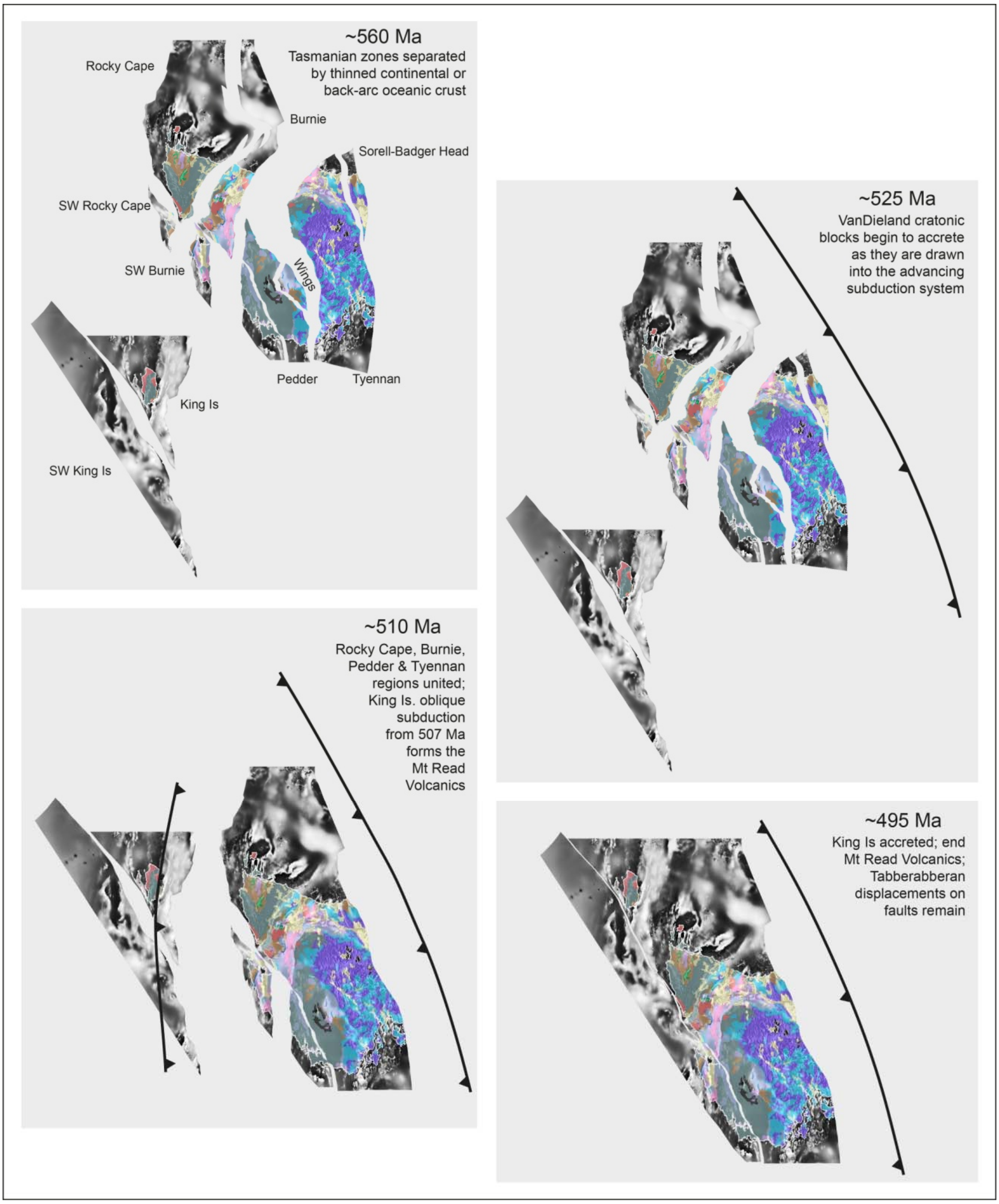


Figure 3-23. Reconstruction of western Tasmanian assembly. Compare with Figures 3-1 and 3-2. At 560 Ma, VanDieland was a series of cratonic blocks, either loosely connected by transitional crust, or completely separated as in the case of the King Island Zone. These began to accrete at about 525 Ma (the Tyennan Orogeny) and by 510 Ma or shortly thereafter, most of the cratonic blocks had come together. By 495 Ma, the remaining blocks had been accreted and most of the movement along the Braddon River Fault had taken place.

At this time, the only paleopole available from northwestern Tasmania places it close to the East Gondwanan coast, if not abutting it (Li et al., 1997). Evidence from Victoria suggests that the former is more likely. During the Late Ordovician to Silurian Benambran Orogeny, the Bendigo Zone was shortened by ~200 km (Gray et al., 2006; Cayley et al., 2011), while the western Melbourne Zone was shortened by ~40 km in the Middle Devonian Tabberabberan Orogeny (Foster and Gray, 2007), which was accommodated by the Bendigo Zone over-riding the Selwyn Block (Cayley et al., 2011). This shortening is well within the most likely error limits given by Li et al. (1997), α_{95} = 10.4° (i.e. ~700 km east-west and ~1200 km north-south).

Our accretion model for VanDieland requires broadly similar plate kinematics to that defined on the Late Cambrian eastern margin of Gondwana (e.g. Cayley and Taylor, 2001; Cayley, 2011). The horizontal accretion of micro-continental slivers or ribbons took place when a central core comprising the Tyennan and Pedder zones successively accreted or obducted the Burnie-Arthur-Rocky Cape, Sorell–Badger Head and King Island zones during north- to northwest relative movement (Figure 3-22). During this accretion, north-south shortening took place in the Trial Harbour area on the west coast (McFarlane, 2011), in central Tasmania (Calver et al., 2006) and the Arthur Lineament (Holm and Berry, 2002). Our model suggests that the internal collisions were oblique, consistent with the kinematic observations. While the magnetic and gravity interpretation provides a clear sequence of accretion events, the polarities of the subduction zones between the accreted fragments are largely unconstrained. Exceptions are the collision between the Tyennan-Pedder and Burnie regions, where westward obducted mafic-ultramafic complexes indicate east-dipping subduction (Crawford and Berry, 1992), and the King Island-western Tasmanian subduction zone, which is also east-dipping. Cayley (2011) proposed that VanDieland briefly accreted to the eastern margin of Gondwana at about 500 Ma and this is tentatively supported by trilobite fauna (Hally and Paterson, 2014), which shows a convergence of shallow-water species in Gondwana and western Tasmania at that time.

At the global scale, these later events lie between the early Gondwana continent and the great circum-Pacific subduction system that has existed throughout the Phanerozoic (e.g. Coney, 1992; Cawood, 2005). We suggest that at least two Tasmanian fragments broke off in the dispersal of Rodinia at about 580 Ma (Yonkee et al., 2014) and these were left isolated in the newly formed ocean. Whether this was due to rifting or subduction roll back that caused back arc extension in the overriding plate is not clear. After VanDieland had departed, East Antarctica began to accrete adjacent terranes to form Gondwana (Boger, 2011). However, in the region of the Transantarctic Mountains, the Neoproterozoic continental fragments remained stranded in a back-arc setting between a retreating west-dipping subduction zone and the Gondwana margin. VanDieland was left outboard of the subduction system that started in the Antarctic region at about 560 Ma (Goodge et al., 2012) and the larger subduction system that developed in the Cambrian along the early Gondwanan margin (Cawood, 2005; Casquet et al., 2012). During inversion of this back arc setting the micro-continental fragments and megaboudins of VanDieland migrated northwards, and from approximately 520 Ma to approximately 495 Ma they successively accreted within this closing back arc system. Unlike the Beardmore and Bowers Terranes along the Antarctic margin (Stump et al., 2003; Godard and Palmeri, 2013), VanDieland did not accrete back onto the early Gondwanan margin at this time. Rather, it moved closer in the Early Silurian and, together with eastern Tasmania and the Lachlan Orogen, was finally integrated with the Gondwanan craton in the Middle Devonian.

The accretion of different fragments in different ways along the Gondwanan margin in the Ross-Tyennan-Delamerian Orogeny would inevitably lead to different effects at different times along the same margin. The timings range from 554±10 Ma ($^{87}Rb/^{86}Sr$) in South Australia (Turner et al., 2009) and 559±6 Ma (SHRIMP) in Antarctica (Goodge et al., 2012), through to 489±3 Ma ($^{40}Ar/^{39}Ar$) in western Victoria (Miller et al., 2005) and 484±8 Ma (SHRIMP) in Antarctica (Goodge et al., 2012), and perhaps even to approximately 485 Ma (SHRIMP) in north Queensland (Paulick and McPhie, 1999). Some micro-continental ribbons accreted directly onto the early Gondwanan margin, while others aggregated outboard. Our synthesis would suggest that VanDieland represents one of these micro-continental fragments, and was separated from the Western Lachlan Orogen by ocean crust that is now imbricated (Cayley et al., 2011). It suggests that the accretion of VanDieland must have happened after the accretion of the Dimboola Complex in western Victoria. This may have occurred during the early Ordovician but was finalised during early Silurian Benambran and Middle Devonian Tabberabberan orogenies (Glen, 2005; Champion et al., 2009). A modern analogue is the western Pacific

(Crawford et al., 2003b), where some micro-continental ribbons have been accreted directly onto Asia (Metcalfe, 2011), while others, like in The Philippines, have come together but are yet to accrete to the Asian mainland (Yumul et al., 2003).

3.5 FUTURE DIRECTIONS

Many areas remain poorly controlled. At a more detailed scale, mapping of key areas will help to understand the more problematic boundary relationships. Because much of region is difficult to access, mapping and age dating along the western Tasmanian coast is particularly important, and any results should be integrated with the potential field and seismic data. One important area is northwest from Nye Bay, where doubts remain as to the position of the eastern edge of the Burnie Zone and to the relationships of the mafic volcanic sequences there to the rest of the Burnie Zone. Another is along the Mt Hobhouse Fault, where mapping should more accurately locate its position and history.

At the regional scale, there is a need for better paleomagnetic controls on VanDieland, but this can only come after tight age constraints can be placed on individual units. Age controls have been established in some of the glacial sequences of the Smithton Basin and associated rocks (Calver, 2011; Calver et al., 2013a), but older rocks are generally less well constrained, as are supposedly equivalent sequences elsewhere. Tighter age controls have been established in some pre-570 Ma rocks on King Island and a comparison of poles with those in the Rocky Cape Zone would test the hypothesis that the two zones were from different parts of Rodinia. As well, too few Sm-Nd model ages of the basement have been determined to make statistically valid comparisons with other regions. These should be obtained not only from the Paleozoic plutonic rocks, but also from plutonic rocks on the South Tasman Rise. Only then can the hypothesis proposed here be adequately tested.

3.6 ACKNOWLEDGEMENTS

The geophysical data used in this study has come from many years of acquisition and compilation by Geoscience Australia. The regional mapping was from Mineral Resources of Tasmania. Without these data, the study would not have been attempted. Research for this paper has been supported by Monash University as part of a PhD project. Their assistance has been invaluable. We also gratefully thank Clive Calver for his insights gained through many years of mapping much of Tasmania. Without these, the paper would have been much the poorer. Discussions with Nick Direen and Ross Cayley also helped clarify several issues. Reviewers Ross Cayley and Zheng-Xiang Li and an unnamed reviewer further improved this manuscript. This research was funded from ARC Discovery Grant DP11010253.

PART B: Declaration for Thesis

Monash University

Declaration for Thesis Chapter 4

Declaration by candidate David H Moore

In the case of Chapter 4, the nature and extent of my contribution to the work was the following:

Nature of contribution	Extent of contribution (%)
Did the research, wrote almost all of the paper	95%

The following co-authors contributed to the work. If co-authors are students at Monash University, the extent of their contribution in percentage terms must be stated:

Name	Nature of contribution	Extent of contribution (%) for student co-authors only
Peter Betts	Supervision, guidance on text	5%

The undersigned hereby certify that the above declaration correctly reflects the nature and extent of the candidate's and co-authors' contributions to this work*.

Candidate's Signature	[redacted]	28 August 2015
Main Supervisor's Signature	[redacted]	26 August 2015

*Note: Where the responsible author is not the candidate's main supervisor, the main supervisor should consult with the responsible author to agree on the respective contributions of the authors.

Chapter 4

We measure shadows, and we search among ghostly errors of measurement for landmarks that are scarcely more substantial.

E.P. Hubble (in D. Overbye, 1991, Lonely hearts of the cosmos)

The basement under Bass Strait; connecting Paleozoic Victoria and Proterozoic Tasmania

ABSTRACT

The geology of western Tasmania has been interpreted north across Bass Strait into Victoria using airborne magnetic and marine gravity data. The three westernmost zones in Tasmania, the King Island, Rocky Cape and Burnie zones, are inferred to form the Selwyn Block in Victoria. The Eastern Tasmanian Zone correlates with the Tabberabbera Zone in Victoria. These relationships suggest the Tasmanian Tiers Fracture Zone is equivalent to the Victorian Governor Fault. Regions of Victorian geology that are interpreted to correlate with western Tasmania contain rare outcrops of equivalent geology. The Ceres Gabbro in Victoria is correlated with a package of magnetic rocks west of King Island that is tentatively considered to be Neoproterozoic. The Cambrian andesitic to rhyolitic volcanic rocks of the Jamieson, Licola and Glen Creek windows correlate with rocks above the Burnie Zone, as are most of the Mt Read Volcanics in Tasmania. The Burnie Zone equivalents in the Selwyn Block include all known ultramafic rocks, similar to the Burnie Zone in Tasmania. Reinterpretation of the deep seismic reflection line in central Victoria indicates Burnie Zone equivalent rocks were thrust southwest over the Rocky Cape Zone equivalents.

A link between western Tasmania and central Victoria can be interpreted from Upper Devonian rocks in the Selwyn Block. The eastern end of the Upper Devonian Cobaw Complex and the Warburton Granodiorite both contain calc-silicate enclaves that are interpreted to have been derived from a northern equivalent to the Smithton Basin. The Mount Disappointment Granodiorite has a high Ni and Cr content and pseudomorphs after orthopyroxenes, consistent with it having been partly sourced from the underlying basaltic rocks similar to those on the eastern margin of the King Island Zone. Quartzite cobbles in a Devonian conglomerate in the southeastern Melbourne Zone may have been derived from Rocky Cape Group equivalents. In the western Melbourne Zone, both ε_{Nd} values and $^{87}Sr/^{86}Sr$ ratios suggest older crust in the north and younger crust in the south. This is interpreted to correspond to the transition from the Mesoproterozoic to Neoproterozoic Rocky Cape Zone to the Ediacaran part of the King Island Zone. The deep magnetic responses under the Strathbogie Complex and the Lysterfield

Granodiorite are attributed to metamorphism of part of an extension of the Smithton Basin, most likely an equivalent to the Spinks Creek Volcanics.

4.1 INTRODUCTION

The landward margin of the Terra Australis Orogen was formed by the breakup of Rodinia (Cawood, 2005). In the eastern Australian-Antarctic sector, the last breakup events took place at approximately 580 Ma (e.g. Greenfield et al., 2011) (Chapter 3). Thereafter the Pacific Ocean was kept open by a stable convective cell, upwelling along a mid-ocean ridge system and subducting in a more complex fashion around the Pacific margin (Coney, 1992; Collins, 2003). This system provided the opportunity for the accretion of exotic terranes into the Terra Australis Orogen. For much of the Early Paleozoic in eastern Australia, the orogen was generally in extension but this was punctuated by shorter compressional phases (Collins, 2002) that coincided with the arrival of exotic terranes (Cayley et al., 2002; Cayley, 2011; Moresi et al., 2014). These exotic (or suspect) terranes are important in understanding the broader orogen (Coney, 1989), since:

- the timing of their arrivals may mark the time of switching from extension to compression in the orogen,
- their internal structure and stratigraphy may have been quite different to the enclosing orogen. This means that the exotic terrane may have had very different chemical and/or physical properties to the enclosing orogen, which in turn could have led to the generation of mineral systems or igneous rocks with different characteristics to those nearby, and
- the collision of an exotic terrane could have profoundly influenced the development of the enclosing orogen (Moresi et al., 2014).

Exotic terranes appear to have played a significant role in modifying the margin of East Gondwana (e.g. Moresi et al., 2014). Further, the presence of these exotic terranes along this margin resolves many outstanding issues, such as not being able to correlate the geology of the Terra Australis Orogen along strike. One of the most perplexing issues is the link between the geology of western Tasmania and the southern Lachlan Orogen in Victoria, which are separated by the relatively narrow Bass Strait.

Although many studies have examined the sedimentary and igneous rocks deposited outboard of the southern Australian sector of the Terra Australis Orogen, until recently the likelihood of pre-existing cratonic fragments having been incorporated into the Australian margin has generally been given less prominence (e.g. Spaggiari et al., 2003b; Glen, 2005). Little changed for about 50 years after David (1950, p. 135) wrote that "the tectonic history of this region is hard to decipher" when discussing differences in the geological histories of Tasmania and mainland southeastern Australia.

The Selwyn Block model opened up the possibility that these problems might be solved. This model (Cayley et al., 2002; Cayley et al., 2011) (Chapter 2, Section 2.3) hypothesised that the Proterozoic western Tasmanian crust extended under Bass Strait and into central Victoria and lies beneath the Melbourne Zone. The Selwyn Block model explained:

- the presence of a carbonate-rich Lower Paleozoic outcrop at Cape Liptrap (Figure 4-1) on the Victorian coast that has more in common with those in Tasmania than the coeval turbidite-rich rocks in Victoria,
- the meta-igneous rocks unconformably below the carbonates mentioned above, and in the Barrabool Hills west of Geelong and on Phillip Island,
- the continuity of magnetic responses from western Tasmania across Bass Strait and into Victoria, suggesting geological continuity,
- the Cambrian felsic to intermediate volcanic rocks in the eastern Melbourne Zone, a region that elsewhere is dominated by Ordovician black shale and Silurian to Middle Devonian turbidite rocks,
- the differences in structural grain between the southern and northern parts of the Melbourne Zone in central Victoria,
- the change in seismic character under the Melbourne Zone when compared to the adjacent Bendigo Zone, and
- the increase in earthquake intensity in central Victoria (Gray et al., 1998), consistent with significant lithological variation at depth.

Where the correlative rocks are exposed in western Tasmania, they have been subdivided into 7 zones (Figures 3-1 and 3-3, Chapter 3). From the west these are

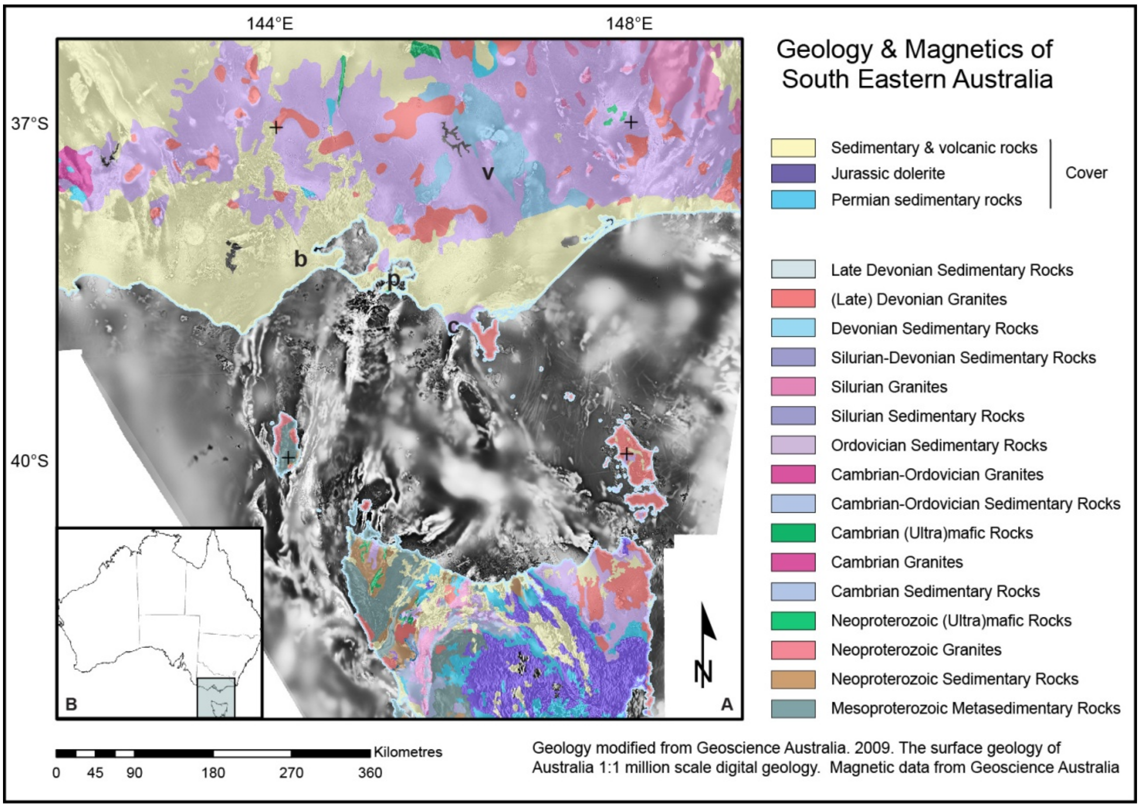


Figure 4-1. A. Geology of southeastern Australia with an intensity layer the total magnetic intensity of the region. b marks the location of the Barrabool Hills, c, Cape Liptrap, p, Phillip Island and v, the Cambrian volcanic rocks in central Victoria. B Location map.

- the King Island Zone, of deep water sedimentary rocks that were metamorphosed at 1290 Ma and again at 760 Ma (Turner et al., 1998; Berry et al., 2005) and that is bounded to the east by 580 Ma rift tholeiites (Meffre et al., 2004),
- the Rocky Cape Zone, a three-layer package, with the upper layer the Ediacaran-Cryogenian Smithton Basin containing volcaniclastic rocks, siltstone, mudstone, dolomite, 580 Ma rift tholeiite (Calver et al., 2004), diamictite, chert and conglomerate, overlying Mesoproterozoic clean marginal marine quartzite and siltstone of the Rocky Cape Group (Everard et al., 2007; Halpin et al., 2014), which in turn overlies an unseen basement containing rocks with a significant component of 1600 to 1650 Ma zircons and with a T_{DM} model age of approximately 1700 Ma (Black et al., 2010),
- the Burnie Zone, largely of Neoproterozoic deep water turbidite rocks intruded by 710 Ma basaltic rift tholeiite sills (McDougall and Leggo, 1965; Crook, 1979; Black et al., 2004) that are overlain by ?580 Ma

mafic volcaniclastic sandstone, shale and rift-related basalt, and Cambrian mafic-ultramafic rocks (Vicary, 2004) and that is often overlain by the Upper Cambrian Mt Read Volcanics (Mortensen et al., 2015),

- the Pedder Zone and Tyennan Zone, both of which comprise similar geology to the Rocky Cape Zone, but are separated by eclogite facies metamorphic rocks (Chmielowski and Berry, 2012). Both are overlain by minor amounts of Mt Read Volcanics,
- the Sorell-Badger Head Zone, of turbidite rocks similar to those in the Burnie Zone and, in the west, of ?Ediacaran melange, chert, volcaniclastic sandstone, black shale, rift-related dolerite and dolostone (Calver and Reed, 2001), and
- the Glomar Zone, which lies in submarine plateaux to the south of the above zones, and which includes granites with ages of approximately 1050 Ma and 1120 Ma and that was metamorphosed at approximately 920 Ma (Fioretti et al., 2005a; Berry et al., 2008).

Cayley (2011) suggested uniting these western Tasmanian rocks with the Selwyn Block under the single term, VanDieland. These Proterozoic and Cambrian rocks contrast with the enclosing part of the Terra Australis Orogen, locally known as the Lachlan Orogen. The Lachlan Orogen in Victoria and eastern Tasmania is comprised of Cambrian to Middle Devonian turbidite rocks that have generally been metamorphosed to epizone or anchizone grades (Patison et al., 2001; Wilson et al., 2009) although, to the east of the region under detailed consideration, the rocks have been metamorphosed to migmatites in the ~430 Ma Benambran Orogeny (VandenBerg et al., 2000). Where the Selwyn Block is absent, the turbidite rocks have a substrate of Cambrian ocean floor basalt (Cayley et al., 2011). The Selwyn Block is generally covered by up to 15 km of Ordovician black shale and Silurian to Middle Devonian turbidite rocks (VandenBerg et al., 2000).

The collision of VanDieland with western Victoria provides a potential mechanism for the semi-continuous deformation from about 460 Ma to 370 Ma (Gray et al., 2003; Cayley et al., 2011; Moresi et al., 2014). Deep seismic data image a distinct change in the middle crust under the Melbourne Zone (Cayley et al., 2011) and other independent tests since the model was first suggested by VandenBerg et al. (2000) have continued to support its existence (Clemens et al., 2014; Pilia et al., 2015).

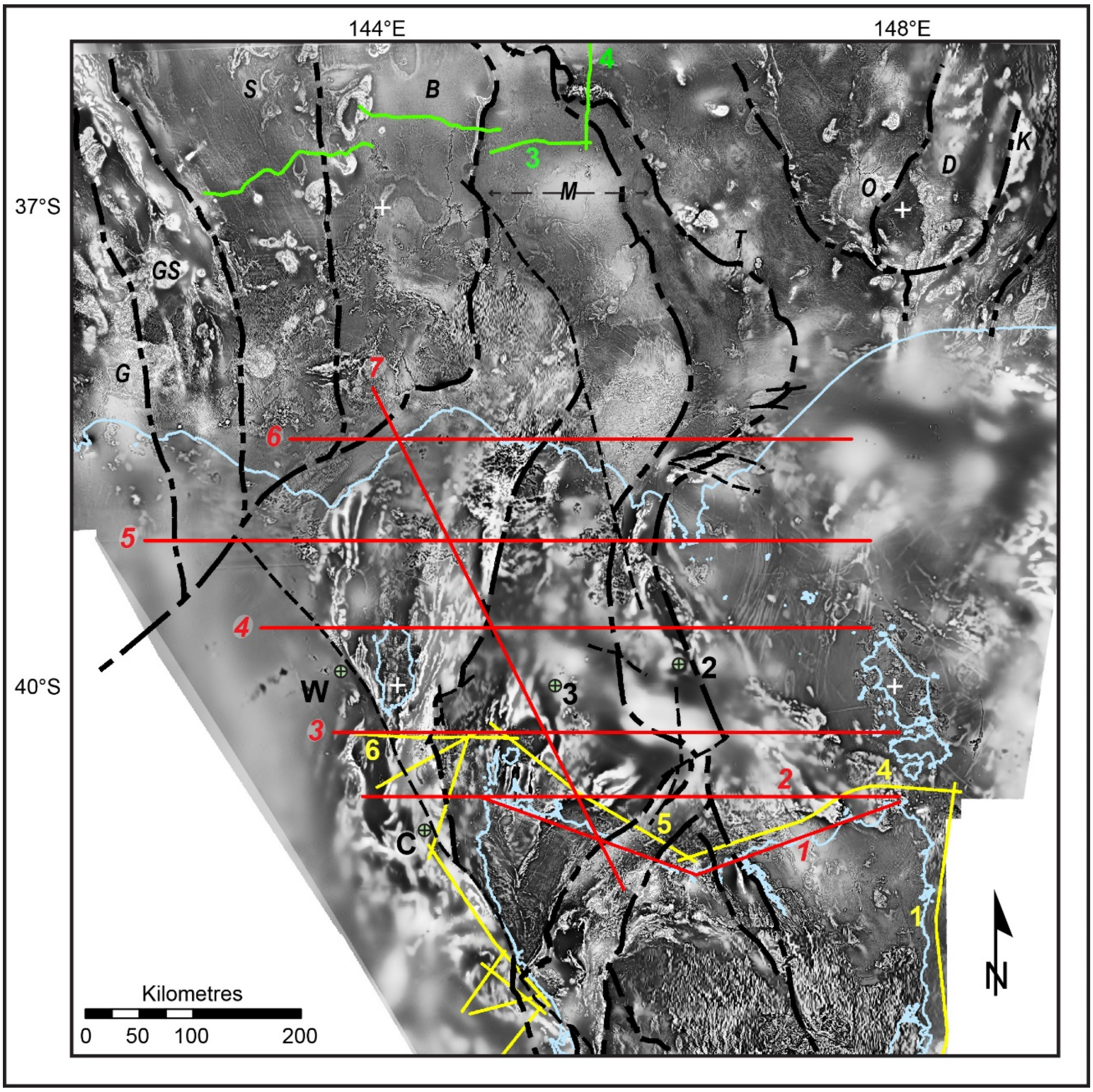


Figure 4-2. Total magnetic intensity, showing the lines modelled (red), the offshore TASGO seismic survey 148 (yellow, with line numbers referred to in the text), the Victorian 2006 seismic survey (green) and offshore boreholes that intersected the Proterozoic basement. Victorian seismic lines 3 and 4 are interpreted in Figure 4-17. Boreholes are 2, Bass 2, 3, Bass 3, C, Clam 1, and W, Whelk 1. Thicker black dashed lines mark the (near-)surface zone boundaries, thinner black dashed lines indicate faults. Red numbers refer to the line numbers in the text. Victorian zones *G* Glenelg, *GS* Grampians-Stavely, *S* Stawell, *B* Bendigo, *M* Melbourne, *T* Tabberabbera, *O* Omeo, *D* Deddick, *K* Kuark. Same area as Figure 4-1. Magnetic data from Geoscience Australia.

Despite these recent gains in understanding the role of VanDieland in eastern Gondwanan accretionary dynamics, the detailed connections between western Tasmanian geology and that of central Victoria remain poorly resolved. For example, McLean et al. (2010) suggested that the large magnetic anomalies present under the Strathbogie Complex in central Victoria and further south, and which lie at depths close to the top of the Selwyn Block, resulted from ultramafic sheets and are equivalent to the ultramafic rocks seen in western Tasmania. However, the largest ultramafic complex in Tasmania has an area less than 250 km^2 and, even assuming continuity between the largest group of occurrences, occupies an area of approximately 1,200 km^2 (Crawford et al., 2014). By contrast, the area of the central Victorian anomalies exceeds 23,000 km^2 (McLean et al., 2010). While some of the size difference can be attributed to post-Middle Devonian erosion in Tasmania, this ten to hundred-fold size difference invites an investigation of possible alternatives.

The deep seismic images of central Victoria show distinct layering in the Selwyn Block. Cayley et al. (2011) attributed the upper layer to Cambrian felsic volcanic rocks and correlated them with those seen in the Jamieson and Licola erosional windows (VandenBerg et al., 2006), but the lower layers remain unexplained. Furthermore, the gravity inversions in Cayley et al. (2011) used a density of 2780 kgm^{-3} for the upper layer, a density more typical of intermediate to mafic rocks. The background density used was 2670 kgm^{-3}.

Better understanding of the relationships between the geology of western Tasmania and the Selwyn Block would help to resolve questions like this and so improve our understanding of the Selwyn Block. As well, it could aid understanding of both the Paleozoic evolution of Tasmania and of the surrounding parts of the Lachlan Orogen, including the origins of the Devonian granites in the region.

The Selwyn Block is overlain by up to 15 km of Silurian and Devonian sedimentary rocks of the Melbourne Zone, with only rare, isolated windows of the basement geology exposed at Ceres (west of Geelong), Phillip Island, at Cape Liptrap and as the Cambrian volcanic rocks in the east of the Melbourne Zone. However, there are many Upper Devonian granites that intrude the Melbourne Zone and these granites could provide insights into the basement rocks because they are derived, at least in part, from basement melts (Chappell and White, 1974; Keay et al., 1997; Rossiter and Gray, 2008; Clemens and Phillips, 2014). The ability to trace Tasmanian geology across Bass Strait into the region where the granites formed could allow for a better understanding of the granites.

Although Chapter 3 outlines a strato-tectonic map of the pre-Ordovician rocks of western Tasmania, no such map exists for the parts of VanDieland under Bass Strait or the Selwyn Block in Victoria as they mostly lie under water or several kilometres of younger rocks. This chapter takes both the geophysical (magnetic, gravity and seismic) and geological (outcrop, drill hole, granite geochemistry and xenolith) data in an attempt to extrapolate these divisions north. Geological cross sections have been constructed using Gm-Sys® forward modelling software (Stewart and Betts, 2010) to further constrain the 3D geometry of the basement rocks beneath the Bass Basin and to provide a better model of VanDieland and adjacent areas. The result is the first strato-tectonic map of the northern part of the micro-continent and a description of some of the relationships between VanDieland and the Lachlan Orogen.

4.2 METHOD

The oldest basement exposed on the north coast of Tasmania and on King Island have been studied in detail and integrated with the marine drilling results, offshore seismic, airborne magnetic and onshore and offshore gravity data (Chapter 3). The major packages and boundaries were then interpreted across Bass Strait and correlated with the geology of Victoria (Figures 4-1 and 4-2). Finally, the geological interpretation was used to construct seven gravity and magnetic constrained geological cross-sections using Gm-Sys® (Stewart and Betts, 2010) to further constrain the correlation across Bass Strait. The sections were constructed to a depth of 30 km. The MOHO depth is constrained by AGSO (Australian Geological Survey Organisation) offshore seismic survey 148, and typically occurs at a depth of 10 to 12s TWT (e.g. Drummond et al., 2000) (Chapter 3), and thus interpreting sections below ~30 km would need to include significant amounts of mantle, a largely unconstrained parameter. Although this approach used the best available data, several problems remained. While the onshore outcropping Pre-Mesozoic geology is well known in both Victoria and Tasmania (e.g. VandenBerg et al., 2000; Corbett et al., 2014), only six offshore bore holes were considered to have intersected Paleozoic or Proterozoic rocks (Table 4-1). The remaining 25 bore holes in the Geoscience Australia database (Blevin et al., 2003) did not intersect basement, and therefore significant parts of the interpretation were less well constrained.

The 2006 Victorian deep seismic survey suggests Cambrian and Ordovician meta-turbidite rocks in western Victoria are underlain by ocean floor basaltic rocks that are largely concealed (Cayley et al., 2011). Away from the margins of the Stawell and Bendigo zones (Figure 4-2), where the metabasalts crop out, the thickness of the meta-turbidite package varies from about 0.6 s TWT (~2 km) to 6 s TWT (~18 km). Below this lies stacked slices of mafic oceanic and metasedimentary rocks (Cayley et al., 2011). A similar distribution is suggested in eastern Victoria (VandenBerg et al., 2000; Moresi et al., 2014). In eastern Victoria, the base of the outcropping sequence occurs at the eastern edge of the Selwyn Block, where Cambrian pillow basalts and boninites have been thrust over Late Cambrian chert and Ordovician meta-turbidites (VandenBerg et al., 2000; Spaggiari et al., 2004; VandenBerg et al., 2006). In Tasmania, Roach and Leaman (1996) used geophysical modelling to interpret the presence of west-directed thrust slices of mafic or ultramafic rocks below parts of the Mathinna Supergroup. Furthermore, Figure 3-17, Chapter 3 indicates west-verging ramp anticlines of mafic rocks in the seismic data on Geoscience Australia line 148/04. There are no outcrops of these rocks and so the interpretation is speculative.

Hole	East; Decimal degrees	South; Decimal degrees	Basement Lithology	Interpreted Unit	Inferred Age
Bass 2	146.304	39.886	Tuffaceous mudstone, mafic volcanics	Crimson Creek Formation	Neoproterozoic K-Ar 589±3 Ma
Bass 3	145.2825	39.9975	Quartzite	Rocky Cape Group	Mesoproterozoic
Clam 1	144.2167	40.8631	Argillite, Basalt	Grassy Group	Ediacaran K-Ar 630±21 Ma
Whelk 1	143.5572	39.8979	Basalt	Grassy Group?	Neoproterozoic?
White Ibis 1	145.254772	39.963781	Quartzite	Rocky Cape Group	Mesoproterozoic

Table 4-1. Offshore drill holes that intersected Proterozoic basement.

With the exception of AGSO survey 148 around Tasmania, almost all of the available seismic data from Bass Strait were collected over basins; hence there are minimal data in areas with little hydrocarbon potential. Thus, regions with the best geological control often had the weakest seismic control. Also, the seismic data were processed to maximise contrast in the basins, and so basement structures were not effectively imaged. These problems are compounded because basement structures are steep. Nevertheless, previous seismic interpretations in the Bass Basin (Setiawan, 2000) and the Torquay Embayment (Madsen, 2002) provided tighter depth constraints on the shapes of the basins along the modelled profiles.

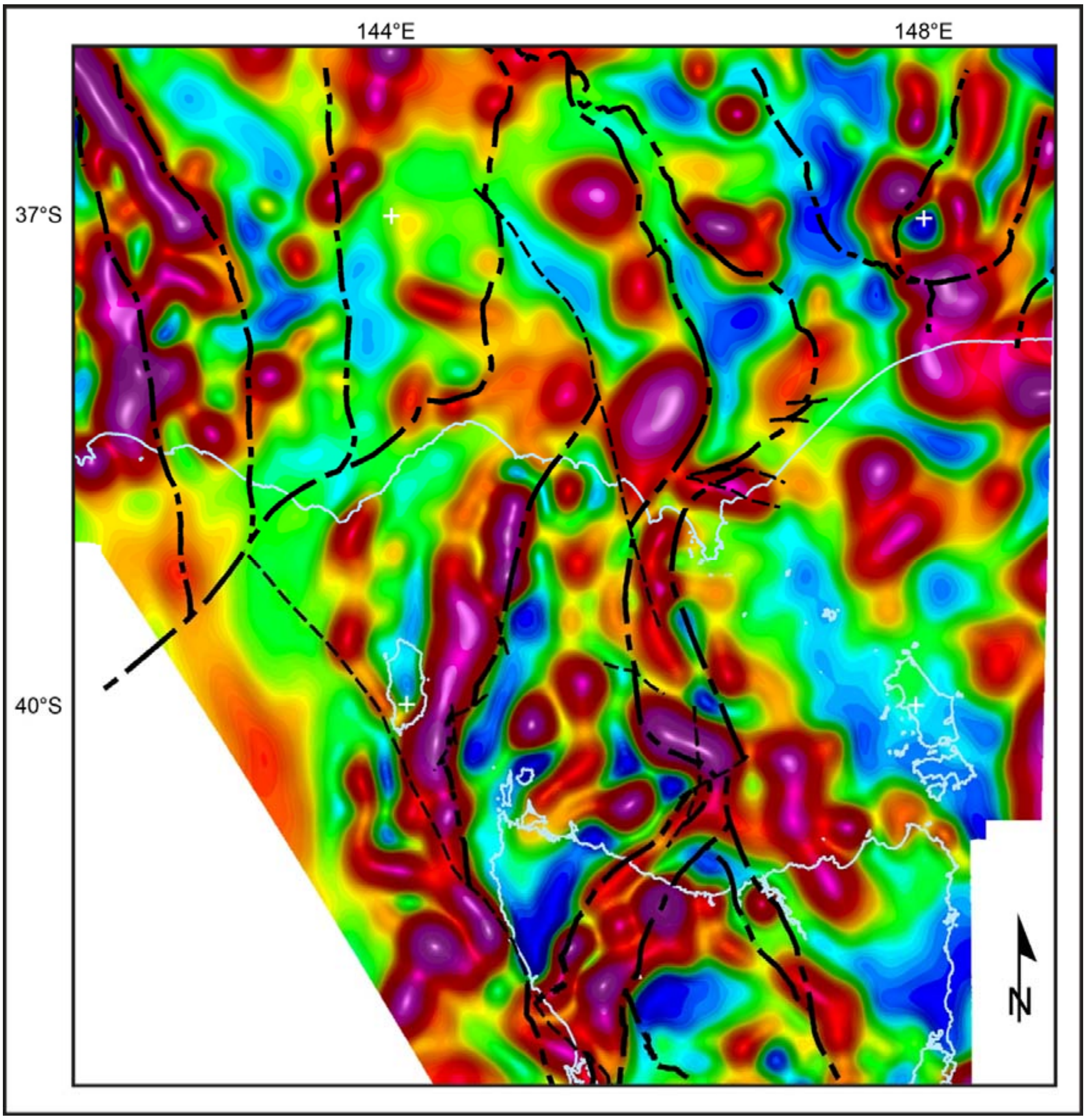


Figure 4-3. Upward continued total magnetic intensity. Thicker black dashed lines mark the (near-)surface zone boundaries, thinner black dashed lines indicate faults. Same area as Figure 4-1. Magnetic data from Geoscience Australia.

The gravity data set used was compiled by Geoscience Australia prior to 2010. Since the marine gravity was almost entirely from hydrocarbon exploration surveys, there were gaps in the data of up to 100 km in areas where there was little or no basin development. Onshore data in northwestern Tasmania also contained gaps of up to 50 km. Consequently, the gridding process generated artefacts. Furthermore, there are no gravity surveys that cross the land-sea interface—most marine data are deliberately collected well away from coastal areas. Hence stitching of onshore and offshore data sets is problematic. Thus, parts of the gravity profiles on Sections 1, 2, 5, 6 and 7 (Figures 4-9, 4-10, 4-14, 4-15 and 4-16) are unreliable in the region where the onshore and offshore data sets were merged.

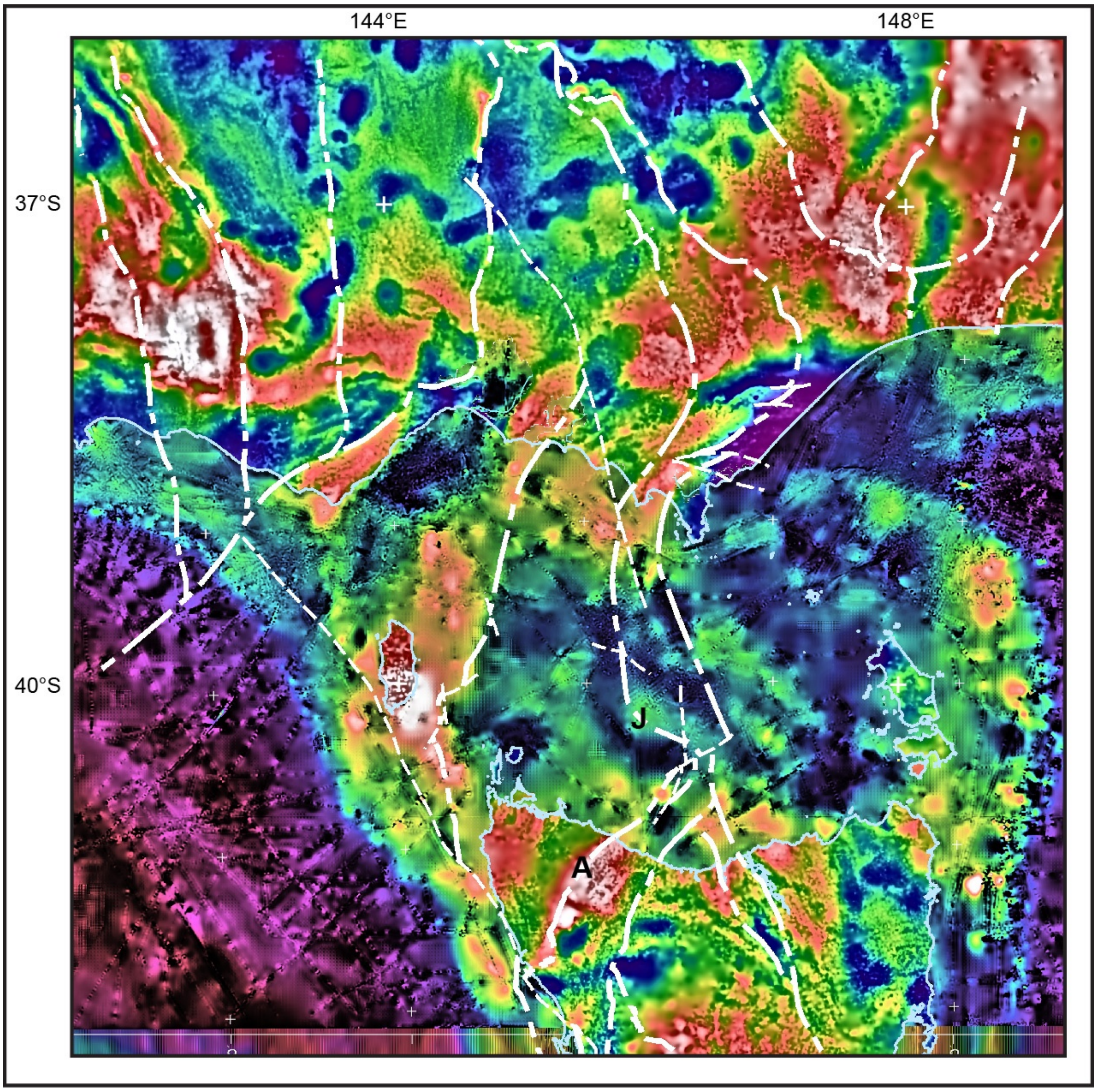


Figure 4-4. Isostatic gravity of the same region as Figure 4-1. Thicker white dashed lines mark (near)-surface Zone boundaries, thinner white dashed lines mark faults. The onshore and offshore isostatic gravity models used slightly different parameters. A marks the location of the Arthur Complex, and J, a gravity high interpreted to be from Jurassic Dolerite. See also Figure 4-8. Gravity data from Geoscience Australia.

Density data in the Mesozoic and Cenozoic sequences were lacking, and this also had a significant effect on the gravity models. Having a basement depth map for the Torquay Embayment (Madsen, 2002) and a two-way-time map and depth-time conversion factors for the deeper regions of the Bass Basin (Setiawan, 2000) implied densities for these stratigraphic columns that could be extrapolated more widely. Where no data were available, we have been guided by values listed in Berkman (2001) and Skladzien (2007). All models used a background density of 2670 kgm^{-3}.

The magnetic data were derived from surveys flown in Victoria and Tasmania from 1996 to 2008 by the Geological Survey of Victoria and Geoscience Australia. The onshore data were acquired at approximately 250 m line spacing, while the offshore data were typically flown at 400 m spacing, with the deeper water surveys often being flown at 800 m spacing. Subsequently the data were grid stitched together using a cell size of 0.05 minutes (approximately 80 m) by Geoscience Australia and released in 2010.

While the collected magnetic data were of high quality, much of the region is covered by magnetic Cretaceous to Recent basaltic volcanic rocks. These rocks have been subjected to rapid chilling, resulting in uneven distributions of fine grained magnetic minerals and hence high frequency changes in susceptibility and significant remanence (Clarke and Emerson, 1991). This made modelling their thicknesses problematic and the modelling of magnetic units beneath them correspondingly less reliable. Remanence problems were also present in the Jurassic Tasmanian Dolerite, but this covered significantly less area and was mostly over the homogenous, very weakly magnetic turbidites of the Mathinna Supergroup sedimentary rocks. Other lithologies have been assumed to carry no remanent magnetism.

Because rocks are only ferrimagnetic below the Curie Point, ~580°C (Ross et al., 2006), broad scale magnetic models such as in this study require depth estimates below which the rocks become non-magnetic. In the onshore Otway Basin of western Victoria, Purss and Cull (2001) gave an average gradient of approximately 31°Ckm^{-1}, implying a Curie Point depth of approximately 18 km. This is in close agreement with Driscoll (2006), who suggested a figure of approximately 32°Ckm^{-1} for those parts of Victoria most likely to be investigated for geothermal purposes. However, Purss and Cull (2001) gave an average figure of 25.6°Ckm^{-1} for cratonic areas of Victoria, implying a Curie Point depth of approximately 22 km. Because the region of investigation ranges from onshore cratons to offshore basins and there was little detailed control on the Curie Point depths for the region, an arbitrary maximum depth of 20 km has been used for the magnetic modelling. In order to obtain closer fit to the extracted profiles, remanence was modelled in the Cenozoic basalt and the Jurassic Tasmanian Dolerite. In all cases, the remanence was assumed to be normal, that is collinear and additive to the Earth's present magnetic field.

I have chosen not to use the model by Eshaghi et al. (2015) for Curie Point depths in southeastern Australia as the cell size, 25 km, was too large to effectively accommodate the models presented here. Furthermore, it seemed to be at variance with some of the known geology. For example, the seismic tomographic model presented by Rawlinson and Kennett (2008) indicates the likelihood of a strong thermal high (i.e. a very shallow Curie Point depth) under the Newer Volcanic Province in western Victoria. Here volcanism persisted well into the Holocene (Price et al., 2003). This region is less prominent in the image presented in Eshaghi et al. (2015) than parts of northern Victoria where no Cenozoic volcanism is known, or much of Tasmania, where the youngest volcanic rocks are little older than 8 Ma (Quilty et al., 2014).

At depths of 20 km or more, there are fewer controls on lithology distributions. The rocks are likely to be metamorphosed to amphibolite facies, and so their densities will be greater than for their near-surface equivalents. Furthermore, it is possible that the rocks at these depths are different to those at the surface. For example, in northern Tasmania, the Cooee Dolerite was intruded into the Burnie Formation. The dolerite geochemistry is consistent with intrusion early in the development of the rift in which the Burnie Formation lies (Crawford and Berry, 1992). However, the seismic or other data sets do not constrain the depth to the contact or indicate the nature of the underlying rocks.

By comparing the zircon inheritance patterns of Tasmanian granites and their host sedimentary rocks, Black et al. (2010) demonstrated that there is likely to be an older crustal unit below the Rocky Cape Group, here called 'Lower Rocky Cape Group'. This unit has an excess of 1600 to 1650 Ma zircons when compared to the presently outcropping rocks. Halpin et al. (2014) suggested that the Rocky Cape Group was deposited in environments that ranged from tidal to the outer shelf, implying that there must be another concealed package below it. There are no other data to indicate what this package might be or its physical properties. This unknown basement is assumed to be denser than the overlying rocks, but the geometries given to it are a best fit approximation. The models also assume a constant density over significant distances, illustrated in Section 7, the diagonal section (Figure 4-16). Here the major changes modelled in the depths to the top of the 'Lower Rocky Cape Group' might at least be partly explained by density variations in either the lower parts of the Rocky Cape Group or the 'Lower Rocky Cape Group'. Assuming as close as possible to a constant density was considered to be the simplest approach to this almost completely unknown package. The unit was assumed to be non-magnetic.

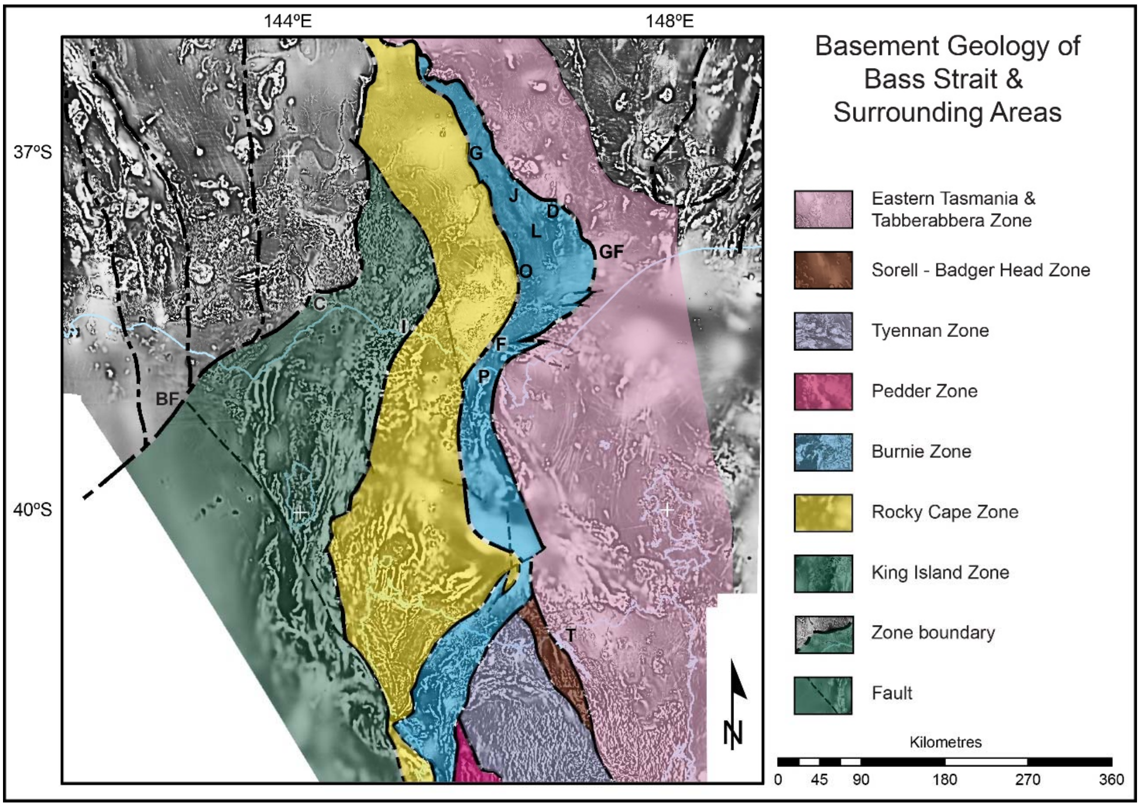


Figure 4-5. Interpreted basement geology of Bass Strait and surrounds; background image Automatic Gain Control (AGC) filtered TMI (see Dentith and Mudge, 2014). Same area as Figure 1. VanDieland and the Tabberabbera and Eastern Tasmania zones coloured. Thicker black lines mark the zone boundaries and thinner dashed lines mark faults. A marks the location of the Arthur Complex, C, the Ceres Gabbro, D, the Dolodrook Window, F, drill hole FOS1, G, the Glen Creek Window, I, the Proterozoic basalt on Phillip Island and the megacrystic basalt at West Point, J, the Jamieson Window, K, King Island, L, the Licola Window, O, the Boola Formation, P, Cape Liptrap, T, the Tamar Fracture Zone, BF marks the Bambra Fault and GF, the Governor Fault. Magnetic data from Geoscience Australia.

The ends of most sections indicate rises in the gravity responses. These have been accounted for by interpreting wedges of mantle, mostly with a density of 3000 kgm^{-3} and no magnetic response. Some of the rises may be an artifice caused by the ends of the sections. In other areas (e.g. the western end of line 5) the mantle wedge may be more realistic, since the water depths are significant and there has also been a significant amount of extension in forming the Otway Basin.

4.3 RESULTS

4.3.1 BASEMENT MAP

Figures 4-5 and 4-7 are plans of the final interpretation and Figures 4-9, 4-10 and 4-12 to 4-16 are of the interpreted cross sections. Figure 4-5 shows much of the basement of western Tasmania to continue across Bass Strait and under the Melbourne Zone of southern and central Victoria. The southwestern Melbourne Zone is underlain by the King Island Zone; the central part is underlain by the Rocky Cape Zone and the eastern part by the Burnie Zone. Although the boundaries have been principally based on the magnetic data, other data were also considered and the boundaries often modified to accord with them. As proposed by Reed (2001), the western edge of the Eastern Tasmanian Zone, the Tamar Fracture Zone, continues north to become the Governor Fault in eastern Victoria.

The eastern edge of the magnetic Grassy Group defines the eastern edge of the King Island Zone. This boundary places both the Ceres Metagabbro and the Proterozoic or Cambrian mafic rocks that crop out on the south coast of Phillip Island in the King Island Zone. In Bass Strait, Whelk 1 (Figure 4-2) is interpreted to have intersected basalts of the Ediacaran Grassy Group (Esso Exploration and Production Australia Inc., 1970). In south-central Victoria, the northeastern edge of the King Island Zone appears to be terminated by a northwest-striking fault that crosses the western edge of the Melbourne Zone (Figure 4-6). Where it crops out crossing the eastern edge of the Bendigo Zone, this unnamed fault dips at 60 to 70° to the southwest as it displaces Cambrian to Devonian rocks (Gray and Willman, 1991). The northwest-striking fault is interpreted to have been a Tabberabberan Orogeny (ca 380 Ma) or later reactivation of an older Tyennan fault. The western boundary of the King Island Zone is undefined.

In Tasmania, the Rocky Cape Zone is marked by a broad central magnetic high of 50 to 300 nT, coincident with the Spinks Creek Volcanics (an upper unit in the Smithton Basin; Calver et al., 2014). To either side, the Rocky Cape Group and the lower parts of **the** Smithton Basin are marked by 50 nT magnetic lows that are approximately 35 km wide. In Bass Strait, the eastern low becomes narrower, so that at the Victorian **coastline** it is no longer present. In Bass Strait, Bass 3 (Figure 4-2) (Esso Exploration

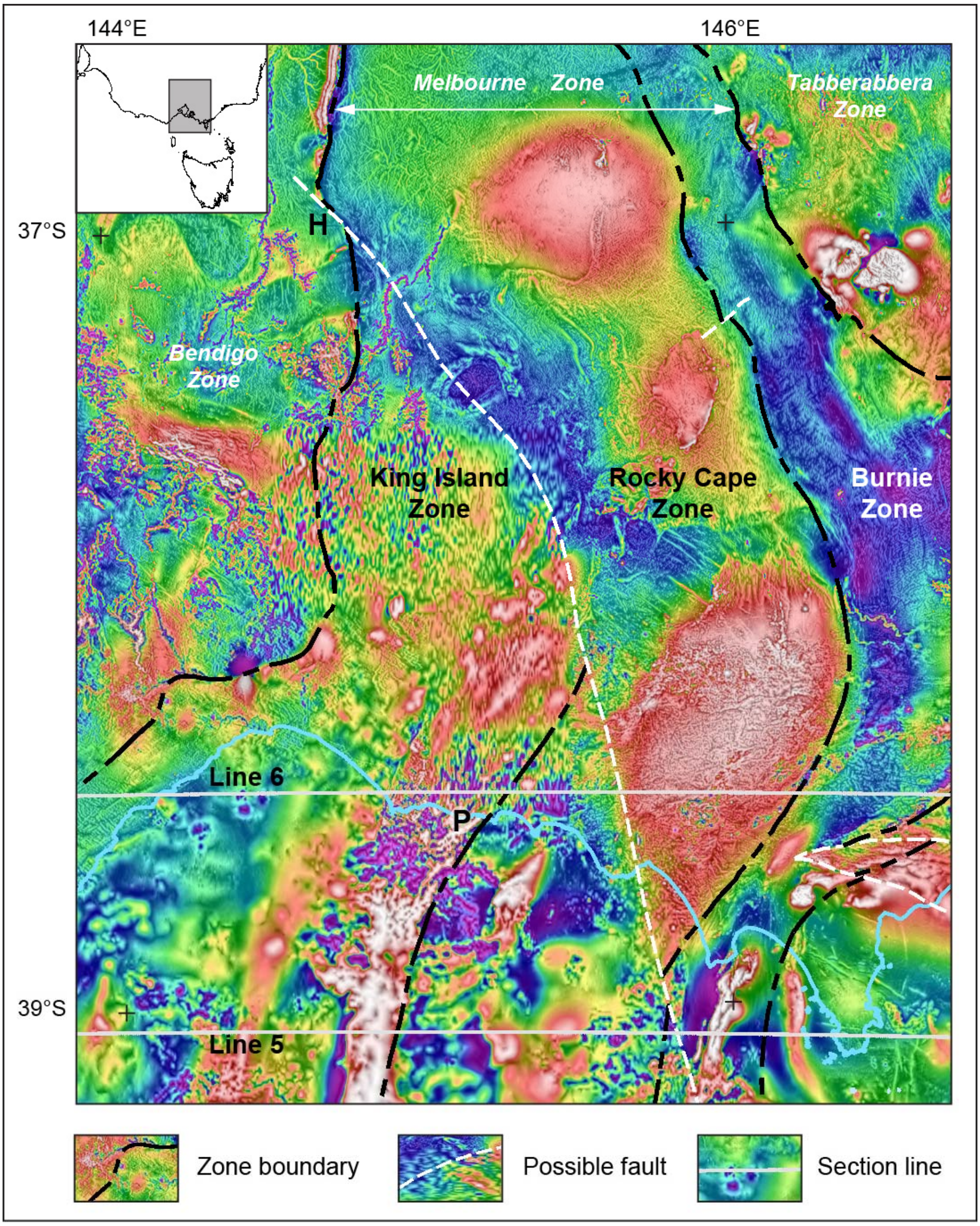


Figure 4-6. Zone boundaries, comparing Selwyn Block and Lachlan Orogen boundaries (smaller, italicised) overlaid on the total magnetic intensity with the intensity layer AGC filtered. What appears to be a fault truncates the 200 nT high on line 6. P marks the position of an Ediacaran MORB tholeiite and H the Heathcote Fault Zone where mafic volcanic rocks have a steeply SW-dipping foliation. The three broad magnetic highs in the Rocky Cape Zone are attributed to metamorphism of equivalents to the Spinks Creek Volcanics. Insert map shows location of main image. Magnetic data courtesy of Geoscience Australia.

Inc. 1967) bottomed in "interbedded quartzite, recrystallized siltstone and fine grainedandstone and black metamorphosed shale", of similar lithological character to the Rocky Cape Group and consistent with the magnetic interpretation. The magnetic parts of the Smithton Basin have a distinct linear pattern that strikes approximately 020° almost as far north as the Victorian coast (Figures 4-2, 4-5 and 4-7). Further north, beneath the Melbourne Zone, prominent magnetic highs occur along strike from magnetic highs interpreted to be sourced from the Spinks Creek Volcanics (Figure 4-6).

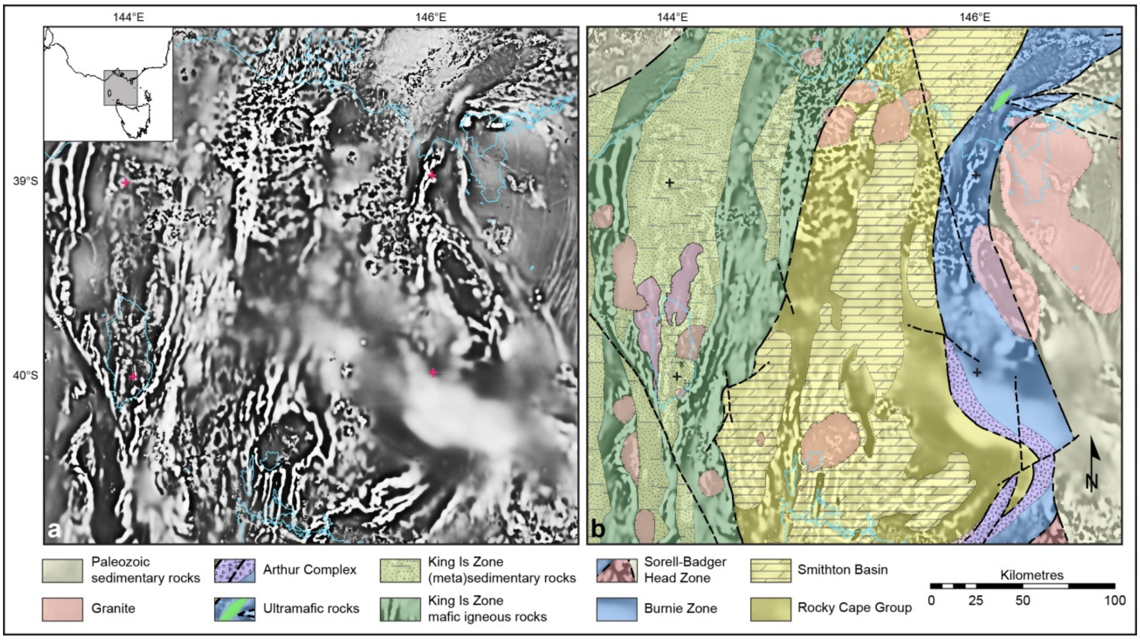


Figure 4-7. Detailed interpretation of the Proterozoic and Early Paleozoic basement of western and central Bass Strait and southern Victoria. a. Automatic gain control filtered total magnetic intensity, with 40% of the first vertical derivative added. Insert gives location. b. Interpretation overlaid on a. Thick dashed lines indicate zone boundaries, thinner lines major faults. The Rocky Cape Group generally has a much lower response and less linear character than the Smithton Basin. Magnetic data from Geoscience Australia.

In Tasmania, the western edge of the Burnie Zone is clearly defined by the 150 nT magnetic and 12 mGal gravity highs associated with the Arthur Metamorphic Complex. These highs continue approximately half way across Bass Strait (Figures 4-4, 4-5 and 4-8), apparently continuing either side of a broad magnetic and gravity high that is interpreted to result from approximately 5 km thickness of overlying Jurassic Tasmanian Dolerite. In Tasmania, the Burnie Zone is characterised by strong magnetic variation, with local highs of up to 1000 nT from the mafic-ultramafic rocks and the Ediacaran basalts and related rocks, and broader 100 nT highs with half wavelengths of 20 km that may be related to granites (Large et al., 1996). The highs persist across Bass Strait and into southernmost Victoria, but in central Victoria the interpreted Burnie Zone generally has a low magnetic response. Long wavelength highs are occasionally present along its eastern margin, but the origin of these is unknown and could either be due to Cambrian granite or Upper Devonian basalt intrusions (VandenBerg et al., 2000).

4.3.2 SECTIONS

Section 1, Tasmanian coast

The western end of the section (to approximately 60 km; Figure 4-9) is interpreted as two packages of metasedimentary rocks and three packages of basalt in the King Island Zone. Approximately 80 km north of the section, the westernmost magnetic package was intersected by Whelk 1 (Figure 4-2) and the subsequent petrology indicated the

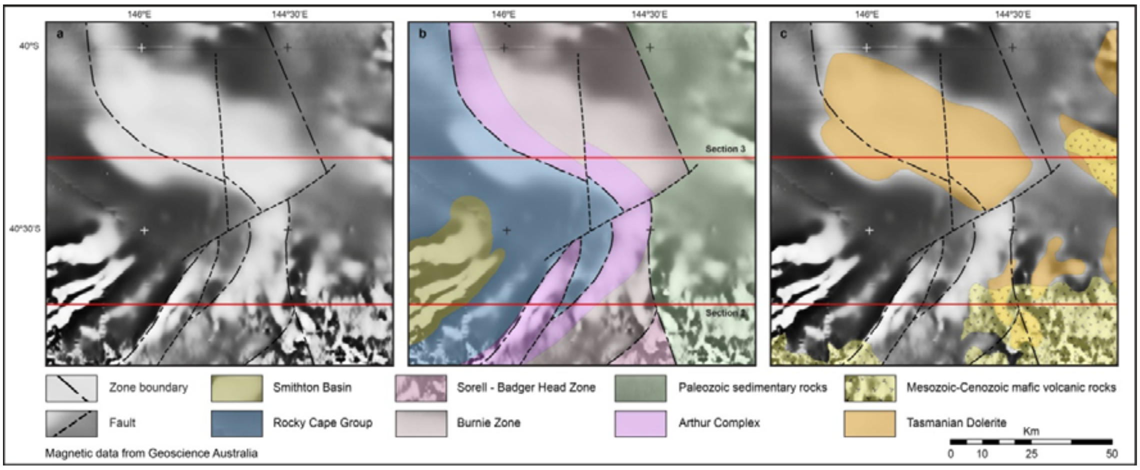


Figure 4-8. Interpretation of the magnetic high in the centre of the Bass Basin. a. AGC filtered magnetic data with major boundaries and faults. b. Basement interpretation. c. Jurassic Tasmanian Dolerite and younger mafic volcanic rocks.

presence of both siltstone and basalt (Esso Exploration and Production Australia Inc., 1970), perhaps equivalent to the Cryogenian-Ediacaran Grassy Group (Hall, 1998) that crops out on the eastern coast of King Island, where it unconformably overlies the Mesoproterozoic Surprise Bay Formation (Calver, 2012). The modelled susceptibility is 8.2×10^{-3} SI and the density 2745 kgm^{-3} above 20 km depth and 2760 kgm^{-3} below this. The central magnetic package has no onshore equivalents. It is modelled as having a range of susceptibilities from 8.1 to 25×10^{-3} SI and a density of 2695 kgm^{-3}. The eastern package was intersected by Clam 1, which is approximately 20 km south of the section (Lunt, 1970). This hole bottomed in argillite and basalt, the latter with a K/Ar whole rock age of 630±21 Ma, which suggests correlation with the Robbins Creek Formation, the basal unit of the Grassy Group. Most of this unit is modelled as having a

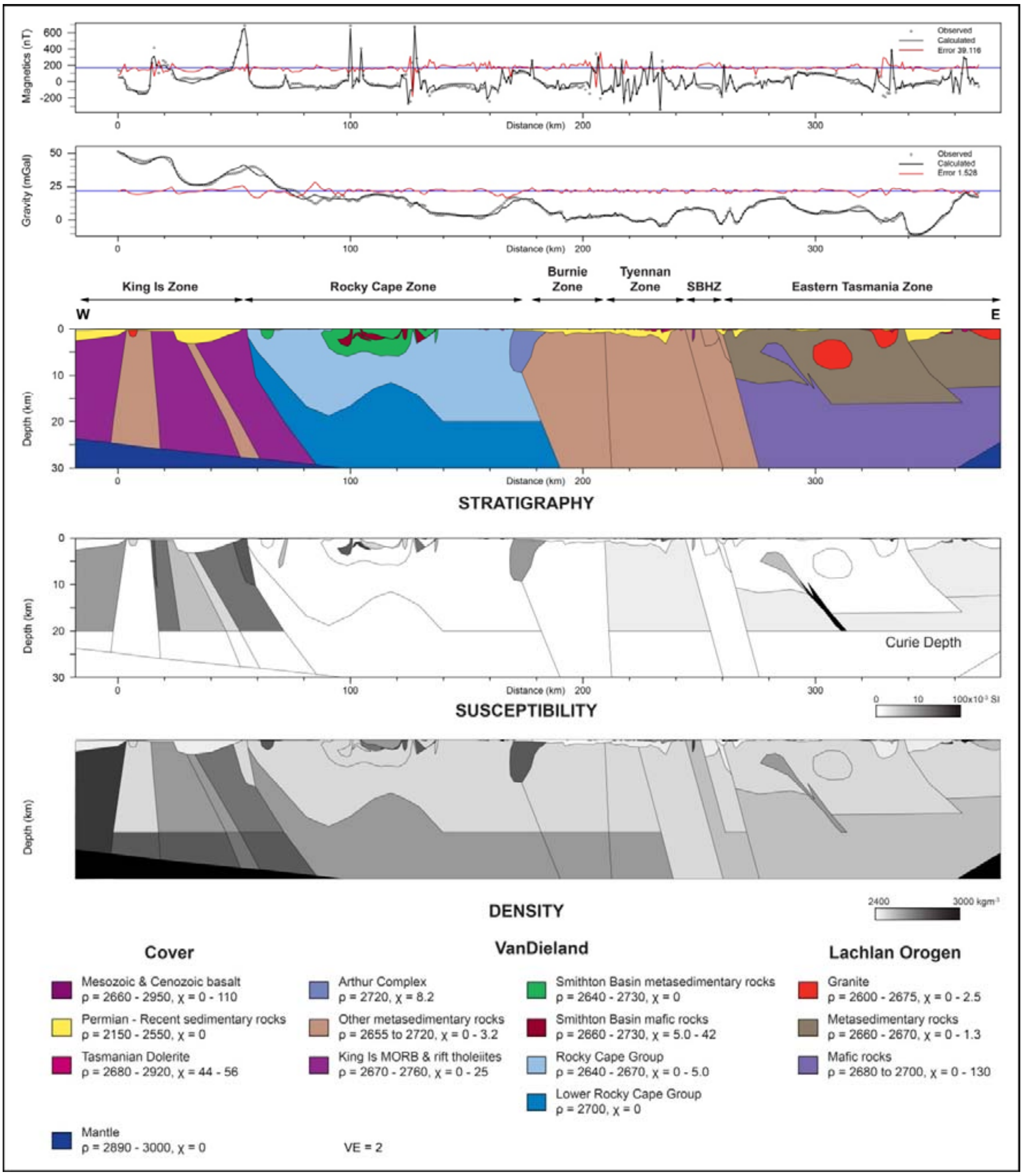


Figure 4-9. Section 1, along the Tasmanian coast. Location shown in Figure 4-2. SBHZ is the Sorell-Badger Head Zone. Susceptibilities as 10^{-3} SI units.

susceptibility of 13×10^{-3} SI and a density 2700 kgm^{-3}, but the eastern margin has a susceptibility of 23×10^{-3} SI and a density 2670 kgm^{-3}. This is consistent with the Neoproterozoic sequence on King Island dipping east and younging from (glacio)-marine sedimentary rocks into rift tholeiite and MORB in a back-arc basin setting. Two weakly magnetic sedimentary units are present. The western one has a susceptibility of 0.01×10^{-3} SI and a density 2655 kgm^{-3}, rising to 2720 kgm^{-3} at depth. It has no onshore equivalent, but appears to continue to the north under the Otway Basin. The eastern metasedimentary unit is interpreted as having a susceptibility of 3.2×10^{-3} SI and a density 2690 kgm^{-3}.

Both the magnetic and non-magnetic units are interpreted to have been intruded by weakly magnetic granite (2.5×10^{-3} SI) with a density of 2670 kgm^{-3}. These are all overlain by the Cretaceous-Tertiary Otway and Bass basins. The Otway Basin rocks are non-magnetic and a have density of 2550 kgm^{-3}. Further east, these young rocks are also non-magnetic but with a density of 2450 kgm^{-3}. Overall, the gravity response increases to the west; this is interpreted as resulting from the thinning of the crust associated with the development of the Otway Basin. The modelled density used for the mantle is 2890 kgm^{-3}. Most of the boundaries between the major rock packages dip to the east at 40° to 60°. An exception is the westernmost boundary, which dips to the west at approximately 75°. Steep westerly dips are modelled on the western ends of all of the sections. It is not clear whether they are real or artefacts resulting from being close to the section ends.

The boundary between the King Island and Rocky Cape zones does not crop out. This boundary is modelled with a listric geometry, dipping at 70° to the east near the surface, but flattening to less than 40° at depth, and lies from 55 to 85 km along the profile. To the east, the Rocky Cape Zone continues to 180 km, where it terminates against the Arthur Metamorphic Complex. The Rocky Cape Zone consists of three broad packages, an upper Smithton Basin, the Rocky Cape Group and the Lower Rocky Cape rocks. Onshore in western Tasmania, the Smithton Basin (including the overlying Scopus Formation) has approximately 4 km measured thickness (Everard et al., 2007), but the model suggests that it may reach a thickness of approximately 5.5 km in the deepest parts. It is mostly comprised of non-magnetic to weakly magnetic sedimentary rocks, including dolomite, diamictite and clastic sedimentary rocks ranging from conglomerate to mudstone (Everard et al., 2007). It also includes basalts of the Spinks Creek Volcanics, which to the south of the studied area have a median susceptibility of 31×10^{-3} SI units with a peak value of 110×10^{-3} SI (N=140 measurements, written communication from J. Everard, 18 Feb 2013). The Smithton Basin is a broad synclinorium in the centre of the Rocky Cape Zone, and is distinctly more magnetic than the underlying Rocky Cape Group (Figures 4-2, 4-3 and 4-9).

The Rocky Cape Group lies unconformably below the Smithton Basin. It crops out extensively in northwest Tasmania where it can be seen as marginal marine siltstone, mudstone and clean quartzite. Along Section 1, it is generally characterised as non-magnetic and with a density of 2640 kgm^{-3}. A small west-dipping unit at 72 km is weakly magnetic ($5.0x10^{-3}$ SI) and was slightly denser than other Rocky Cape Group units (2670 kgm^{-3}), perhaps due to a pyrrhotite-rich siltstone. These compare with susceptibilities recorded by the author that had a median value of $0.04x10^{-3}$ SI (N=124) in the quartzite-rich units and $0.16x10^{-3}$ SI in the siltstones (N=80).

The Lower Rocky Cape Group does not crop out, and so there is minimal control on any of the rock properties or body shapes assigned. Along Section 1, a constant density of 2700 kgm^{-3} was used. The unit shallows to 3.5 km deep in the western part of the section. The 70° dip of the contact with the King Island Zone and the 30° dip of the contact with the overlying Rocky Cape Zone suggest either west-directed thrusting or east-block-down extension; all of the models generated show this pattern. Another high is modelled at approximately 120 km along the section, where the Lower Rocky Cape Group was interpreted at approximately 12 km from the surface.

The Arthur Metamorphic Complex forms the boundary between the Rocky Cape Zone and the Burnie Zone and lies at approximately 170 km along the section. In western Tasmania, it is a zone of blueschist facies meta-sedimentary and mafic volcanic rocks that are interpreted to have formed in the Tyennan Orogeny and that underwent significant sinistral strike-slip movement (Holm and Berry, 2002) (Chapter 3). Geophysical forward modelling by Leaman and Webster (2002) suggests the complex has a shallow, steeply west-dipping part that overlies a deeper boundary that dips east at about 50°. Here, the shallow complex was modelled with a density of 2720 kgm^{-3} and a susceptibility of $8.2x10^{-3}$ SI. The area is covered by non-magnetic Cenozoic sedimentary rocks (density 2300 kgm^{-3}) and partly by Cenozoic basalt (density 2750 kgm^{-3}, susceptibility 0 to $24x10^{-3}$ SI, normal remanence 0 to 0.101 Am^{-1}). Cenozoic cover continues east along the section, overlying non-magnetic deep marine turbidite rocks of the Burnie Formation (density 2664 kgm^{-3}) in the Burnie Zone. The Cenozoic basalt is variably magnetic (susceptibility 0 to $28x10^{-3}$ SI, normal remanence 0 to 0.601 Am^{-1}) and dense (2670 to 2800 kgm^{-3}).

The eastern boundary of the Burnie Zone, at approximately 210 km along the section, is a fault zone that crops out on the coast at Penguin (Chapter 3). The fault has a modelled dip of 80° to the east. This fault separates two packages of rocks with only slightly different susceptibilities and densities (0 and 2670 kgm^{-3} for the Burnie Zone and $0.6x10^{-3}$ SI and 2664 kgm^{-3} for the Tyennan Zone) and so the control on the dip of the boundary was poor. In the coastal outcrops, the rocks in the western part of the Tyennan Zone are breccias, and thence a poly-deformed conglomerate, pelite and psammite succession that becomes progressively more metamorphosed to the east. After approximately 220 km along the section, the rocks are metamorphosed to amphibolite facies. Further east, the Proterozoic rocks are covered by non-magnetic Permian glacial deposits, sandstone and mudstone that are modelled with a density of 2460 kgm^{-3}. The Permian rocks continue to the east and conceal the non-magnetic Tyennan Zone basement. However, the Tyennan is inferred to be present as it crops out approximately 6 km south of the section as a poly-deformed quartz-mica schist (Calver & Everard, in Corbett et al., 2014) and is interpreted to extend to 245 km along the section line. Both Tasmanian Dolerite (density 2700 kgm^{-3}, susceptibility$18x10^{-3}$ SI) and Cenozoic basalt (density 2670 to 2680 kgm^{-3}, susceptibility 9.4 to $84x10^{-3}$ SI, normal remanence to 0.2 Am^{-1}) overlie the Permian rocks and the Tyennan basement rocks.

The Sorell-Badger Head Zone extends from 245 km to 260 km along the section. Both the western and eastern boundaries are interpreted to dip east at approximately 60°. The western boundary lies along strike from an outcrop of blueschist rocks; to the east are poly-deformed Neoproterozoic fine grained, rift-related deep water turbidite rocks, chert and metabasalt of the Port Sorell Formation (Calver and Reed, 2001). The rocks were mostly forward modelled as non-magnetic, with a density of 2670 to 2680 kgm^{-3}, but include a small, steeply dipping basalt body with a susceptibility of 6.3 $x10^{-3}$ SI and a density of 2740 kgm^{-3}. Median susceptibilities measured in the field were $0.05x10^{-3}$ SI (N=76) for the sedimentary rocks of the Port Sorell Formation.

The Badger Head Group is a ?Neoproterozoic meta-turbidite package that crops out from approximately 250 to 257 km. The rocks are modelled as non-magnetic, with a density of 2630 kgm^{-3}. Measured susceptibilities had a median of $0.10x10^{-3}$ SI (N=71). The model follows that by Zengerer (1999), who suggested that the Badger Head rocks had been thrust over the Port Sorell Formation, so that Port Sorell Formation underlies the Badger Head Group.

A narrow lens of serpentinite, layered orthopyroxenite, websterite and gabbro occurs at the eastern edge of the outcropping Badger Head Group (Reed et al., 2002). It has a strong magnetic response that indicates it almost entirely occurs to the south of the section. The model includes a lens of this mafic to ultramafic assemblage, but it is too small to be shown on Figure 4-9. The modelled susceptibility of 2.5×10^{-3} SI and density of 2730 kgm^{-3} compares with a median measured susceptibility of 26×10^{-3} SI (N=112) for the serpentinite and 0.38×10^{-3} SI (N=51) for the gabbro. The unit is interpreted as defining the leading edge of a zone of thrusting associated with activity along the Tamar Fracture Zone. To the east, rocks from both western and eastern Tasmania are present. The western Tasmanian rocks include equivalents to the Mt Read Volcanics, the Denison Group and the Owen Conglomerate (Reed and Vicary, 2005), none of which are known from eastern Tasmania and so imply a Proterozoic substrate. However, there are also thin fault slices of Devonian deep marine turbidite rocks that correlate with rocks of the Eastern Tasmanian Zone (Rickards et al., 2002; Mineral Resources Tasmania Data Management Group, 2011). Because of scale constraints, the present section shows the 3 km-wide area at approximately 260 km as dominantly associated with western Tasmania. At depth, the boundary between eastern and western Tasmania has a modelled dip of 60° to the east. Its eastern edge is covered by Permian and Cenozoic sedimentary rocks (density 2310 to 2320 kgm^{-3}), flat-lying Tasmanian Dolerite (modelled susceptibility 42×10^{-3} SI, measured median susceptibility 36×10^{-3} SI from 164 measurements, density 2790 kgm^{-3}) and Cenozoic basalt (modelled susceptibility 2.5×10^{-3} SI, measured median susceptibility 18×10^{-3} SI from 187 measurements, density 2700 kgm^{-3}).

Paleozoic meta-turbidites and granites dominate outcrops in the Eastern Tasmanian Zone. Mostly these rock types have no discernible modelled magnetic response, although the easternmost package of metasedimentary rocks has a modelled susceptibility of 1.2×10^{-3} SI. This compares with the 0.15×10^{-3} SI median measured magnetic susceptibility of the Mathinna Group. The western part of the Mathinna Group coincides with long-wavelength magnetic responses (Figures 4-2, 4-3 and 4-5), which are interpreted to be sourced from magnetic bodies at depth. Seismic sections suggest the presence roll-over anticlines and a strengthening in the seismic responses below depths of approximately 15 km (Chapter 3, Figure 3-17). Similar interpretations have been made in western Victoria and have been attributed to mafic volcanic rocks (Cayley et al., 2011). Roach and Leaman (1996) have previously modelled the broad magnetic high as of mafic rocks. The forward model (Figure 4-9) suggests the anomaly is caused by a body with magnetic susceptibilities that vary from 5.0×10^{-3} SI to 130×10^{-3} SI and densities between 2700 and 2690 kgm^{-3} for the upper and lower sections of the anticline. Interpretation of this part of the section is limited as the model exceeded the software limits on the number of surfaces allowed. If more surfaces were allowed, more complex body shapes would have been modelled from 320 km along the section. In outcrop, the Ordovician meta-turbidite rocks above and to the west of the magnetic high are strongly deformed, with recumbent folding mapped (Seymour et al., 2011). Further east, the deformation is of lower intensity and the metamorphic grade is lower (Patison et al., 2001) and the rocks exposed there are Silurian and Devonian; the relationship between the older and younger sequences may either be faulted or unconformable (Seymour et al., 2011). The forward model suggests slightly different densities for the western (2670 kgm^{-3}) and eastern (2660 kgm^{-3}) areas.

Granites form a significant part of the outcrop of the Eastern Tasmanian Zone. They are non-magnetic, and have modelled densities between 2600 kgm^{-3} and 2675 kgm^{-3}. Both the granites and country rocks are overlain by Tasmanian Dolerite, Cretaceous and Cenozoic sedimentary rocks and Cenozoic basalt. The sedimentary rocks include the southeastern end of the Bass Basin, the Cretaceous and Paleocene Durroon Sub-basin. Forward modelling of the associated 15 mGal gravity low suggests a maximum depth of 3.4 km with a density of 2400 kgm^{-3}. Tasmanian Dolerite crops out in two areas and is also inferred to underlie the eastern end of the Durroon Sub-basin. The modelled magnetic susceptibilities fall from 56×10^{-3} SI in the western body to 43×10^{-3} SI in the eastern part, while the density increases from 2680 to 2920 kgm^{-3}. The central body is also remanently magnetised (0.5 Am^{-1}). Minor thin sheets of basalt are present in the western part of the Eastern Tasmanian Zone. These range in susceptibility from 6.3 to 110×10^{-3} SI, with three bodies also being normally remanently magnetised (0.1, 2.2 and 8.7 Am^{-1}). Densities range from 2710 to 2950 kgm^{-3}. These are the edges of a much larger body that lies offshore, mostly between lines 1 and 2.

Section 2, offshore Tasmania

The first 80 km of this section (Figure 4-10) is almost identical to the similar part of Section 1. This includes all of the King Island Zone and the western Rocky Cape Zone. Elsewhere, the section is mostly across deeper water than Section 1 and so the magnetic sources are deeper. Hence there are fewer high frequency magnetic responses. Most of the modelled high frequency responses represent the seaward extensions of basalt or dolerite occurrences mapped onshore.

The Smithton Basin extends from 61 to 179 km along Section 2. The upper unit of the Smithton Basin (between 80 and 180 km) is modelled as weakly to non-magnetic, with a maximum susceptibility of 1.3×10^{-3} SI. Densities range from 2600 to 2660 kgm^{-3}. The Spinks Creek Volcanics, present between 95 and 180 km, is folded (Figure 4-10). In the west, the folds have approximately north-south axes but from 130 km the axes trend towards 070° (Figure 4-11). Modelled dips were typically 60 to 70° where the section was normal to the structural grain. In the east, many modelled dips were flatter, and are interpreted to reflect the apparent dips of the strata. Modelled susceptibilities range from 5.7 to 60×10^{-3} SI, and densities from 2620 to 2810 kgm^{-3}. Northwest-dipping faults are also interpreted at approximately 160 km. The lower Smithton Basin is generally magnetic, with susceptibilities ranging from 0.01 to $17 \times 10^{--3}$ SI.

Apart from a small unit at71 km with a susceptibility of 3.7×10^{-3} SI, the Rocky Cape Group is interpreted to be non-magnetic. The modelled density was 2620 to 2640 kgm^{-3}. Below this, the Lower Rocky Cape Group is non-magnetic, with a density of 2690 kgm^{-3}, increasing to 2700 kgm^{-3} below 20 km depth. Minor amounts of non-magnetic Cenozoic sedimentary rocks with density values between 2180 to 2400 kgm^{-3}, and basalt rocks with magnetic susceptibilities between 1.3 and 110×10^{-3} SI and densities between 2670 and 2800 kgm^{-3}, overlie the Rocky Cape Zone. The most westerly basalt occurrence is modelled with normal remanence of 0.7 Am^{-1}.

The Arthur Metamorphic Complex is present from 200 to 220 km. The magnetic response suggests a body that was repeated by a late Tyennan steeply west-dipping fault, possibly during deformation $\mathit{Ca}D_3$ of Holm and Berry (2002). Both parts of the Arthur Metamorphic Complex have similar modelled magnetic (11 and 12×10^{-3} SI) and density (2695 and 2700 kgm^{-3}) characteristics. At depth, the Arthur Metamorphic Complex is modelled as a single boundary dipping east at approximately 60°.

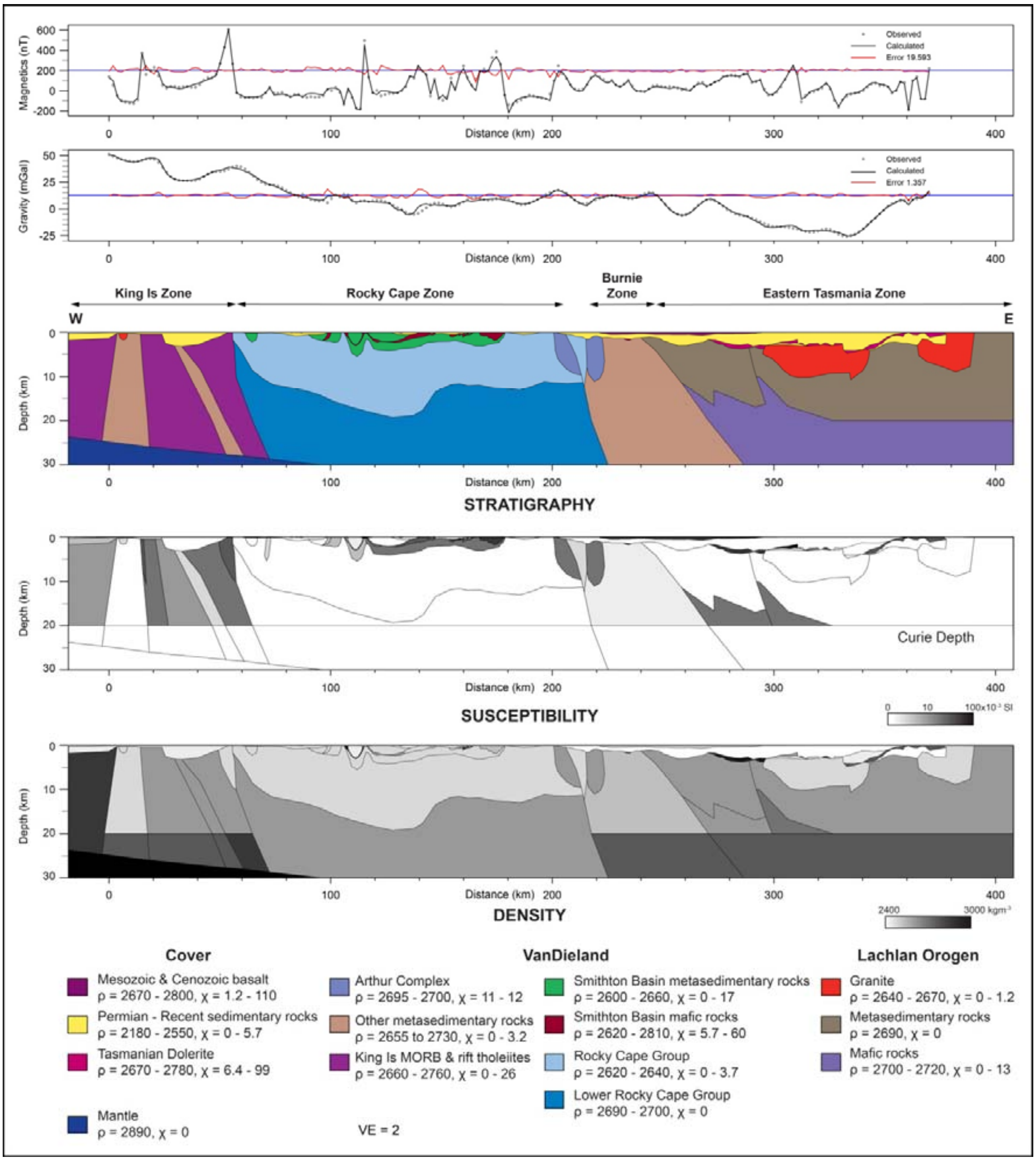


Figure 4-10. Section 2, offshore Tasmania. Location shown in Figure 4-2. Susceptibilities as x10^{-3} SI units.

The Burnie Zone is weakly magnetic (susceptibility 1.9x10^{-3} SI), with a density of 2680 kgm^{-3} increasing to 2720 kgm^{-3} at depth (Figure 4-10). This increase in density is attributed to either a higher metamorphic grade at depth or a change in lithology to an underlying mafic oceanic crust. The Burnie Zone is interpreted to be successively overlain by flat-lying, minor amounts of Jurassic Tasmanian Dolerite (susceptibility 42x10^{-3} SI, density 2710 kgm^{-3}), Cenozoic sedimentary rocks (non-magnetic, density

2240 kgm^{-3}) and Cenozoic basalt (susceptibility 53x10^{-3} SI, density 2670 kgm^{-3}). The eastern boundary of the Burnie Zone is considered to be the northern extension of the Tamar Fracture Zone, as the Sorell-Badger Head Zone pinches out to the south of the section (Figure 4-5). The boundary, at approximately 250 km, has a modelled dip of 35° to the east.

The eastern end of the section has a basement interpreted to be a northern extension of the Eastern Tasmanian Zone. As described above, these comprise Early Paleozoic sedimentary rocks (non-magnetic, with a density of 2690 kgm^{-3}) and Devonian granites (non-magnetic, with density 2640 kgm^{-3}). The model requires a deeper, dense layer (2700 to 2720 kgm^{-3}) that, in the west was also magnetic (11 to 13x10^{-3} SI). This is consistent with Section 1, and is considered to be a package of interlayered mafic and sedimentary rocks in west-verging slices thrust onto the Proterozoic craton.

Permian-Triassic Parmeener Supergroup sedimentary rocks, Jurassic Tasmanian Dolerite, Cretaceous Durroon Basin and Cenozoic sedimentary rocks and Cenozoic volcanic rocks all overlie the Paleozoic basement. Outcrops of these units occur on Tasmania and small offshore islands, and are imaged along the western end of deep seismic line AGSO 148/04. Only minor amounts of Parmeener Supergroup crop out, however it has a modelled thickness up to 2 km on the eastern end of the section (non-magnetic, density 2320 to 2350 kgm^{-3}) to satisfactorily match the observed and calculated geophysical responses. However, the gravity grid in this area had no data points and relied on extrapolations from onshore and from an offshore survey approximately 30 km away. Neither does the Parmeener Supergroup have a distinct seismic response. Thus, the modelling here has significant uncertainties Semi-continuous bodies of Tasmanian Dolerite (susceptibility 0.6 to 99x10^{-3} SI, remanence to 6.45 Am^{-1}, density 2670 to 2780 kgm-^{3}) underlie the younger sedimentary rocks. The dolerite bodies are typically between 0.4 and 1 km thick. They are generally modelled as flat-lying, but at approximately 350 km, dolerite below the Durroon Basin has a dip of up to 15°. It is interpreted as having been tilted during the basin formation in the Aptian (Hill et al., 1995). The basin is well imaged by seismic line AGSO 148/04, giving a maximum depth of 4.2 km. With a density of 2240 kgm^{-3} for the overlying Cenozoic sedimentary rocks, this implied a density of 2360 kgm^{-3} for the basin fill.

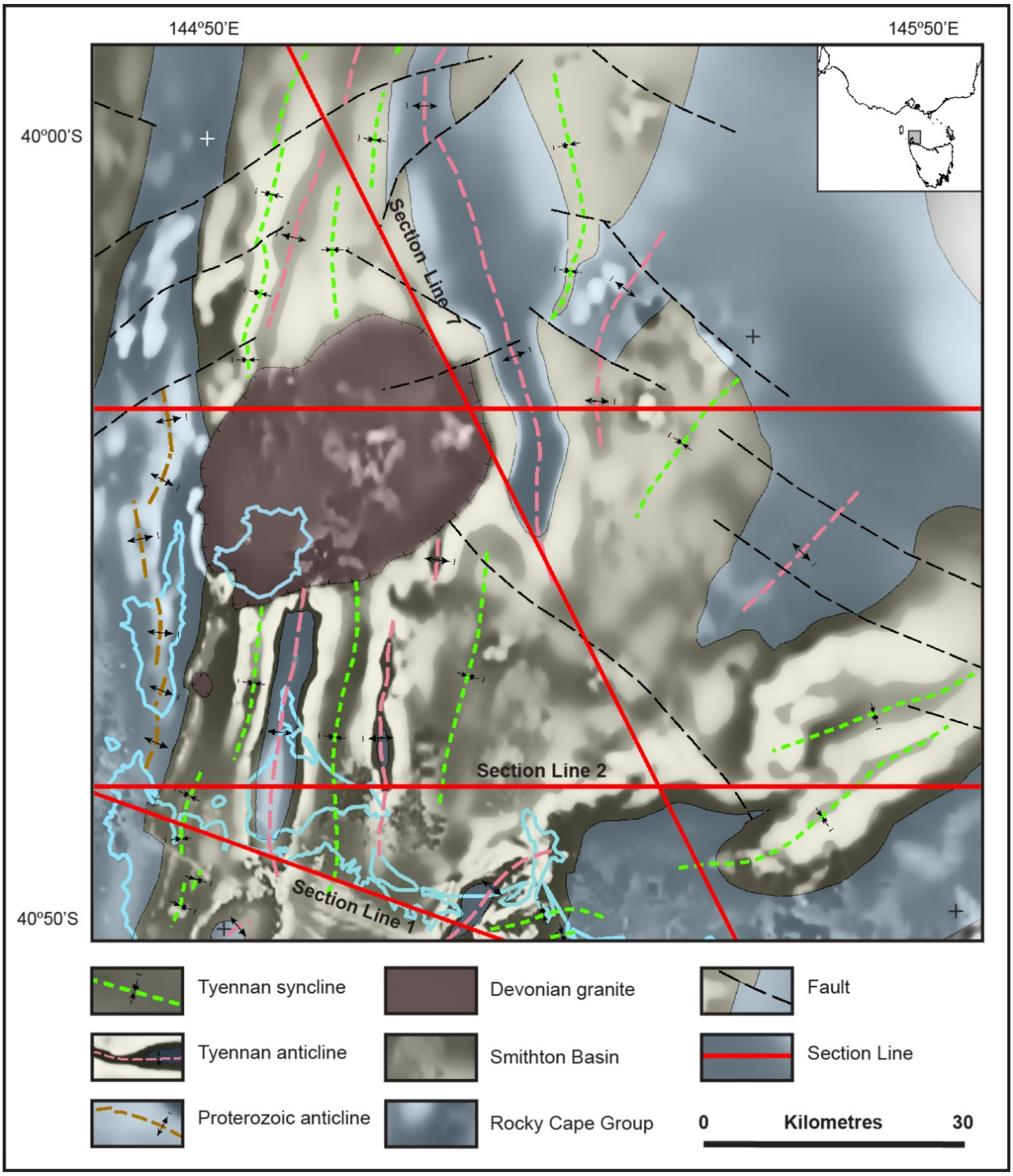


Figure 4-11. Magnetics of offshore northwestern Tasmania, showing interpreted fold axes extrapolated from onshore mapping. Inset gives location. Magnetic data from Geoscience Australia.

Cenozoic basalt is interpreted between approximately 220 and 295 km. This is interpreted as having a maximum thickness of 0.7 km, with a susceptibility of 57×10^{-3} SI and a density of 2670 kgm^{-3}. The magnetic images show the unit to be a northern extension of the flows described in Section 1.

Section 3, Central Bass Basin

The western end of this profile is southwest of King Island, while its eastern end reaches the Furneaux Group in eastern Bass Strait (Figures 4-2 and 4-12). It crosses a broad magnetic high that is from a body that lies under the Bass Basin. The first 125 km of the profile is within 5 km of AGSO seismic line 148/06. Whelk 1 (see above, Section 1) was drilled approximately 40 km north of the western end of the profile. The section was pinned at 355 km to the extracted profile, as the eastern end of the extracted profile was approximately 25 nT below the null model value.

The western end of the profile includes both metasedimentary and rift tholeiitic mafic volcanic rocks of the King Island Zone. Metamorphic monazites from sedimentary rocks near the southern tip of King Island yielded ages of approximately 1290 Ma (Berry et al., 2005), approximately 12 km north of 40 km from the western end of the section. The modelling is consistent with older boundaries dipping east at approximately 45° having been truncated by later faults dipping west at approximately 50°. In map view, the smaller, late, northeast-striking faults truncate the older dominantly north-striking major boundaries. The Braddon River Fault, with a displacement of up to 70 km, is an exception to this pattern, as it strikes at 330°. This Tyennan Orogeny fault lies 35 km from the western end of the section and dips west at approximately 70°. More detailed mapping and modelling by Meffre et al. (2004) indicated similar dips for the early boundaries. The eastern boundary of the King Island Zone was taken as the easternmost magnetic unit, at approximately 70 km from the western end of the section. Here it dips east at approximately 55°, but is cut by a later fault striking at 030° with an apparent dip of 45° to the west (Figure 4-12).

In the mafic volcanic rocks, susceptibilities range from 0.63 to 42×10^{-3} SI units, with most close to 10×10^{-3} SI. Densities range from 2670 kgm^{-3} to 2800 kgm^{-3}. Modelled susceptibilities in the sedimentary rocks are less than 0.6×10^{-3} SI, but mostly 0. Susceptibilities collected from Stokes Point on King Island, approximately 12 km north of the section, had a median of 0.19×10^{-3} SI (n=67). Modelled densities ranged from 2680 to 2690 kgm^{-3}.

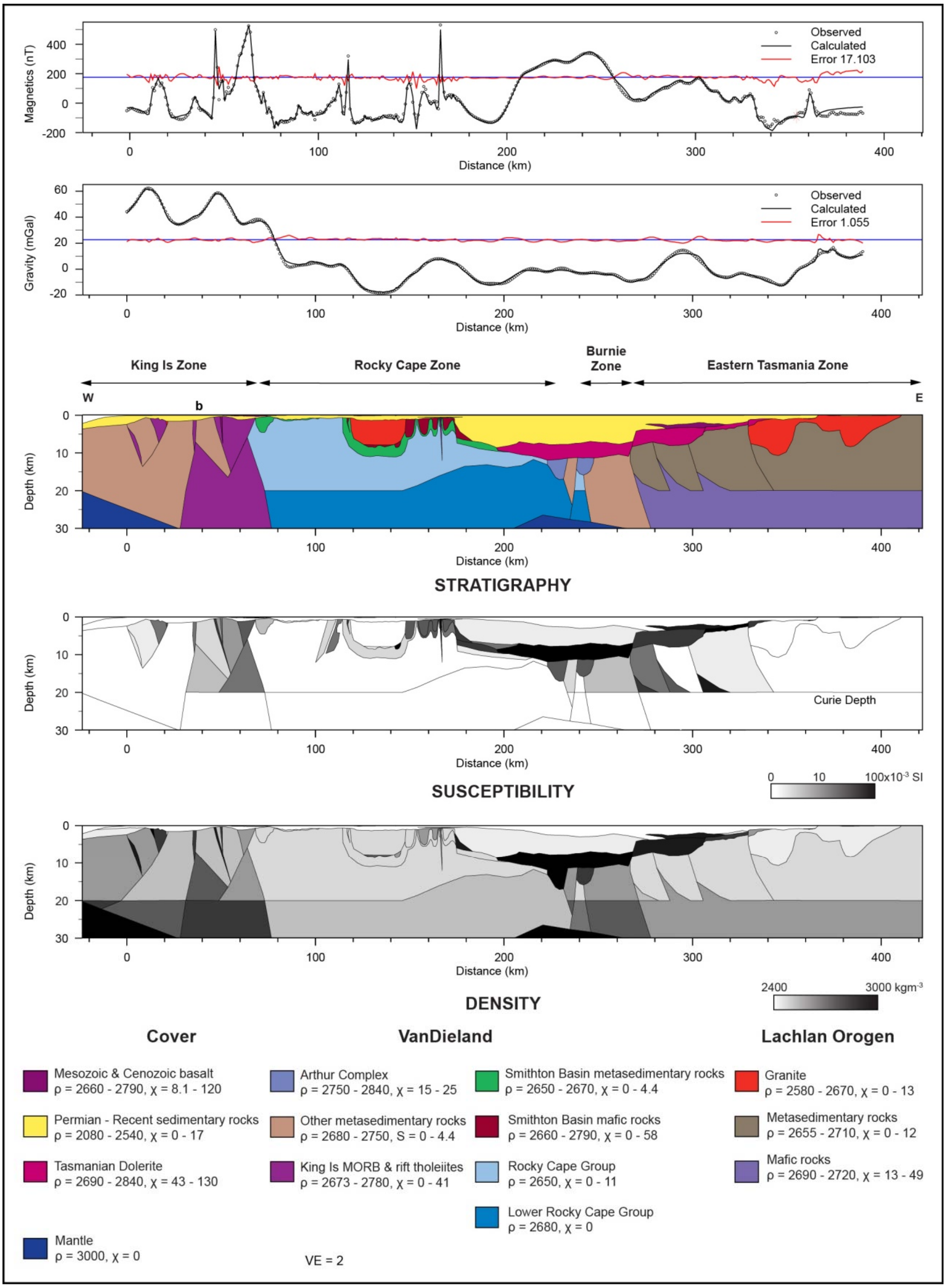


Figure 4-12. Section 3, central Bass Basin. Location shown in Figure 4-2. Susceptibilities as 10^{-3} SI units. b indicates the location of the BraddonRiver Fault.

the Smithton Basin is projected north from outcrops in northwest Tasmania. However, a west-dipping outcrop of the basal unit of the Smithton Basin (Forest Conglomerate) is present on Albatross Island (Everard et al., 1997), approximately 10 km south of the section at ~100 km from the western end of the section. This conglomerate is coincident with the eastern edge of a magnetic low, and accordingly the Smithton Basin was interpreted to extend from 68 km to 102 km along the section. The interpretation suggests a 0.3 km thick, generally flat-lying layer of Smithton Basin rocks in the east (susceptibility 0.6, density 2650 kgm^{-3}) and a upright syncline in the west with the limbs with dips of up to 60° (susceptibility $5.7x10^{-3}$ SI, density 2660 kgm^{-3}).

Further east, the Smithton Basin rocks have been intruded by the Devonian Three Hummock Island Granite. However, the distinctive fold patterns in the magnetics continue north of the intrusion (Figures 4-7 and 4-11), and accordingly the Smithton Basin is interpreted to occur beneath the granite. In the weakly magnetic parts of the Smithton Basin, the rocks are modelled with susceptibilities of 2.5 to $23x10^{-3}$ SI and densities between 2650 and 2690 kgm^{-3}. The magnetic units are interpreted to correlate with the Spinks Creek Volcanics, with susceptibilities that range from 5 to $130x10^{-3}$ SI and densities from 2660 to 2790 kgm^{-3}. The magnetic units are mostly modelled as upright folds with steeply dipping to vertical limbs.

Hunter Island lies approximately 20 km south of the section line at ~110 km along the profile. Here, Rocky Cape Group quartzite and siltstone crop out in north-striking moderately to steeply dipping limbs of folds that have been overprinted by west-dipping reverse faults (Hall, 2001). Two of the siltstone units are correlated with magnetic responses that continue north to the section and have modelled susceptibilities of 2.5 and $16x10^{-3}$ SI units and densities of 2650 and 2660 kgm^{-3}. Elsewhere, the Rocky Cape Group is non-magnetic, with a density of 2650 kgm^{-1}.

The magnetic low at 153 km (Figure 4-12) was interpreted as the core of an anticline that exposed the Rocky Cape Group in the Mesozoic. Drill holes Bass 3 (Esso Exploration Australia Inc, 1967) and White Ibis 1 (Premier Oil Australasia, 1999) approximately 30 km north of the section line, lie on the edge of this low, and intersected quartzite typical of those in the Rocky Cape Group.

The Lower Rocky Cape rocks underlie the entire Rocky Cape Zone. This package is non-magnetic and is modelled with a density of 2680 kgm^{-3}. The unit appeared to decrease in depth to the east, under the Bass Basin.

Both Mesozoic to Cenozoic sedimentary rocks and Cenozoic basalt overlie the Rocky Cape and Smithton Basin sequences. The sedimentary rocks are non-magnetic and have a modelled density of 2080 kgm^{-3}, while the basalt has susceptibilities that range between 8.0 and120x10^{-3} SI, have modelled remanence between 0 and 3.4 Am^{-1} and densities that vary from 2660 to 2720 kgm^{-3}.

The Bass Basin is interpreted to extend from 174 km to 330 km along the section. Setiawan (2000) suggested successive west-dipping normal faults stepping downwards from the eastern edge of the basin and a maximum basin thickness of 4.4 s TWT (equivalent to approximately 8.8 km depth) at 198 km along the section. Most of the basin is ~7.5 km thick. The western, deepest parts of the basin have a modelled susceptibility of 18x10^{-3} SI and a density of 2540 kgm^{-3}, while most of the basin has a modelled susceptibility of 0.6x10-3 SI and a density of 2530 kgm^{-3}, with the upper part of the basin modelled as non-magnetic and having a density of 2200 kgm^{-3}. A wedge of mantle (density 3000 kgm^{-3}) is included below the Bass Basin to allow for the crustal thinning associated with basin formation. This disagrees with the model of Kennett and Salmon (2012) which suggested a mantle depth of over 30 km. However, their model used a 0.5 x 0.5° grid (approximately 55 x 45 km), which may be too coarse to allow for any rise in the mantle under the Bass Basin. Furthermore, their model did not use any data from the area of the Bass Basin, and so any mathematical interpolation over the region would not take its presence into account.

A broad 300 nT magnetic anomaly coincident with a complex 22 mGal gravity anomaly occurs near the deepest part of the Bass Basin. This is interpreted to result from a 1 to 5 km thick sheet of flat-lying Tasmanian Dolerite, the top of which lies immediately below the Bass Basin. This dolerite horizon extends from 195 km to 365 km along the section, with the deepest and thickest parts in the west and the thinnest and shallowest parts in the east. Interpreted susceptibilities and densities vary from 43 to 130x10^{-3} SI and densities from 2690 kgm^{-3} to 2840 kgm^{-3}. Poonboon 1 was sited on the western end of the feature and bottomed at 3266 m in Cretaceous siltstone without encountering any mafic rocks (Trigg et al., 2003), implying that the causative body lies below this depth. In this region, the top of the dolerite has a modelled depth of approximately 7.4 km. An overlying Cretaceous basaltic volcanic unit contributed to the broad magnetic feature (susceptibility 74x10^{-3} SI, density 2790 kgm^{-3}), but it is considered to be a less significant factor as it is interpreted to be less than 1 km thick. Chat 1 was drilled approximately 14 km north of the 274 km point along the profile and bottomed in Upper Cretaceous olivine basalt at 3104 m (Furr, 1986).

The Arthur Metamorphic Complex forms the eastern boundary of the Rocky Cape Zone, but responses from this are largely obscured by the overlying Tasmanian Dolerite and Bass Basin. The Arthur Metamorphic Complex is interpreted to have been duplicated by a steeply (75°) west dipping fault active during the Tabberabberan Orogeny (Figures 4-8 and 4-12). The two parts of the Arthur Metamorphic Complex have slightly different characteristics. The western fault block is modelled to be more magnetically susceptible ($36x10^{-3}$ SI) and dense (2840 kgm^{-3}) than the eastern fault block ($29x10^{-3}$ SI and 2750 kgm^{-3}), suggesting that perhaps it contains a higher proportion of mafic rift tholeiites (c.f. Holm et al., 2003). As in the previous sections, the Arthur Metamorphic Complex is interpreted to be of limited thickness, and below it the boundary between the Rocky Cape and Burnie Zones dips to the east between 60 to 70°.

Apart from a small sliver faulted into the Arthur Metamorphic Complex at approximately 235 km, the Burnie Zone extends from approximately 245 to 265 km. The rocks are weakly magnetic, with a modelled susceptibility of $6.9x10^{-3}$ SI and a density of 2680 kgm^{-3} above 20 km, but below this the modelled density increases to 2730 kgm^{-3}. As in modelled Section 2, this may be due to either an increase in metamorphic grade or the presence of mafic oceanic crust. The eastern boundary of the Burnie Zone, the equivalent to the Tamar Fracture Zone, is modelled as dipping ~60° to the east.

As in Sections 1 and 2, from 265 km to the end of the section, the rocks are interpreted to be Paleozoic turbidite rocks underlain by oceanic crust, similar to the interpretation of the 2006 Victorian deep seismic survey (Cayley et al., 2011). The sedimentary rocks are modelled as weakly magnetic (0 to $14x10^{-3}$ SI, but mostly 0), with densities ranging from 2655 kgm^{-3} to 2710 kgm^{-3}. The more mafic rocks (susceptibilities 29 to $67x10^{-3}$ SI above the Curie Depth, densities 2680 to 2700 kgm^{-3}) appeared to lie in the hanging walls of west-verging reverse faults with dips of approximately 60°.

Granite was the dominant near-surface lithology interpreted from 330 km to the eastern end of the section, although small areas of Tasmanian Dolerite are present to 340 km and from 358 to 365 km ($109x10^{-3}$ SI, density 2800 kgm^{-3}) and a small area of Paleozoic metasedimentary rocks crops out at 375 km, on Chappell Island. The granites are all modelled as being non-magnetic. Modelled densities range from 2590 to 2600 kgm^{-3}.

Section 4, North King Is

Principal controls on this section include the outcrops on the north end of King Island, from 86 to 96 km from the western end of the section, and the Furneaux Group islands, which are as close as 4 km from the eastern end. The western end of the section (Figure 4-13) finishes in water depths of over 1000 m, implying significant crustal thinning. The presence of Cretaceous Sorell-Otway Basin sedimentary rocks, of unknown thickness, to the west of the modelled section further complicates interpretation in this area. A wedge of mantle (density 3060 kgm^{-3}) has been added to partly take account of these factors, but the model does not satisfactorily resolve with the extracted profile at the western end. Another wedge of mantle (density 3000 kgm^{-3}) is added to the eastern end of the profile in order to account for the 25 mGal rise in the gravity response there. However, part of the reason for the gravity low at approximately 365 km along the profile may caused by a lack of data and gridding issues related to the merging of land and sea-based surveys. If so, then the amount of mantle that is modelled may be an over-estimate. Furthermore, the densities and thicknesses of the other lithologies modelled there may be in error.

The 70 nT lows from 18 to 27 km along the section are considered to be caused by problems related to stitching of two adjacent grids. No attempt has been made to fit bodies to them.

As in the previous sections, the western end is largely comprised of (meta)-sedimentary rocks interlayered with mafic volcanic rocks, with the mafic rocks dominantly in the east. From 0 to 50 km, the mafic rocks dip steeply to the west, but further east the modelled dips are to the east, initially at 80° or more, but by 120 km from the western

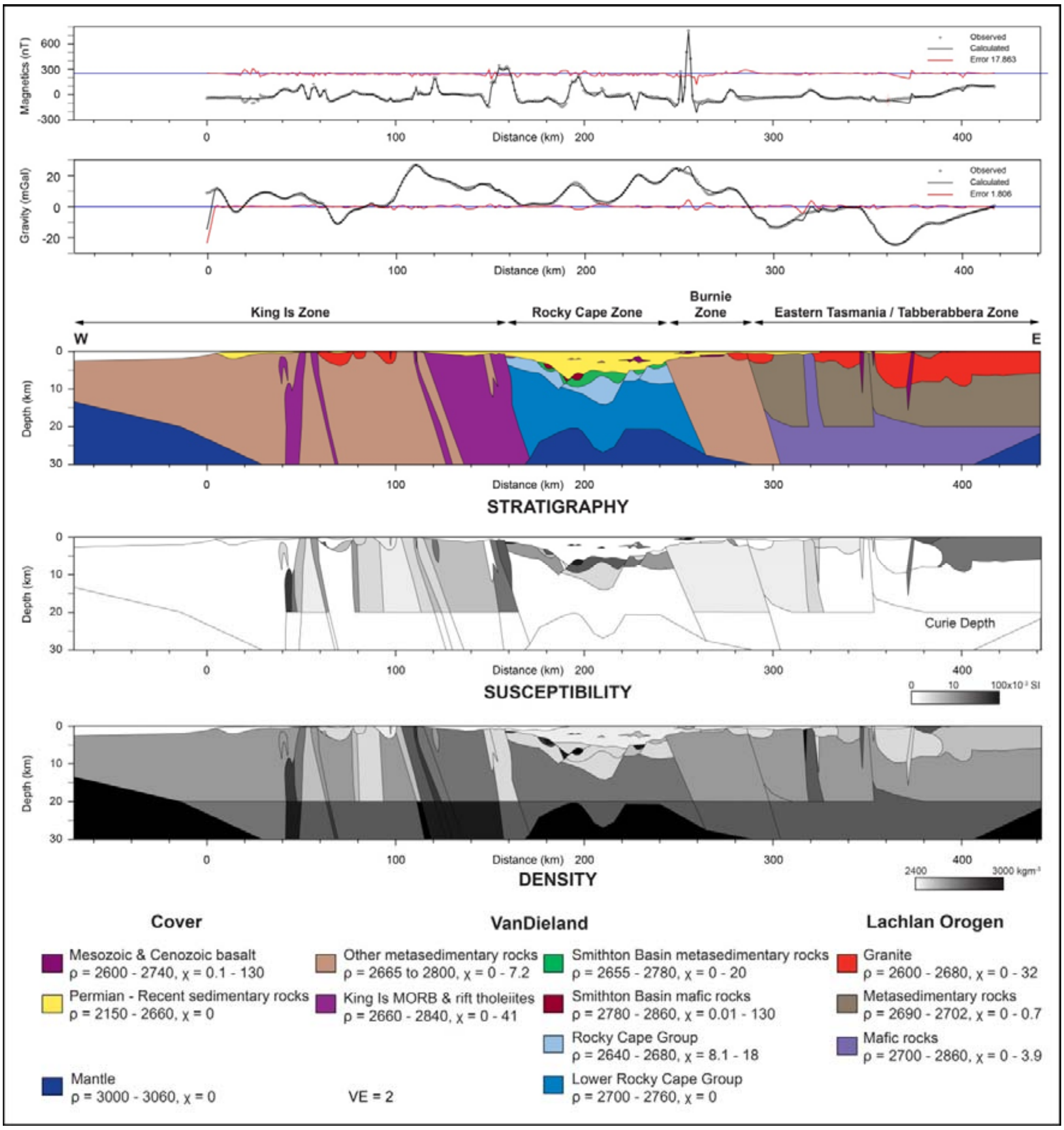


Figure 4-13. Section 4, King Island North. Location shown in Figure 4-2. Susceptibilities as 10^{-3} SI units.

end of the section, at 55°. Modelled susceptibilities range from 3.7 to 40×10^{-3} SI, and densities from 2665 to 2770 kgm^{-3}. At approximately 40 km, a lens modelled with a susceptibility of 3.2×10^{-3} SI and density of 2690 kgm^{-3} is considered to be of mafic volcaniclastic rocks folded within the sedimentary rock package. Mafic rocks in the east of the King Island Zone are northern extensions of those seen on the southeast coast of King Island, where the median susceptibility measured by the author in picrite was 34×10^{-3} SI (N=44), while that in basalt was 0.47×10^{-3} SI (N=10). Calver (2008) recorded dips to the east, typically between 40° and 60°. The easternmost magnetic unit is taken as the eastern boundary of the King Island Zone. This boundary dips east at approximately 60°.

The Mesoproterozoic (meta)-sedimentary rocks are typically modelled with susceptibilities of less than 3x10^{-3} SI, although a metasedimentary package at 78 km from the western end of the section has a modelled susceptibility of 7.2 x10^{-3} SI. This package can be traced south to the west cost of King Island, where Calver and Everard (2014) recorded the presence of garnet-bearing phyllite and slate of the 1300 Ma Surprise Bay Formation. Susceptibility measurements have a median of 0.14x10^{-3} SI with the maximum from 81 measurements of 0.40x10^{-3} SI, while amphibolite grade meta-sediments elsewhere on King Island had a median susceptibility of 0.18x10^{-3} SI and a maximum of 1.41x10^{-3} SI (N=137). These data are too low to account for the modelled value. Contact metamorphism of an unusual unit also seems unlikely, since the abnormally susceptible body was modelled as extending to depth of 20 km. It suggests that another lithology could be present in the anomalous package on the profile. Possibilities include interlayered basalt, an example of which were seen in Currie Harbour (median susceptibility 0.87x10^{-3} SI, upper value 1.5x10^{-3} SI, N=24), or amphibolite dykes similar to those present further south, at Ettrick Beach (median susceptibility 2.2x10^{-}3 SI, upper value 10.2x10^{-3} SI, N=20) (Calver, 2007). Three generations of dykes were recorded by Calver and Everard (2014) on Cape Wickham at the north end of King Island, and these had a median susceptibility of 0.79x10^{-3} SI, with an upper value of 13x10^{-3} SI (N=158). It suggests that amphibolite dykes too small to have been interpreted on the 500x500 m grid used in the present study might have been sampled by the extracted profile. The Braddon River Fault is interpreted to cut the section at approximately 40 km, but juxtaposes two equally weakly magnetic metasedimentary packages (1.3x10^{-3} SI) and so could not be modelled.

Both Neoproterozoic and Paleozoic granites are known on King Island, and both have been interpreted to intersect the section. The westernmost granite is non-magnetic at the near-surface, but at depth is associated with a magnetic layer with a modelled susceptibility of 32x10^{-3} SI in the west and 5.5x10^{-3} SI in the east. This magnetic layer is interpreted as either part of the granite or from a contact aureole. The former interpretation is favoured by analogy with granites in western Victoria, where many granites have magnetic rims and non-magnetic cores. Neoproterozoic granites are distinguished by the presence of many mafic dykes with susceptibilities of approximately 0.8x10^{-3} SI. As discussed above, dykes crop out on the section line, where they intrude the 760±12 Ma Cape Wickham Granite (Turner et al., 1998; Calver and Everard, 2014). The Cape Wickham Granite and associated metamorphism is coeval with the formation of the Smithton Basin and both have been attributed to extension related to the breakup of Rodinia (Chapter 3). Model 4 requires the granite to be weakly magnetic (1.8x10-3 SI), but a small dyke within it gives a modelled susceptibility of 3.3×10^{-3} SI. Measured responses from granites in the northern part of King Island gave a median susceptibility of 0.07×10^{-3} SI (N=34).

A 13 mGal gravity low at approximately 15 km from the western end of the section was modelled as a 2 km thick northern extension of the Cretaceous Sorell Basin, with a density of 2200 kgm^{-3}. At the eastern end of the zone, a 3 mGal low was attributed to a 1.3 km thick western extension of the Bass Basin, also with a density of 2200 kgm^{-3}.

As in previously described sections, the Rocky Cape Zone is interpreted as three layers, with the Neoproterozoic Smithton Basin overlying the Mesoproterozoic Rocky Cape Group, which in turn overlies an unseen Mesoproterozoic-Paleoproterozoic basement. The broad magnetic high that marks the Spinks Creek Volcanics in the Smithton Basin is less prominent, although even here the average response level over the Smithton Basin is approximately 100 nT higher than that over the adjacent Rocky Cape Group and this rose another 150 nT over the most magnetic parts of the Basin. The broad response is interpreted to be sourced from the metasedimentary rocks (susceptibilities 0 to 20×10^{-3} SI, density 2655 to 2780 kgm^{-3}), with the higher responses reflecting Spinks Creek Volcanic equivalents (susceptibilities 0.01 to 130×10^{-3} SI, densities 2780 to 2860 kgm^{3}). The Rocky Cape Group is interpreted as being relatively thin, typically 3 to 5 km thick, with susceptibilities that ranged from 3.9 to 18×10^{-3} SI. Densities range from 2640 to 2680 kgm^{3}. As it does not crop out and is non-magnetic, the Lower Rocky Cape rocks has fewer constraints, but are interpreted to comprise most of the lower half of the Rocky Cape Zone. The contact between the Rocky Cape and Lower Rocky Cape zones is somewhat flatter than in other sections, with modelled dips of approximately 20°. The assumed density above 20 km is 2700 kgm^{3} and below that, 2760 kgm^{3}.

Most of the Bass Basin on the section overlies the Rocky Cape Zone. Following Setiawan (2000), it is interpreted to have formed in northwest-striking rifts formed by southwest-dipping normal faults. On the section line, they are prominent at approximately 195 km and 220 km, where vertical displacements of up to 2.5 km are modelled. Modelled densities range from 2150 kgm^3 in the upper parts to 2660 kgm^3 in the deepest parts of the basin. Minor basaltic layers were present in many holes drilled in the Basin (e.g. Trigg et al., 2003). These are interpreted in several places, typically at depths between 1 and 3 km, and as giving rise to anomalies of up to 850 nT. Modelled susceptibilities are as high as 130×10^{-3} SI; densities range from 2600 to 2780 kgm^3. Three basalt bodies are modelled with remanence of 0.25, 1.4 and 2.4 Am^{-1}. A block of mantle (density 3000 kgm^3) is placed at the base of the model in order to compensate for the crustal thinning in the formation of the Bass Basin. This compensates for the broad 20 mGal gravity high at 250 km.

Unlike previous sections, there is no clear evidence of the Arthur Metamorphic Complex on the section line. Accordingly, the western boundary between the Rocky Cape and Burnie zones is modelled as a simple fault that dips east at 50°. The Burnie Zone is modelled as a simple body with a susceptibility of 1.8×10^{-3}SI and a density of 2700 kgm^3, rising to 2730 kgm^3 at depth. Drill hole Bass 2 was drilled approximately 25 km south of the section line 290 km from the western end. It intersected tuff that yielded a K/Ar age of 589±3 Ma (Hall, 1998). If the age is valid, it may have come from an equivalent to the Crimson Creek Formation. This Ediacaran rift-related succession of basalt and associated deep water metasedimentary rocks commonly overlies the Burnie Formation in Tasmania.

The boundary between the Burnie and Eastern Tasmanian (or Tabberabbera) zones is modelled as dipping East at approximately 60°. East from here, the Ordovician to Devonian meta-turbidites rocks (susceptibility less than 0.7×10^{-3} SI, density 2690 to 2710 kgm^{-3}) are interpreted as having been thrust to the west on steeply dipping faults. These faults brought up sections of the underlying ocean-floor basalt. The magnetic feature associated with the fault at approximately 320 km can be traced in the magnetic images for approximately 170 km. Its northern end appears to be truncated by the Selwyn Block boundary, while at its southern end, the magnetic features associated with it are reduced by the presence of the overlying Bass Basin and subsumed in responses from the Tasmanian Dolerite.

Granite is present on all of the many small islands in eastern Bass Strait and is interpreted to sub-crop along most of the eastern end of the section line. Many are modelled as magnetic, with susceptibilities as high as $19x10^{-3}$ SI, and with densities that range from 2620 to 2680 kgm^{-3}. The magnetic character is considered to be from the many dykes in the area (Figures 4-2 and 4-5), which were emplaced from the Jurassic onwards during the breakup of Gondwana (Soesoo et al., 1999; Moore and Wong, 2002). At 323 km along the profile, the dykes were sufficiently numerous that they are modelled as a cluster with a susceptibility of $2.9x10^{-3}$ SI in the granite and $3.6x10^{-3}$ SI in the underlying sedimentary rocks. Small areas of Paleozoic meta-turbidite rocks are also present on Flinders Island and other larger islands. Accordingly, a small area of sub-cropping meta-sedimentary rocks is interpreted at approximately 380 km to account for a 5 mGal gravity anomaly there. However, the gravity data are poor, as the sparse offshore data has been combined with onshore data from an island approximately 20 km away. Thus the reliability of the grid used may be compromised by both merging and interpolation errors and the anomaly may not be real. The gravity response rise at the eastern end of the section, and this requires the addition of a wedge of mantle to have the model conform to the extracted profile. The continental shelf edge is approximately 80 km east of the end of the line, suggesting that this may have some validity. However, the rise in the gravity response may be another artefact, at least part, because of the problems discussed above.

Section 5, Wilsons Promontory

The western end of the section, as far as 60 km, covers the eastern part of the Delamerian Orogen (Figure 4-2 and 4-14). Its eastern boundary is the Bambra Fault, seen in the onshore seismic data to dip to the northwest (Finlayson et al., 1996), and here modelled with an apparent dip of approximately 60° (Figure 4-2), implying a true dip of 70°. A broad north-trending 40 nT magnetic high at approximately 15 km from the western end of the section with an accompanying 5 mGal low is modelled as a non-magnetic granite overlying a mafic magnetic unit with a susceptibility of $19x10^{-3}$ SI beneath it. The eastern boundary of the magnetic package is interpreted to be a southern extension of the Yaramyljup Fault, which separates the higher grade western part of the Glenelg Metamorphic Complex from the lower grade, eastern part (Morand et al., 2004). Elsewhere, the Neoproterozoic-Cambrian sedimentary rocks have modelled susceptibilities less than 1.9x10^{-3} SI and densities of 2690 kgm^{-3}. More mafic sections are modelled with susceptibilities of up to 2.5x10^{-3} SI and densities of from 2710 to 2860 kgm^{-3}. All of the area is covered by Cretaceous and Cenozoic Otway Basin sedimentary rocks, typically 7 km thick (Schneider, 2005) with modelled densities from 2350 to 2590 kgm^{-3}. The rocks are mostly non-magnetic, although a 0.4 km thick horizon is modelled as having a susceptibility of 7.6x10^{-3} SI. Water depths at the western end of the section exceed 1400 m. This suggests significant of crustal thinning during the formation of the Otway Basin and accordingly a wedge of mantle (density 3000 kgm^{-3}) is included at the western end.

As in previously described sections, the western part of the King Island Zone is largely of meta-sedimentary rocks, while the eastern part is of mafic volcanic rocks that are the northern extension of the Grassy Group MORB tholeiites that crop out on the southeast coast of King Island (Meffre et al., 2004). The sedimentary rocks are mostly non-magnetic, although a package at 128 km and another at 230 km are modelled with susceptibilities of 3.8x10^{-3} SI. Modelled densities range between 2630 and 2720 kgm^{-3}. Susceptibilities in the mafic units range from 3.2 to 23x10^{-3} SI and densities from 2640 to 2770 kgm^{-3}. Contacts are sub-vertical in the west, while in the east contacts generally dip to the east at greater than 60°. An exception is the contact at 233 km, which dips to the west at 85°. The section line crosses the northern end of the Braddon River Fault, but it is not modelled as the fault juxtaposes two metasedimentary packages with similar geophysical properties. However, it can be traced to within 25 km of the section line on the magnetic images (Figures 4-2 and 4-5), and is projected to intersect the section 67 km from the western end where there is a 4 mGal dip in the gravity response.

The Mesozoic and Cenozoic cover thins over the King Island Zone, from as much as 7.5 km in the west to less than 1 km north of King Is, 190 km from the western end of the section. This is generally non-magnetic, with densities that ranged from 2250 to 2590 kgm^{-3}. Minor thin layers of mafic volcanic rocks are present, with susceptibilities that range from 8.2 to 130x10^{-3} SI.

The Rocky Cape Zone extends from 246 to 323 km along the section line. Both the western boundary, with the King Island Zone and the eastern boundary, with the Burnie Zone, have dips of 85° to the east. The Smithton Basin is the upper layer of the zone and is restricted to a relatively small area between 276 and 303 km. Its internal and external contacts are modelled with shallow dips of 30° or less. Only two units are recognised, an upper unit with susceptibility of 2.5×10^{-3} SI and a density of 2640 kgm^{-3}

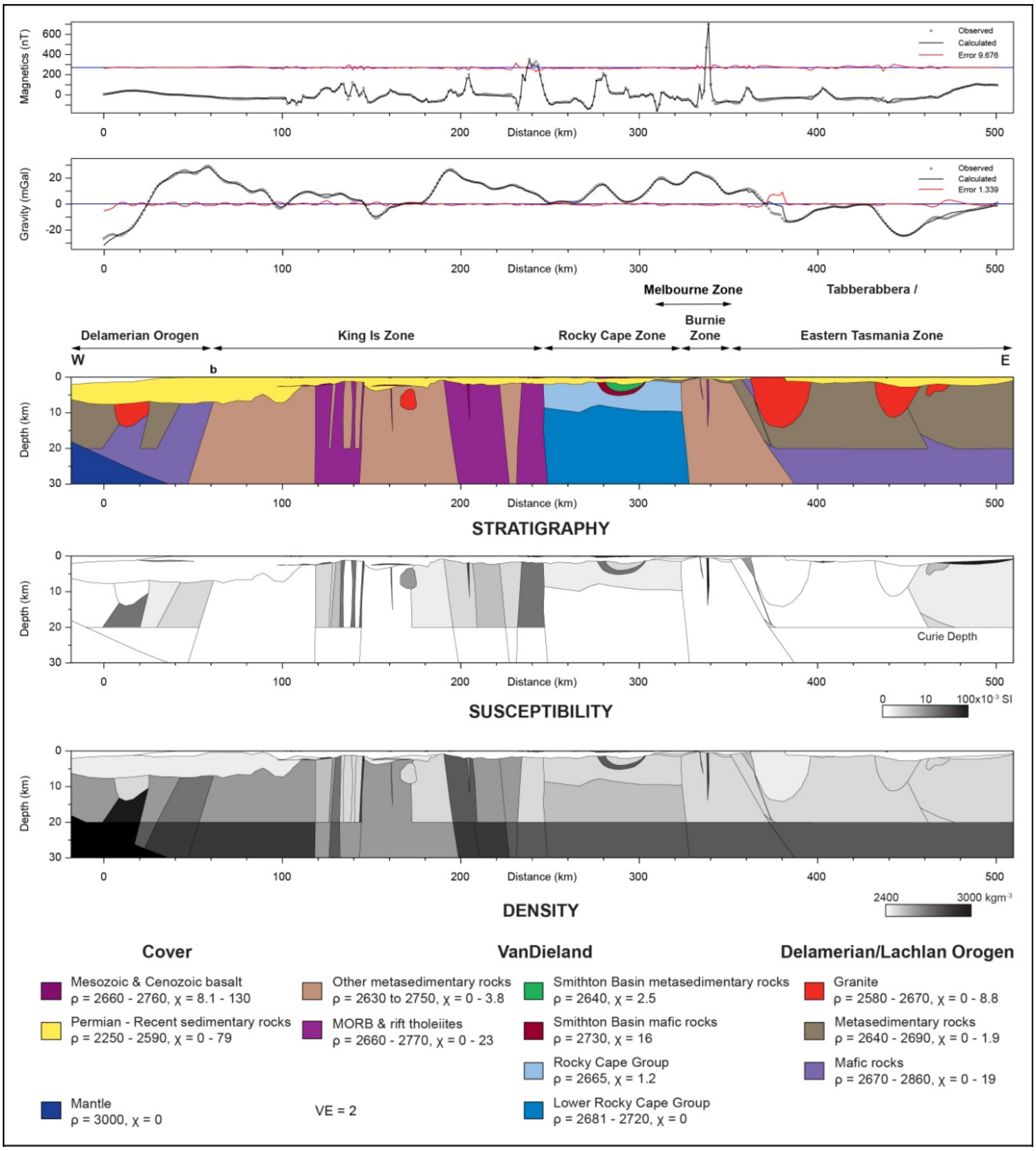


Figure 4-14. Section 5, Wilsons Promontory. Location shown in Figure 4-2. Susceptibilities as 10^{-3} SI units. b marks the Bambra Fault, here the boundary between the Delamerian Orogen and VanDieland.

and a lower unit, the mafic Spinks Creek Volcanics, with a susceptibility of 16×10^{-3} SI and a density of 2730 kgm^{-3}. The Rocky Cape Group, below the Smithton Basin, is modelled with a susceptibility of 1.3×10^{-3} SI and a density of 2665 kgm^{-3}. This is the highest density modelled for the Rocky Cape Group and, together with the magnetic response, suggested an increased proportion of pyrrhotite-rich pelite was present, as susceptibilities measured in northwestern Tasmania returned higher values from fine grained rocks than the coarser quartzites (see Section 1, above). The non-magnetic Lower Rocky Cape unit is modelled with densities ranging from 2681 kgm^{-3} above 20 km depth to 2720 kgm^{-3} below this.

A discontinuous 0.6 km thick layer of Melbourne Zone rocks is modelled from 307 km to 350 km in the section to accord with the mapped geology in southern Victoria, where rare inliers of the Selwyn Block are present (VandenBerg et al., 2000; Welsh et al., 2011). The model uses a non-magnetic sedimentary sequence with a density of 2640 kgm^{-3}, which compares with median densities for Melbourne Zone sedimentary rocks of 2636 kgm^{-3} for mudstone (N=1), 2707 kgm^{-3} for shale (N=1), 2459 kgm^{-3} for siltstone (N=8), and 2581 kgm^{-3} for sandstone (N=23) (Skladzien, 2007). The Melbourne Zone is interpreted to continue east to the eastern edge of the Burnie Zone. A layer of Cretaceous and Cenozoic sedimentary rocks up to 2 km thick (density 2250 kg^{-3}) is interpreted to overlie both the Rocky Cape and Burnie Zones, and this in turn, is overlain by a thin layer of Eocene basalt (Price et al., 2003) with modelled susceptibilities that range from 8.5×10^{-3} to 130×10^{-3} SI. Basalt also extends above the western Burnie Zone.

The Burnie Zone is modelled as non-magnetic, with a density of 2670 kgm^{-3} above 20 km and 2720 kgm^{-3} below this. However, a magnetic anomaly is present at 334 km that lay 25 km directly along strike from the pre-Tyennan Orogeny tholeiitic basalt and meta-gabbro present at Cape Liptrap (VandenBerg et al., 2000) and a second was present at 338 km that appears to be of similar character, but is larger and does not crop out. The anomaly at 334 km is modelled with a source having an east dip of 75° and a susceptibility of 80×10^{-3} SI and a density of 2780 kgm^{-3}, while the larger anomaly at 338 km is modelled as having a vertical source with a susceptibility of 90×10^{-3} SI and a density of 2750 kgm^{-3}. The outcropping meta-gabbro has a median measured susceptibility of 0.6×10^{-3} SI, insufficient to explain the anomalies, but a Tasmanian ?Neoproterozoic rift-related tholeiitic basalt that is geochemically similar to the basalt seen on Cape Liptrap (Woof, 2006) has a median response of 22×10^{-3} SI, (N=235) with values as high as 336×10^{-3} SI. It is possible that the anomalies at 334 and 338 km may be from the basaltic units present at Cape Liptrap. If so, they may be more or less coeval with the basalts of the Ediacaran Crimson Creek Formation.

The boundary between the Burnie Zone and the Tabberabbera Zone, at 355 km along the profile, dips ~40° to the east. As in the previous sections, the base of the Tabberabbera Zone is modelled with a west-directed thrust against the VanDieland craton. However, unlike other sections, a second west-dipping magnetic unit is modelled approximately 10 km in the hanging wall of the contact. It is correlated with the magnetic feature at 320 km on Section 4, the King Island section, as another mafic package on a west-verging thrust fault inside the Tabberabbera Zone that is truncated by the Burnie-Tabberabbera Zone boundary approximately 30 km north of the section line. The mafic package is modelled with a susceptibility of 7.5×10^{-3} SI and a density of 2680 kgm^{-3}. The model also has another east-dipping fault at 455 km that has a package of magnetic mafic rocks (susceptibility 5.7×10^{-3} SI) in its hanging wall. These separate weakly magnetic Tabberabbera Zone sedimentary rocks in the east (susceptibility 1.3×10^{-3} SI) from non-magnetic sedimentary rocks to the west. Both sedimentary packages have a modelled density of 2665 kgm^{-3}.

Granite dominates the near-surface Paleozoic rocks in the eastern end of the section. It crops out on Wilsons Promontory and the nearby islands and is interpreted to have produced the gravity lows seen in the region. One is modelled as magnetic, with a susceptibility of 6.9×10^{-3} SI. Others were non-magnetic. Modelled densities range from 2580 kgm^{-3} for the Mt Norgate Granite on Wilsons Promontory to 2620 kgm^{-3}. The 8 mGal mismatch between the gravity model and the grid from 370 to 380 km corresponds to Wilsons Promontory, suggesting a problem in stitching the onshore and offshore data and that the modelled density for the Mt Norgate Granite might be quite different to the actual density.

A thin (less than 2.7 km thick) Gippsland Basin sequence overlies the Paleozoic rocks. Diffuse magnetic responses were present from 399 to 408 km and east of 468 km that are interpreted as a magnetic sedimentary layer at the base of the Gippsland Basin. Both the Strzelecki Group and the Golden Beach Subgroup in the Gippsland Basin are known to have a volcanic component (O'Halloran and Johnstone, 2001; Duddy, 2003) and it is possible that the magnetic layer is part of one or other of these units. Both units are modelled as having relatively high susceptibilities (58 and 79×10^{-3} SI) but low densities (2540 and2580 kgm^{-3}), more consistent with a partly lithified sedimentary rock than a volcanic flow. Non-magnetic sedimentary rocks (density 2300 kgm^{-3}) are modelled above the Tabberabbera Zone. A minor near-surface flow (susceptibility 130×10^{-3} SI, density 2750 kgm^{-3}) is interpreted at 438 km.

Section 6, Phillip Island

This section was south of almost all of the outcropping Paleozoic rocks in Victoria and was approximately 50 km south of a forward modelled profile by McLean et al. (2010). Section 6 (Figure 4-15) extends from the southern parts of the Stawell Zone in the west to the southern Tabberabbera Zone in the east. The extracted section was 'pinned' to the model value at 8 km east of the western end of the extracted profile, as this end of the extracted profile was approximately 28 nT below the expected value for a uniform background with no influence from magnetic bodies. This meant that the final model is based on a profile higher than that extracted from the gridded data.

The Cambrian meta-turbidite rocks of the Stawell Zone are modelled in the far west of the section. They are non-magnetic with a density of 2686 kgm^{-3}. However, much of the magnetic modelling in the region was disrupted by the presence of at least one weakly magnetic near-surface unit in overlying the Otway Basin with modelled susceptibilities of up to 31×10^{-3} SI. The eastern boundary of the Stawell Zone, the Avoca Fault, is extrapolated approximately 70 km south along a 5 mGal gravity high. This anomaly is interpreted to be sourced from highly strained Cambrian mafic volcanic rocks observed in a mine dump and drill core (Taylor et al., 1996). The Avoca Fault is interpreted to intersect Section 6 at ~35 km where it has a modelled dip of 70° to the west. The basalt and sedimentary rocks in the hanging wall of Avoca Fault are modelled as magnetic (29×10^{-3} SI) and dense (2730 kgm^{-3}).

A non-magnetic wedge of Ordovician Bendigo Zone meta-turbidites is modelled from 35 to 45 km along the section; the modelled density was 2670 kgm^{-3}. The eastern boundary is the Bambra Fault, a complex fault zone with fault segments that dip to the northwest and southeast (Finlayson et al., 1996). Modelling indicates an apparent dip of 30° for the northwestern segment, which accords with the 50° true dip interpreted by Finlayson et al. (1996). Section 6 indicates the Bambra Fault truncates the Avoca Fault at a depth of approximately 15 km. The southeastern fault segment has not been modelled. Figure 4-4 and the model suggest that movement on the northwest-dipping segment of the Bambra Fault controlled the present thickness of the Otway Basin by normal movement allowing deposition in the Late Jurassic-Early Cretaceous and by subsequent inversion at approximately 90 Ma (Hill et al., 1995).

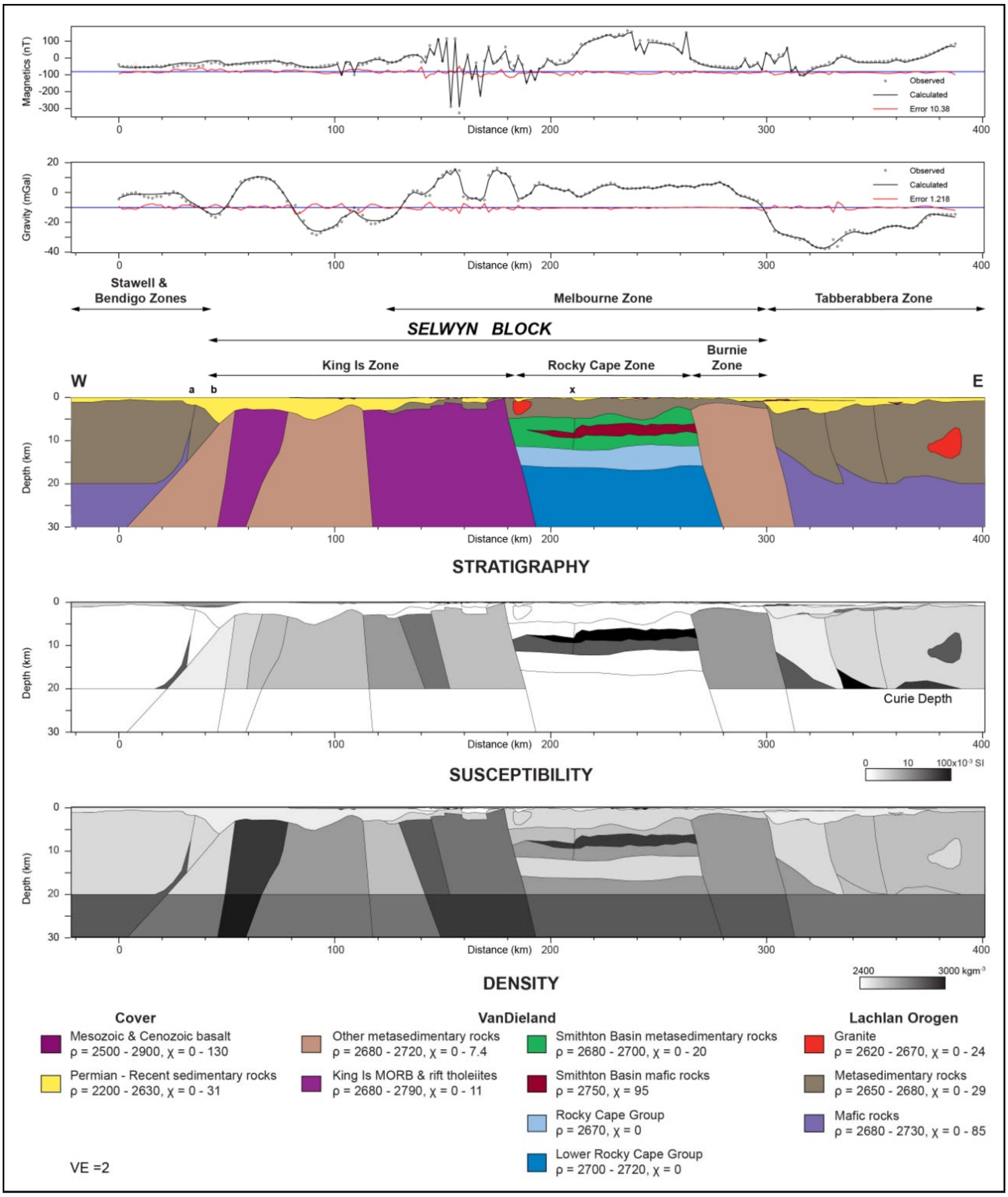


Figure 4-15. Section 6, Phillip Island. Location shown in Figure 4-2. Susceptibilities shown as x10^{-3} SI. a marks the Avoca Fault, b, the Bambra Fault and x the possible continuation south of an unnamed fault that displaces the Heathcote-Mt William Fault. See also Figure 4-6.

The seismic interpretation by Finlayson et al. (1996) indicated that the overlying Barwon Downs Graben, part of the Otway Basin, had a depth of 3.8 s TWT, or approximately 6 km. This has been used to constrain this part of the section and requires a density of 2500 to 2540 kgm^{-3} for this sedimentary package. The Otway Basin is interpreted to be at least 1 km thick from 29 km to 144 km along the section, with thickest part through the northern end of the Torquay Embayment, at 95 km (Madsen, 2002).

The King Island Zone extends from 45 km to 180 km along the section. As with previous sections, the boundaries between the mafic and metasedimentary units dip to the west in the west and to the east in the east. Dips in the west are typically from 60° to 75°, while those in the east range from 50° to 70°. The westernmost mafic unit can be traced approximately 35 km north from the section to the Ceres Gabbro, described by Morand (1995) as having been metamorphosed to amphibolite facies. The median susceptibility of the Ceres Gabbro is $0.28x10^{-3}$ SI (N=165), compared to the modelled susceptibility of from 3.7 to $4.4x10^{-3}$ SI. Weathering of the outcrop may explain at least part of this discrepancy. Near the eastern edge of the zone, at 177 km, the section line passes 4 km north of a small outcrop of MORB tholeiite (Henry and Birch, 1992) with a median measured susceptibility of $6.1x10^{-3}$ SI (N=92), compared to a modelled susceptibility of $3.1x10^{-3}$ SI. The modest discrepancy may be due to the presence of the overlying Eocene basalt (Price et al., 2014) that hampered effective modelling. Henry and Birch (1992) compared the MORB tholeiite to the exposed Cambrian 'greenstones' elsewhere in Victoria, but images of the magnetic and gravity data (Figures 4-2 to 4-5) suggest a stronger correlation with the 575 Ma MORB tholeiites on King Island (Calver et al., 2004; Meffre et al., 2004). Elsewhere, interpreted susceptibilities in the mafic units range from 3.7 to $11x10^{-3}$ SI where modelled as above the Curie Depth. Modelled densities in the same region range from 2680 to 2764 kgm^{-3}. Meta-sedimentary units have modelled susceptibilities of 1.9 to $4.1x10^{-3}$ SI. Ordovician Melbourne Zone chert and thin-bedded turbidite rocks are present less than 15 km north of the section line in the eastern part of the King Island Zone (VandenBerg et al., 2000), and these are included as a layer of non-magnetic rocks with densities of 2650 to 2660 kgm^{-3}. Much of the eastern end of the zone was also covered by less than 0.2 km of Eocene basalt (Price et al., 2014) and these are modelled with susceptibilities to $130x10^{-3}$ SI and densities from 2620 to 2700 kgm^{-3}. One unit also is modelled with remanence of 1.3 Am^{-1}.

The Rocky Cape Zone extends from 180 km to 265 km along the section. Both boundaries are modelled as dipping east at approximately 60°. The zone is marked by a 200 nT high, the western side of which is truncated by a fault at 210 km. This truncation can be traced northwest to where it crosses the Heathcote Fault Zone (Figure 4-6). Here, the Heathcote Fault Zone appears to have been cut by a fault with an apparent displacement of approximately 1 km, and foliations in the mafic volcanic rocks in the footwall of the displaced region show a consistent 70° to 80° southwestern dip (Gray and Willman, 1991). Although this displacement is no older than Middle Devonian as it displaced the Melbourne Zone (which was only deformed in the Middle Devonian Tabberabberan Orogeny), it may indicate reactivation of an older structure.

McLean et al. (2010) modelled a parallel section over the same magnetic high but 55 km north of Section 6. Their section suggested that the causative body was an ultramafic unit (susceptibility 40×10^{-3} SI, density 2800 kgm^{-3}) with a top at a depth of 7 km and a base at 15 km and was over 70 km wide. They correlated the ultramafic body with those seen in the Burnie Zone in Tasmania. Paleozoic Melbourne Zone rocks overlay this. Section 6 suggests an alternative possibility, with the Melbourne Zone approximately 4 km thick and underlain by an 8 km thick non-magnetic sequence of the Smithton Basin. The Spinks Creek Volcanics are modelled as magnetic (susceptibility 95×10^{-3} SI, density 2750 kgm^{-3}) and form a layer up to 3 km thick that extends from 191 to 266 km along the section. The top is at a depth of 6 km. Below this lies another layer of Smithton Basin rocks with a susceptibility of 20×10^{-3} SI and a density of 2700 kgm^{-3}. The Rocky Cape Group (non-magnetic, density 2670 kgm^{-3}) and the Lower Rocky Cape unit (non-magnetic, density 2700 to 2720 kgm^{-3}) underlie this package. The Woolamai Granite (non-magnetic, density 2620 kgm^{-3}) is intruded near the western edge of the Zone. Thin layers of Mesozoic and Cenozoic sedimentary rocks (non-magnetic, density 2270 to 2580 kgm-3) and basalt (susceptibility 0 to 130×10^{-3} SI, density 2500 to 2900 kgm^{-3}) cover much of the region, but a small outcrop of Melbourne Zone rocks is present 4 km south of the section 217 km from its western end.

The Burnie Zone extends from 265 km to 300 km. It is modelled with a susceptibility of 7.4×10^{-3} SI and a density of 2700 kgm^{-3}, rising to 2720 kgm^{-3} at depth. The Melbourne Zone is modelled as continuing across the Burnie Zone. Eocene basaltic volcanic rocks (Price et al., 2014) (susceptibility 57×10^{-3} SI, density 2590 kgm^{-3}) are modelled as overlying the boundary between the Melbourne Zone and the Tabberabbera Zone. Mesozoic and Cenozoic sedimentary rocks (susceptibility 0 to 31×10^{-3} SI, density 2400 to 2450 kgm^{-3}) also cover the entire area; the thickness increases from less than 0.2 km in the west to 0.8 km in the east.

The Tabberabbera Zone extends from 265 km to the end of the section. Following mapping further north (VandenBerg et al., 2000), it is interpreted as largely comprised of Paleozoic turbidite rocks on a basement of mafic volcanic rocks. Unlike most of the previous sections, the sedimentary rocks are modelled as being weakly magnetic, with susceptibilities ranging from 1.7×10^{-3} to 2.9×10^{-3} SI. The boundaries between the areas with different susceptibilities are interpreted to dip east at approximately 65°. Densities range from 2670 to 2680 kgm^{-3}. Furthermore, the model suggests that the mafic volcanic basement is both slightly closer to the surface and more magnetic, with susceptibilities from 21×10^{-3} to 85×10^{-3} SI. A magnetic granite (susceptibility 24×10^{-3} SI, density 2670 kgm^{-3}) is interpreted at the eastern end of the section to account for the 100 nT high at this location.

Cretaceous to Recent Gippsland Basin sedimentary and volcanic rocks overlie the Paleozoic sequence. The sedimentary rocks have modelled susceptibilities ranging from 0 to 11×10^{-3} SI and densities from 2400 to 2630 kgm^{-3}. These relatively high susceptibilities may be due to the mineralogical immaturity of the Strzelecki Group, the lowest unit in the Gippsland Basin. The sequence thins to the east, with the thickest part, 3.7 km, at approximately 325 km along the section whereas the package is approximately 1 km thick at the eastern end. Volcanic units interlayered with the sedimentary rocks have susceptibilities that range from 0 to 57×10^{-3} SI, and densities between 2570 and 2590 kgm^{-3}.

Section 7, Northwest-southeast

The gravity data on this section are notably poor, with the data from only five ship traverses in the data set available between 95 km and 360 km along the section line. Consequently, forward models of the gravity data for the southeastern part of the King Island Zone and all of the Rocky Cape Zone are less reliable than elsewhere. The section was ‘pinned’ 36 km southeast of the northwestern end of the extracted profile, as this end of the extracted profile was approximately 37 nT below the expected value for a uniform background with no influence from magnetic bodies. The line traversed at least 4 different magnetic surveys that have been stitched together and this may have caused part of the problem.

The Bendigo Zone of the Lachlan Orogen forms the basement at the northern end of the section (Figure 4-16). Outcrops to the north are of non-magnetic Ordovician turbidite rocks (density 2680 kgm^{-3}) that have been metamorphosed to epizone to anchizone levels (Wilson et al., 2009). Seismic surveys further north indicated that deeper than 15-20 km, the section has a significant mafic volcanic component (susceptibility 3.4 to 26×10^{-3} SI, density 2730 to 2800 kgm^{-3}) (Cayley et al., 2011). These mafic rocks are also present in the hangingwalls of the major thrust faults, and the model suggests that this has happened under the onshore Otway Basin. The southern boundary of the Lachlan Orogen sequence is the Bambra Fault, 20 km from the northern end of the section. It has an overall dip of approximately 50° to the northwest, as imaged in seismic sections by Finlayson et al. (1996).

The Otway Basin overlies the Lachlan Orogen rocks, with displacements on the Bambra Fault indicating that it was the headwall of a half graben during the early formation of the Otway Basin and was active during inversion at 90 Ma (Hill et al., 1995; Cooper and Hill, 1997). However, modelling near the Bambra Fault is compromised by an apparent mismatch between the onshore and offshore data sets of approximately 3 mGal. The modelled thickness of the basin is approximately 6 km, assuming a density of 2530 kgm^{-3} for the lower parts and 2350 kgm^{-3} for the upper parts. Pliocene to Recent basaltic flows overlie much of the Otway Basin. Here they are modelled as mostly with susceptibilities of 10×10^{-3} SI but with minor values between 1.2 and 14×10^{-3} SI where particular flows are crossed. The basalts have modelled densities of 2750 to 2800 kgm^{-3}.

The King Island Zone extends from south of the Bambra Fault to 190 km along the section. As in previous sections, the zone comprises alternating units of mafic and sedimentary rocks, with sedimentary rocks more often in the northwest and the mafic rocks more common in the southeast. Susceptibilities in the mafic rocks range up to 30×10^{-3} SI. Densities range from 2680 to 2770 kgm^{-3}. The mafic rocks at the southeastern end of the zone are abnormally weakly magnetic when compared to the previous sections, with responses mostly less than 7x10^{-3} SI. This is because the section line runs along a fault that was the site of local alteration of the rocks. The sedimentary

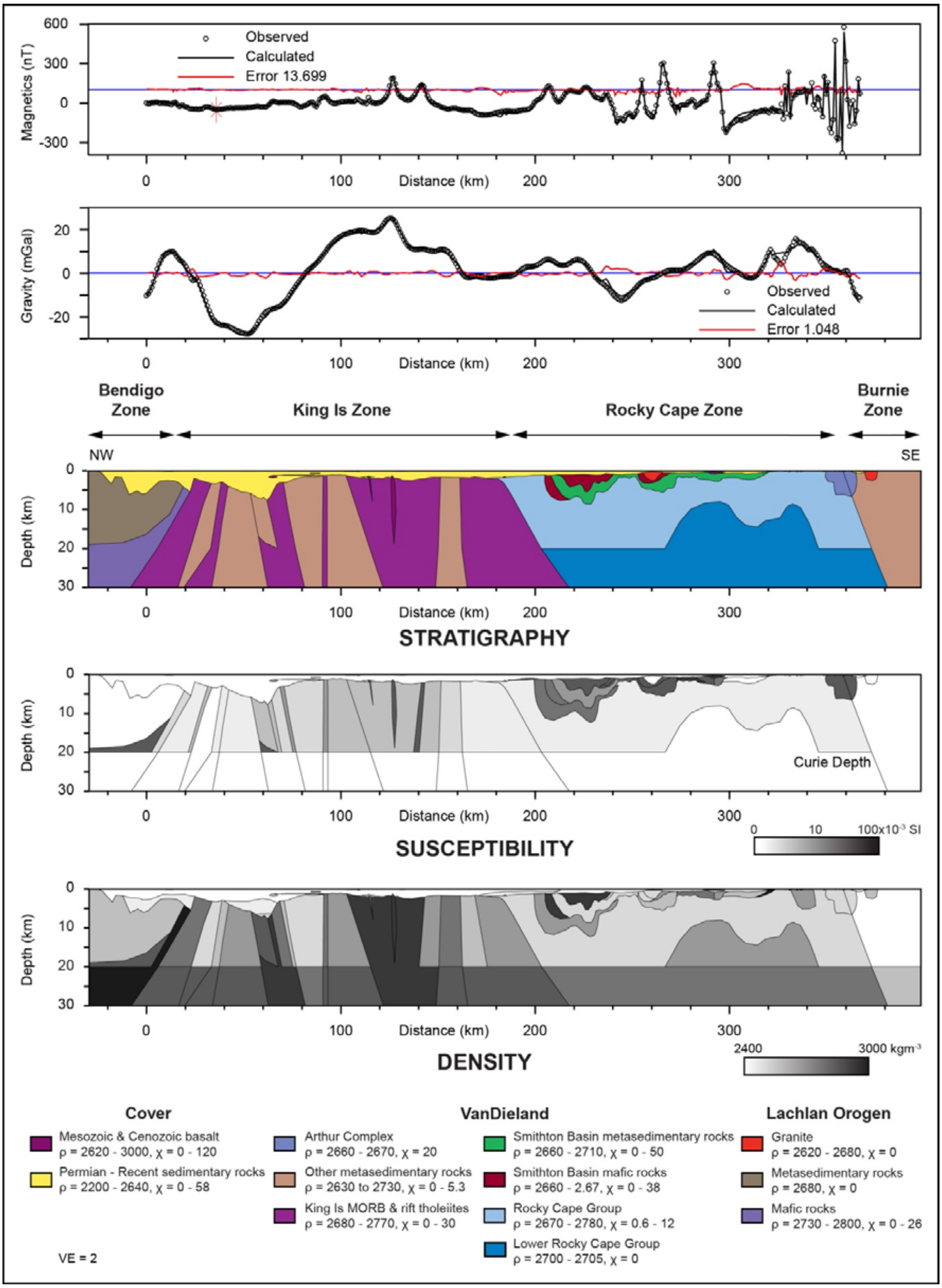


Figure 4-16. Section 7, northwest-southeast. Location shown in Figure 4-2. Susceptibilities as 10^{-3} SI units.

rocks are less magnetic, with the maximum susceptibility 5.3×10^{-3} SI. Modelled densities range from 2630 to 2740 kgm^{-3}. Apparent dips are variable but steep, with both northwest and southeast apparent dips modelled.

The Cretaceous Torquay Embayment lies above the King Island Zone. Seismic interpretation by Madsen (2002) provided the basin-basement surface that is used in the model. This suggests that the maximum depth of the sub-basin along the section line is 6.2 km deep, just north of the Snail Fault that lay at 65 km along the section line. Like the Bambra Fault, the Snail Fault acted as a the headwall fault in a half graben that was initially filled with Upper Jurassic to Barremian sediments (modelled density 2640 kgm^{-3}) and then by Aptian to Albian sediments (density 2550 kgm^{-3}) and then overlain by Cenozoic sedimentary rocks (density 2300 to 2350 kgm^{-3}) that extend over the entire King Island Zone (Hill et al., 1995). Minor Cretaceous and Cenozoic basaltic dykes and flows were also modelled. Susceptibilities range from 3.7 to 49×10^{-3} SI and densities from 2660 to 2770 kgm^{-3}. One dyke is remanently magnetised (1.5 Am^{-1}).

The Rocky Cape Zone extends from the boundary with the King Island Zone, at 190 km, to 365 km. This boundary is modelled as having an apparent dip of approximately 40°. As in previous models, the zone is divided into three, with the upper layer the Neoproterozoic Smithton Basin, the middle layer the Mesoproterozoic Rocky Cape Group and the lower layer the Paleoproterozoic ‘Lower Rocky Cape Group’. The Smithton Basin is modelled with an upper layer of magnetic volcanic rocks, the Spinks Creek Volcanics, a middle layer of sedimentary rocks and, in one area, a lower layer of mafic igneous rocks. The Spinks Creek Volcanics are present in the axes of synclines, where dips are mostly less than 45°. Susceptibilities ranged from 17 to 38×10^{-3} SI and densities from 2640 to 2760 kgm^{-3}. The sedimentary layer is variably magnetic (susceptibility 0 to 50×10^{-3} SI) and with densities ranging from 2675 to 2710 kgm^{-3}. The lowest layer is only present at approximately 207 km from the northern end, where it has a modelled susceptibility of 18×10^{-3} SI and a density of 2760 kgm^{-3}. The unit may be comparable to the mafic rocks outlined by Everard et al. (2007) that are present near the base of the Smithton Basin.

Non-magnetic Devonian granite (density 2620 kgm^{-3}) intruded the Smithton Basin from 255 to 263 km. Sections 3 and 7 intersect near the southern edge of the granite, at

265 km. The two sections do not compare well, as Section 3 shows approximately 5 km of granite lying above 5 km of Smithton Basin sedimentary rocks with a susceptibility of 2.5×10^{-3} SI, while Section 7 has 0.5 km of granite above 1.2 km of Spinks Creek Volcanics (susceptibility 38×10^{-3} SI), which in turn lie above 5.5 km of Smithton Basin sedimentary rocks with susceptibility 22×10^{-3} SI. However, the granite contact on Section 3 is modelled as sub-vertical and so only a small difference between the plotted path and the extracted path would make a significant difference in the model. As well, where the two sections cross, Section 3 is modelled as having folded Spinks Creek Volcanics either side of the intersection point, with a syncline to the west having a susceptibility of 130×10^{-3} SI and a syncline to the east with a susceptibility of 24×10^{-3} SI and the anticline between the two truncated by the granite. Any minor variation between the extracted path and the plotted path would result in significant differences in the modelled sections. The Rocky Cape Group is interpreted to have more variation on this section than elsewhere. The variations in the densities modelled probably result from the poor gravity coverage noted above. Variations in the susceptibilities are at least partly due to orientation of the section, sub-parallel to the strike of the rocks, particularly between 220 and 300 km. This means that there are many responses on the section that were due to edge effects of magnetic bodies that are either not on the section line or inadequately represented on it. Where exposed on the northwestern tip of Tasmania, the Rocky Cape Group was found to have a wider range of susceptibilities (from 0.02 to 0.2×10^{-3} SI) than quartzite units elsewhere, and one of the siltstones had values as high as 0.38×10^{-3} SI. These compare with overall medians of 0.04×10^{-3} SI for the quartzite units (N=138) and 0.14×10^{-3} SI (N=117) for the siltstones. In the model, susceptibilities range from 0.6 to 12×10^{-3} SI, and densities from 2670 to 2780 kgm^{-3}. The boundaries between the Rocky Cape layers are mostly gently dipping, although those at approximately 200 km along the section line are sub-vertical. The Lower Rocky Cape Group is modelled as non-magnetic, with a density of from 2700 to 2705 kgm^{-3}.

Northwest of 325 km, the rocks of the Rocky Cape Zone are covered by a layer of Cretaceous to Recent sedimentary rocks (density mostly 2200 to 2380 kgm^{-3}) that generally thicken northwards to a maximum depth of 1.8 km near the boundary with the King Island Zone. One 0.4 km thick sedimentary layer that extends from 185 to 200 km has a modelled susceptibility of 58×10^{-3} SI and a density of 2590 kgm^{-3}. It may be of volcaniclastic material as it lies adjacent to a minor ?Cretaceous mafic flow. Other Cretaceous and Cenozoic mafic volcanic rocks are also present, with susceptibilities ranging from 0 to 85x10^{-3} SI and densities from 2620 to 3000 kgm^{-3}. Some volcanic rocks were remanently magnetised, with values ranging from 0.15 to 3.7 Am^{-1}. The misfit in the section between 298 km and 316 km may be caused by another remanently magnetised Cenozoic volcanic rock package to the south of the Smithton Basin that could not be imaged in a deep low. As suggested above, another possible cause may result from difficulties in stitching the onshore and offshore data sets.

The Arthur Metamorphic Complex is modelled as lying from 350 km to 365 km and extending to 6.5 km deep, with a suggested susceptibility of 20x10^{-3} SI and density of 2660 to 2670 kgm^{-3}. Modelling of the shape is difficult as most of the Complex was covered by Tertiary basalt with modelled susceptibilities of from 74 to 120x10^{-3} SI. These appear to have at least in part flowed down paleovalleys, with an unknown thickness of Cenozoic river deposits in them. Carboniferous-Permian diamictite also partly covers the basement. This means that neither the magnetic or gravity data could be relied on to effectively model the shape of the Arthur Metamorphic Complex. Below the Arthur Metamorphic Complex, the model suggested an apparent dip of the contact between the Rocky Cape and Burnie zones of approximately 50°.

The southeastern end of the section covers the Burnie Zone, a non-magnetic package with a modelled density of 2650 kgm^{-3}. Past the end of the modelled data points, the rocks have been intruded by Devonian granite with a modelled density of 2680 kgm^{-3}. Both are covered by the same Carboniferous-Permian diamictite and/or Tertiary basalt.

4.4 DISCUSSION

4.4.1 VICTORIAN OUTCROPS

There have been many difficulties in correlating the Proterozoic and Early Paleozoic geology of Tasmania with that seen in central Victoria. There are only a few relevant outcrops in Victoria, as most of the region is covered by the Ordovician to Middle Devonian rocks of the Melbourne Zone and the Cretaceous and Tertiary Gippsland and Otway basins, and many critical areas are under Bass Strait. In the past, there have been attempts to correlate the Tasmanian Mt Read Volcanics with the Stavely Volcanic

Group in western Victoria (e.g. Shaw et al., 1996; Foster et al., 2005) or the Mathinna Group with the Melbourne Zone (Powell and Baillie, 1992). Recognition of the Selwyn Block (Cayley et al., 2002) has provided a new opportunity to re-examine some of the potential correlations between Tasmania and Victoria.

The high-resolution magnetic data available has made this extrapolation a realistic possibility by providing a single framework over the region into which other disparate and sometimes cryptic observations can be placed. Some units in the Tasmanian parts of VanDieland, such as the Spinks Creek Volcanics and the Grassy Group, have distinctive magnetic and/or gravity responses that can be recognised as continuing north under Bass Strait and into Victoria. These responses from the pre-Mesozoic basement were recognised even in the deepest parts of the Bass Basin.

Even so, some difficulties remain. Responses from the Tasmanian Dolerite and Cretaceous to Recent mafic volcanic rocks partially or completely obscure the magnetic responses of the older packages. Faults of all ages have displaced the basement and the associated geophysical responses. Those with significant (apparent) displacement added uncertainty to the interpretation. Because of this, correlation of individual units across Bass Strait was not possible, even where the units extend into Victoria. Rather, the zones established in Tasmania have been extrapolated north, giving the first broad, continuous pre-Carboniferous strato-tectonic map of this part of VanDieland. Together with the interpretation outlined in Chapter 3, the zones provide a complete map of the internal structure of the micro-continent. Furthermore, by defining the edges of the micro-continent, the results allow for a better understanding of the relationships between this exotic terrane and the enclosing Lachlan Orogen.

The interpretation results (Figures 4-5 and 4-7) suggest that the three western zones in Tasmania (the Burnie, Rocky Cape and King Island zones) lie beneath the Bass Strait and their extensions comprise the Selwyn Block. In Tasmania, the Burnie Zone is characterised by ?710 Ma turbidite rocks (Burnie Formation) that are successively overlain by 580 Ma rift-related basalt and associated sedimentary rocks (Crimson Creek Formation), by 515 Ma mafic-ultramafic rocks, by 496-508 Ma Mt Read Volcanics (Mortensen et al., 2015) and finally by Owen Conglomerate (Chapter 3). In Victoria, the Cambrian andesites and rhyolites of the Jamieson Volcanic Group, exposed in the

Jamieson, Licola and Glen Creek windows (Figure 4-5) through the eastern Melbourne Zone (VandenBerg et al., 1995) are geochemically similar to, and the same age as the Mt Read Volcanics (Crawford et al., 2003a). Like the Mt Read Volcanics they host copper-gold mineralisation (Morey et al., 2002). These windows are interpreted to be the northern extension to the Burnie Zone. In Tasmania, no outcrops of Mount Read Volcanics are known in the Rocky Cape Zone. All are either associated with the Burnie Zone or lie further to the east. The geophysical interpretation suggests these eastern zones in Tasmania do not extend across Bass Strait and so the Jamieson Volcanic Group may indicate the presence of the Burnie Zone at depth.

Crawford and Berry (1992) studied the mafic-ultramafic rocks in Tasmania, most of which occur in the Burnie Zone. Subsequently, Crawford (in Goss and Lenard, 2006) noted the similarity between the serpentinised cumulate harzburgite intersected in hole FOS 1, drilled 30 km north of Cape Liptrap in Victoria (Figure 4-5), with those seen in the Burnie Zone in Tasmania. Ultramafic rocks are also present in the Dolodrook Window (Figure 4-5) in the eastern Selwyn Block and, although the rocks are extremely altered, they have similar primary mineralogy and Cr/Al ratios to some ultramafic rocks in Tasmania and are also associated with basaltic lavas (Spaggiari et al., 2003a). Ultramafic rocks are also present in the Glen Creek Window (VandenBerg et al., 2006) (Figure 4-5), but correlation with occurrences in Tasmania is more tenuous. Pillow basalt of the Maitland Beach Volcanics and the Corduroy Creek Metagabbro crop out at Cape Liptrap (VandenBerg et al., 2000), and clasts of both are found in the Pragian Boola Formation in the southeastern Melbourne Zone (Figure 4-5) (VandenBerg et al., 2006). While neither lithology is unique to the Burnie Zone in Tasmania, basalt is present in the Crimson Creek Formation and both are present in some mafic-ultramafic complexes (Crawford and Berry, 1992). Cape Liptrap is interpreted to be the northern continuation of the Burnie Zone. The Boola Formation mostly overlies the northern continuation of the Burnie Zone, but partly overlies the eastern edge of the Rocky Cape Zone. It is suggested these clasts were eroded from the exposed Burnie Zone during the Bindian Orogeny. Parts of Boola Formation also contain clasts of quartzite (VandenBerg et al., 2006), a characteristic lithology of the Rocky Cape Group in Tasmania (Halpin et al., 2014) but unknown in Victoria prior to the Upper Devonian.

Henry and Birch (1992) described a small basalt inlier on Phillip Island (in the south central part of the Selwyn Block) to which they tentatively ascribed a Cambrian age (Figure 4-5). Both the magnetic and gravity data suggest that the inlier can be correlated south along strike with the 580 Ma rift-related mafic volcanic rocks on southeastern King Island (Chapter 2). In Chapter 3, these mafic volcanic rocks were used to define the eastern edge of the King Island Zone in Tasmania. There may also be instances where the overlying Oligocene Older Volcanic Province has sampled clasts from the underlying 580 Ma basement. Unusual megacrysts of amphibole up to 7 cm long are present in the Oligocene Older Volcanic Province basalts approximately 13 km west of the Phillip Island occurrence, at West Point (Simmons et al., 2014), and 13 km northwest, at Point Leo (Scholte, 2010). If this has taken place, it suggests that the Oligocene basalts have sampled from Proterozoic layer 3 oceanic gabbros, lithologies that are not exposed on either King Island or Phillip Island, but that may reasonably have come from deeper parts of the rift-related mafic sequences present on King Island.

In the western Selwyn Block, Morand (1995) described the Ceres Gabbro as having been metamorphosed to amphibolite facies and shortly after having been subjected to north-south shortening. He suggested it might represent layer 3 of oceanic crust. Woof (2006) recorded the presence of cumulate texture in samples of the gabbro. The magnetic data suggest that the Ceres Gabbro can be traced along strike to a magnetic package west of King Island. The mafic rocks of the King Island Zone are interpreted to have formed as a result of extension during the breakup of Rodinia, between 760 and 580 Ma (Chapter 3). If the Ceres Gabbro is indeed a correlative of one of the mafic packages seen on King Island, it represents the northernmost exposure of the geology of the King Island Zone.

4.4.2 CENTRAL VICTORIAN SEISMIC LINE

Interpretation of data from the Geoscience Australia 2006 central Victorian seismic survey lines 06GA-V3 and V4 (hereafter referred to as Lines 3 and 4, Figure 4-2) suggested that under the Melbourne Zone, the Selwyn Block had three layers (Cayley et al., 2011). The top layer was thin and discontinuous, and shallowed to the northeast, but elsewhere the base of this layer occurred at a depth of approximately 15 km. It had a maximum thickness of approximately 3 km and a modelled density of 2780 kgm^{-3} and was correlated with the andesite-rhyolite-volcanic breccia unit of the Jamieson Volcanic Group. This unit cropped out to the southeast on basement highs of the Selwyn Block, and was inferred to be present on the basement high imaged at the northeastern end of the seismic line. Furthermore, Rossiter and Gray (2008) had suggested that the $^{87}Sr/^{86}Sr$ ratios and barium contents of the central Victorian Upper Devonian granites implied that they had been sourced from Ediacaran or Cambrian felsic rocks and consequently suggested the widespread presence of the Jamieson Volcanic Group at depth. The two lower layers were not ascribed to any particular unit, but the non-reflective middle layer varied in thickness from 0 to 8 km where not fault repeated and had a modelled density of 2600 kgm^{-3}. Their lower layer was up to 25 km thick, had variable responses and a density of 2760 kgm^{-3} (Cayley et al., 2011). Neither was considered magnetic.

An alternative interpretation is presented (Figure 4-17), which considers an understanding of the geology of western Tasmania and its interpreted extension across Bass Strait. The seismic data suggest that there is a difference in the character of the crustal layering at the northern end of Line 4 to that evident on Line 3 and the southern part of Line 4. At the northern end of Line 4, the upper layer is more continuous, strongly reflective and thicker than on Line 3. The middle, non-reflective layer of Cayley et al. (2011) appears to not continue further north than CDP 9500 on Line 4. Thereafter, it appears to include discrete reflectors, suggesting that it may be a different unit. On Line 3, the lower unit of Cayley et al. (2011) can be divided into an upper reflective unit approximately 2 s TWT (6 km) thick and a lower non-reflective unit, the top of which lies at approximately 8 s TWT (24 km). This division is not present at the northern end of Line 4, where the bottom, non-reflective layer appears to be truncated by a north-dipping fault that extends into the overlying Melbourne Zone and that crops out west of CDP 8600 (Welsh et al., 2011). This fault appears to mark the boundary between the 4 layers visible on Line 3 and the southern part of Line 4 and another, different rock package.

The layers of Line 3 and part of Line 4 are consistent with the crustal layers modelled for the Rocky Cape Zone. The lowest unit (Lower Rocky Cape Group in the models) may represent the deep crust inferred by Black et al. (2010), which may represent late Paleoproterozoic to Mesoproterozoic basement (Chapter 3). The strongly reflective unit above it may be from the Rocky Cape Group, which includes both quartzites and mudstones and even when metamorphosed would retain a significant lithological contrast, consistent with the strongly reflective seismic response.

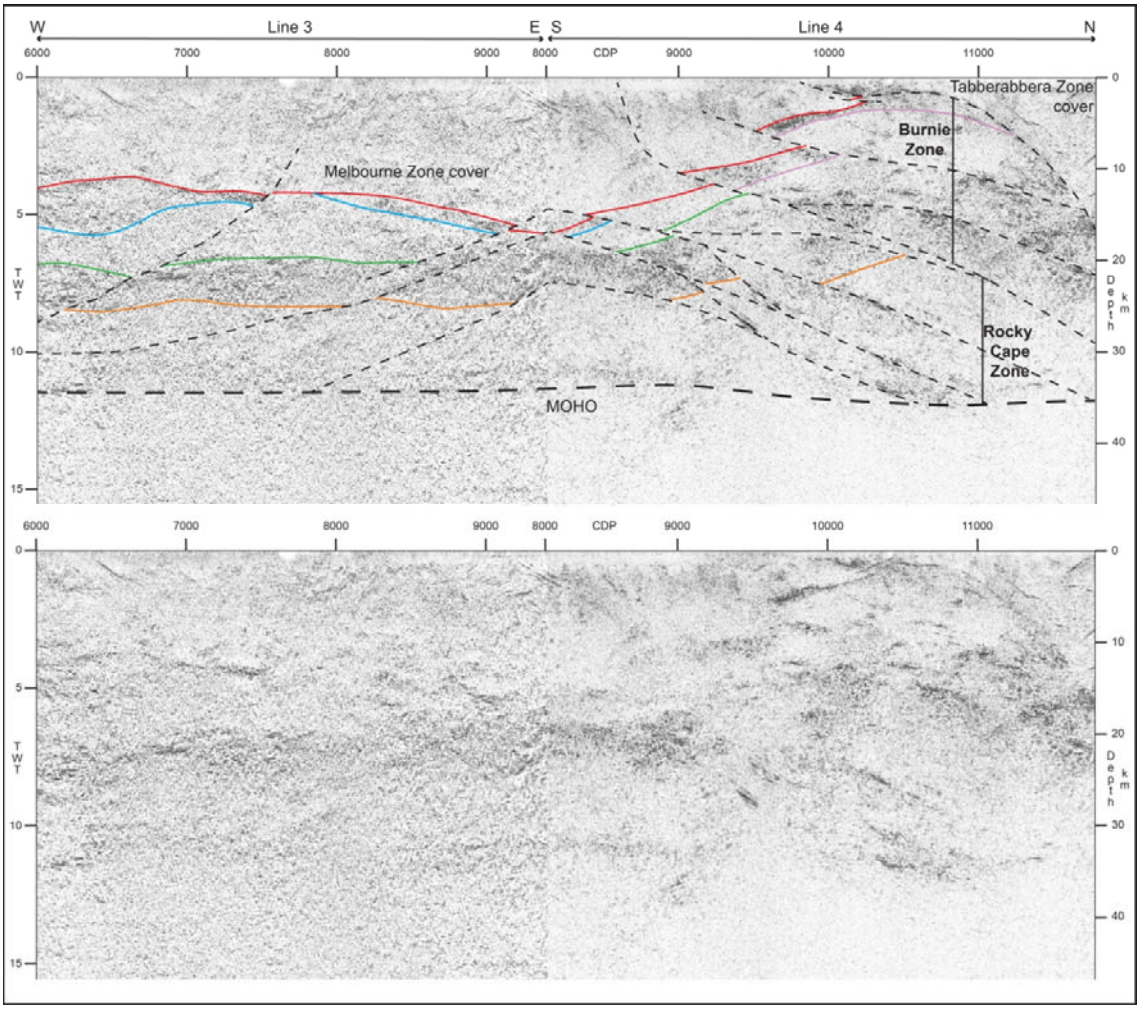


Figure 4-17. Geoscience Australia seismic lines 06GA-V3 and 06GA-V4, lower panel uninterpreted, upper panel interpreted, with emphasis on the basement. Location on Figure 4-2. In the Rocky Cape Zone, the red horizon marks the base of the Melbourne Zone, the blue horizon the base of an upper strongly reflective zone (possibly the Scopus Formation), the green horizon the base of a weakly reflective zone (the Smithton Basin), the orange horizon the base of a lower strongly reflective zone (the Rocky Cape Group). In the Burnie Zone, the lavender horizon marks the base of what is interpreted to be the Jamieson Volcanic Group, an equivalent to the Mt Read Volcanics. Below this, the Burnie Zone has a different character to the Rocky Cape Zone.

Overlying the interpreted Rocky Cape Group package is a layer that is characterised by bland seismic response with little reflective character. This response is similar to the seismic response of the seaward extension of the Smithton Basin off the coast of northwestern Tasmania imaged along AGSO seismic line 148/05. I propose the poorly reflective seismic package in Line 3 is equivalent to most of the Smithton Basin beneath the Melbourne Zone. The upper, discontinuous layer with a modelled density of 2780 kgm^{-3} may be from an equivalent to the Spinks Creek Volcanics, an upper unit in the Smithton Basin in Tasmania that is likely to have significantly different seismic character to the rest of the basin. It is not always present; on AGSO seismic line 148/05 it was mostly so shallow its responses could not be differentiated from those from the sea floor. The presence of these basaltic layers at the top of the Smithton Basin would yield higher densities than andesites of the Jamieson Volcanic Group. An alternative possibility is that the upper discontinuous layer might be from the Scopus Formation, which lies above the Smithton Basin. However, the Scopus Group would probably have lower density values than those forward modelled and so is considered a less valid interpretation. Furthermore, mafic volcanic rocks like the Spinks Creek Volcanics typically show a strong seismic contrast with the enclosing sedimentary sequences (e.g. O'Halloran and Johnstone, 2001).

The change from the interpreted seismic responses imaged on Line 3 (Rocky Cape Zone) to the thicker and more reflective seismic package along the northern part of Line 4 suggests a change in the geological architecture of the Selwyn Block. One possible explanation for this change may be elucidated by the interpretation of AGSO seismic line 148/05 by Barton (1999), which indicated that the eastern boundary of the Rocky Cape Zone is an east-dipping reverse fault, in which rocks of the Burnie Zone have been thrust over the Rocky Cape Zone. The eastern boundary of the Rocky Cape Zone can be traced across Bass Strait to southern Victoria. In all forward models this boundary is also an east dipping fault. The Jamieson Volcanic Group may indicate the presence of the Burnie Zone at depth, and this unit can be traced as far north as the Glen Creek Window, approximately 50 km southeast of Line 4. The Jamieson Volcanic Group crops out on basement highs in the Selwyn Block (VandenBerg et al., 2000; VandenBerg et al., 2006), and the most prominent basement high in the 2006 seismic survey over the Selwyn Block is that at the northern end of Line 4. Consequently, the change in seismic character along the northern section of line 4 is interpreted to indicate the presence of the Burnie Zone, which has been thrust over the Rocky Cape Zone.

An implication of this suggestion is that the Cambrian volcanic rocks are unlikely to be present over most of the Selwyn Block, as proposed by Rossiter and Gray (2008). This will be examined in the following section.

4.4.3 CENTRAL VICTORIAN GRANITES

The enclaves and geochemistry of central Victorian granites provide indirect mechanisms to gain insights into the nature of the middle and deep crust of the Selwyn Block. Many of the central Victorian granites where U/Pb zircon ages have been obtained have been formed at such high temperatures that almost all pre-existing zircons have been recrystallised (Clemens et al., 2014), and so one useful discriminator between the Tasmanian Precambrian and Lachlan Orogen (Chapter 3) is not possible using these granites. The zircons that yield dates that are older than the intrusive ages typically have Lachlan Orogen signatures, and are likely to be late stage xenocrysts accumulated during or after the ascent of the intrusions (Clemens and Bezuidenhout, 2014).

Nevertheless, the geochemical signatures of the granites do help constrain the basement. The initial $^{87}Sr/^{86}Sr$ ratios of these granites imply that they were formed from Neoproterozoic or Cambrian precursors (Gray, 1990; Rossiter and Gray, 2008). Maas and Nicholls (2012) demonstrated that the $^{87}Sr/^{86}Sr$ ratios rise from as low as 0.704 in the south to as high as 0.712 in the north of the Selwyn Block. The ε_{Nd} values are higher in the south, reaching +2, whereas values of -6 are obtained further to the north. These data suggest granites in the south are derived from more juvenile sources, which may include mantle melts, or Ediacaran and Cambrian back-arc or arc rocks, whereas source material for the granites to the north appears to be more evolved and with a greater crustal component. These relationships were particularly evident in the western Selwyn Block, and suggest that, at least here, the granite source rocks became increasing more evolved and presumably older to the north. Maas and Nicholls (2012) also considered the I-type granites on the Selwyn Block were sourced from a more juvenile mafic component, but were not able to distinguish whether this was from the remelting of pre-existing mafic rocks or from a syn-magmatic component. These geochemical data support the geophysical interpretation presented in this chapter (Table 4-2). For example, the southern part of the outcropping Melbourne Zone is interpreted to be underlain by the eastern parts of the King Island Zone, an Ediacaran back-arc basin sequence of mafic volcanic and meta-sedimentary rocks, where ε_{Nd} at 579 Ma ranged between -3.1 and +4.8 (Meffre et al., 2004). The northern part of the Melbourne Zone is interpreted to be underlain by the Rocky Cape Zone—

Mesoproterozoic shallow marine sedimentary rocks and the Neoproterozoic Smithton Basin. No ε_{Nd} or T_{DM} data are available on these units in Tasmania.

Pluton	Basement source rocks	Reference	Basement Zone	Lithologies of the underlying zone
Cobaw Complex	Metasediments, including carbonates	Anderson (1997)	Rocky Cape, King Island	Quartzite, siltstone, mudstone, meta-turbidites, mafic volcanic rocks, dolomite
Cerberean Cauldron	Arc-related greywacke	Clemens and Birch (2012)	Rocky Cape	Quartzite, siltstone, mudstone, dolomite, mafic volcanic rocks
Lysterfield Granodiorite	Intermediate volcanic rocks	Clemens and Bezuidenhout (2014)	King Island	Mafic volcanic rocks including picrites, meta-turbidites
Mt Disappointment Granodiorite	Alkali basalt and metasedimentary rocks	Clemens and Benn (2010)	Rocky Cape, King Island	Mafic volcanic rocks including picrites, meta-turbidites, quartzite, siltstone, mudstone
Strathbogie Complex	Andesite, dacite, greywacke, pelite	Clemens and Phillips (2014)	Rocky Cape	Quartzite, siltstone, mudstone, dolomite, mafic volcanic rock
Tolmie Igneous Complex	(Volcani)clastic greywacke	Clemens et al. (2011)	Burnie	Meta-turbidites, felsic to intermediate volcanic rocks
Warburton Granite	Metasediments, including carbonates	Anderson (1997)	Rocky Cape	Quartzite, siltstone, mudstone, dolomite, mafic volcanic rocks

Table 4-2. Source rocks for Victorian granites.

Rossiter and Gray (2008) found that when the Ba contents of the central Victorian Late Devonian granites are normalised to granites with 70% SiO_2, they have abnormally high Ba contents, up to 1700 ppm. The highest values are concentrated in the central and eastern Melbourne Zone or just east of it, and they considered that this indicated that the source rocks were, at least in part, felsic volcanic rocks. One possibility is that they are from the Cambrian Mt Read Volcanics (Cayley et al., 2011). However, as noted above, the Mount Read Volcanics have only been mapped as overlying the Burnie and Tyennan zones, and not the Rocky Cape Zone that is inferred to underlie much of the area with high Ba granites. An alternative possibility is that the erosional products of the Mt Read Volcanics or geochemically similar rocks were partly deposited on the Rocky Cape Zone. The Scopus Formation, which lies immediately above the Smithton

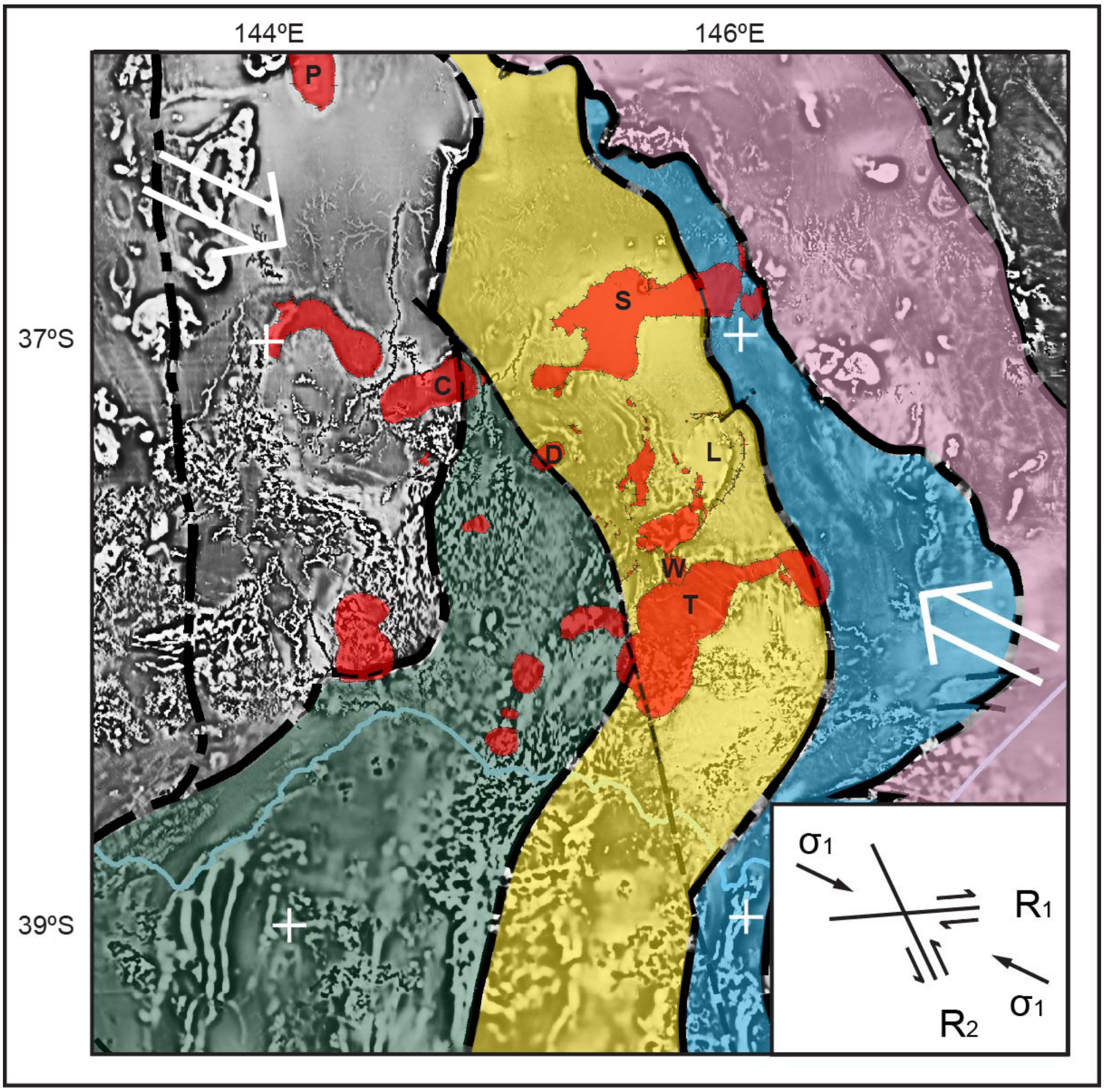


Figure 4-18. Central Victorian Upper Devonian granites (red) and their relationships to the basement outlined in Figure 4-5. C marks the Cobaw Complex, D the Mt Disappointment Granodiorite, L the Cerberean Complex, P the Pyramid Hill Granite, S the Strathbogie Complex and T the Tynong Granodiorite. The 070° trend is evident in their alignment, particularly in the Cobaw and Strathbogie complexes where the dominant strike of the basement is at 340°. This contrasts with the broadly north-south alignment seen in the magnetic Upper Silurian granites in the upper left hand corner of the image and the Pyramid Hill Granite. Background image is the AGC filtered total magnetic intensity. White arrows indicate possible σ_1 direction. Inset ahows suggested Reidel directions that may have influenced the emplacement of the granites. Magnetic data from Geoscience Australia and the Geological Survey of Victoria.

Basin, was deposited in the Late Middle Cambrian, approximately 500 Ma (Everard et al., 2007), and is coeval with the Mt Read Volcanics (Mortensen et al., 2015). It also contains ultramafic clasts, consistent with at least some of its sedimentary rocks having been sourced from the adjacent Burnie Zone (Everard et al., 2007). As well, in the Smithton Basin, Everard et al. (2007) recorded the presence of felsic volcanic clasts in the uppermost Cryogenian Croles Hill Diamictite, and this is another possible felsic source for the Devonian granites. However, Clemens and Phillips (2014) suggested that higher Ba concentrations in granites could also be due to increased melting of a protolith where the granite formed by peritectic assemblage entrainment. If so, then the proposition by Rossiter and Gray (2008), that the high Ba was from a felsic volcanic protolith, may not be valid. The explanation of Clemens and Phillips (2014) is more consistent with the geophysical interpretation presented in this chapter.

Anderson (1997) located two granites from the Selwyn Block that contained calc-silicate enclaves, the Warburton Granodiorite and the Cobaw Complex (Figure 4-18). The Warburton Granodiorite is an S-type granite, for which Elburg (1996a) gave a Sm/Nd isochron of 551±129 Ma, whilst Richards and Singleton (1981) found an intrusive age of 369±13 Ma (K/Ar on biotite). Elburg (1996a) reported ε_{Nd} at 369 Ma of -6.23 to -6.55 and an $^{87}Sr/^{86}Sr$ ratio of 0.7097 to 0.7100. The calc-silicate enclaves typically contain quartz, Ca-plagioclase, calcite, biotite, titanite, diopside, Ti-magnetite, apatite and hornblende. The mafic enclaves present in the granite were considered to be derived from a juvenile mantle-like melt (Elburg, 1996a). The body was intruded into the Humevale Siltstone, the Wilson Creek Shale and the Norton Gully Sandstone. These are at the same stratigraphic level as the Lilydale Limestone, approximately 30 km away but there is no limestone mapped any closer, and it appears likely that the calc-silicate rocks are from a deeper level, potentially the Black River Dolomite or the Smithton Dolomite of the Smithton Basin.

The Cobaw Complex lies on the boundary between the Melbourne and Bendigo zones. It intrudes the Heathcote Volcanic Group and the overlying chert, black shale and turbidite rocks of the Bendigo Zone and the Silurian and Devonian clastic sedimentary rocks of the Melbourne Zone. Anderson (1997) gave a U-Pb zircon age of 356±4 Ma, ε_{Nd} of -4.4, and a T_{DM} age of approximately 1800 Ma, with calc-silicate enclaves ranging from 1700 to 1900 Ma. The calc-silicate enclaves were only found east of the Heathcote Volcanic Group, eliminating it and the overlying rocks of the Bendigo Zone as a possible source unless the enclaves were stoped down into the magma. The calc-silicate enclaves are 80 km from the Lilydale Limestone in the Melbourne Zone, the closest carbonate rocks exposed. Again, it appears more likely that the enclaves have come from a deep, concealed source, perhaps the dolomite-rich units of the Smithton Basin.

Images of the magnetic data over the Melbourne Zone show the presence of broad 200 nT magnetic anomalies under the Strathbogie and Tynong granites and the Cerberean Caldera (compare Figures 4-6 and 4-18). These are part of a larger 60 nT high that extends over 250 km north from the Victorian coastline and is up to 80 km wide. McLean et al. (2010) interpreted the more local highs as resulting from serpentinised ultramafic rocks at depths of approximately 15 to 20 km, thicknesses of 5 to 10 km, over an area of 20,000 km^2 and with a susceptibility of 0.04 SI units and a density of 2.9 kgm^{-3}. This is much larger than the largest area of ultramafic rocks seen in Tasmania, 1,200 km^2.

Another possibility is that this broad response could be from a more widely spread mafic unit. Grant (2002) recorded that, near the Stawell Granite, Carrolls Amphibolite, originally a MORB-like sequence, had a median susceptibility of 4.5×10^{-3} SI units and ranged up to 100×10^{-3} SI units. In Tasmania, the mafic Spinks Creek Volcanics, in the upper part of the Smithton Basin, has a median susceptibility of 30×10^{-3} SI units, with some values exceeding 100×10^{-3} SI (Everard et al., 2007) and this rock type would generally be expected to increase its magnetic response with increasing metamorphism (Isles and Rankin, 2013). It crops out over an area of 30x60 km and its magnetic response can be traced offshore suggesting an extent of at least 10,000 km^2 in a single, continuous sheet. Upward continuation of the magnetic data (Figure 4-3) shows a continuation of the high across Bass Strait and into the Melbourne Zone. The present modelling showed a wide range of values, from 0.01 to 130×10^{-3} SI units (Figures 4-9, 4-10 and 4-12 to 4-16). In Victoria, the more local anomalies are located beneath outcropping intrusions, which suggests that they are more likely to be at least partly due to metamorphic effects during the intrusive event, perhaps on some basaltic rocks similar to the Spinks Creek Volcanics.

Clemens and Benn (2010) found that the Upper Devonian Mt Disappointment Granodiorite (Figure 4-18) had abnormally high Ni and Cr contents, a wide range $^{87}Sr/^{86}Sr_{370}$ ratios and the presence of pseudomorphs after orthopyroxenes. They attributed these to inhomogeneities at the source, with both greywacke and basalt present. The intrusion lies on the boundary between the King Island and Rocky Cape zones and on the northern extension of the MORB-type tholeiite basalts and picrites that crop out on the southeastern coast of King Island (Meffre et al., 2004). It seems likely that the mafic component was derived from these basalts and the greywacke from either the Rocky Cape Group or, more likely, sedimentary rocks interlayered between the basalts.

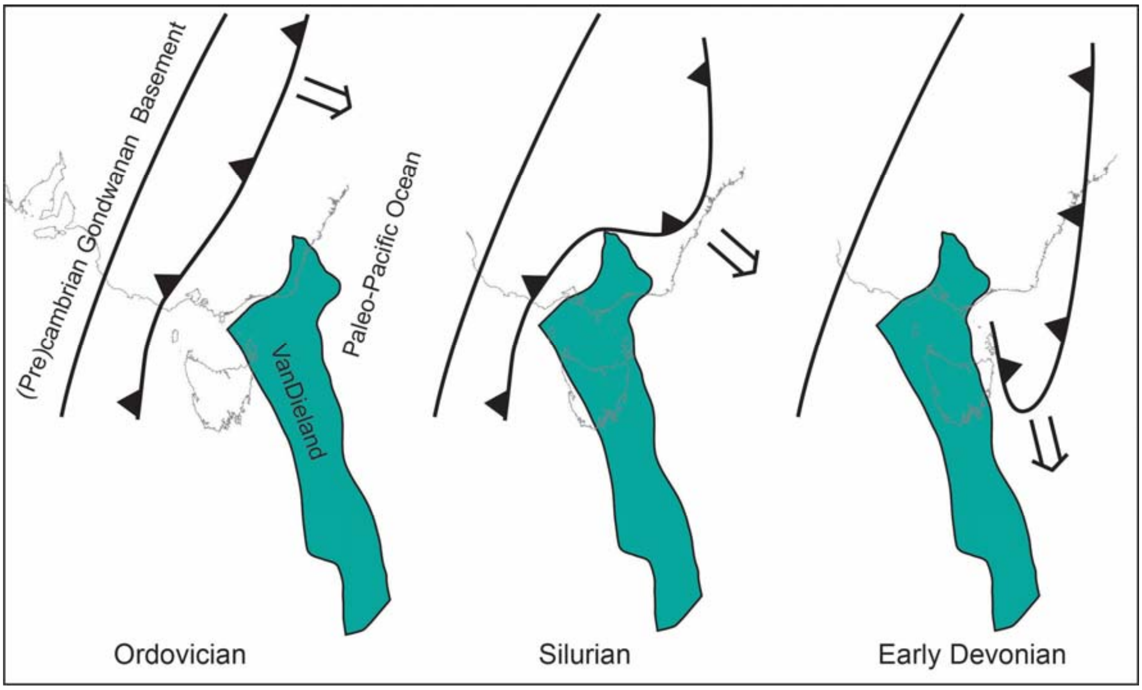


Figure 4-19. Early Paleozoic subduction in southeastern Australia, outlining the development of an orocline in response to the presence of VanDieland and the retreat of a subduction zone. After Moresi et al. (2014).

Finally, a distinctive characteristic of the Upper Devonian granites on the Selwyn Block north of approximately 38°30’ is that their long axes strike at approximately 070° (Figure 4-18). This contrasts with Upper Devonian granites elsewhere and the dominant strike of the host rocks and the long axes of the nearby Upper Silurian granites, which tend to strike close to 000°. This unusual strike also reflects the strikes of the suites in the region (Chappell et al., 1988). I speculate that this is a mid-crustal effect, where the Upper Devonian granites reflect part of a conjugate shear set, the other half being the boundaries either side of the Rocky Cape Zone. This implies that the granites were formed in a stress field where σ_1 was approximately 115° and σ_3 approximately 025°. In contrast, the Upper Devonian Pyramid Hill Granite, intruded west of the Selwyn Block, has a long axis trending 000°, similar to the Upper Silurian granites in the area.

4.4.4 THE OROGEN ENCLOSING VANDIELAND

Following Reed (2001) and Chapter 2, correlation between the Early Paleozoic Eastern Tasmanian Zone with the Early Paleozoic Tabberabbera Zone in Victoria is proposed. This correlation requires that the eastern boundary of the Selwyn Block (the Governor Fault), correlates with the Tamar Fracture Zone in Tasmania (Figure 4-5). However, the Tabberabbera Zone forms the western portion of an oroclinal structure in eastern Victoria and southeastern New South Wales (Cayley, 2012; Moresi et al., 2014). Geodynamic modelling by Moresi et al. (2014) suggested the orocline formed in response to the collision of VanDieland and east Gondwana (Figure 4-19). The orocline developed 'west' of VanDieland as the subduction zone attempted to re-establish itself after the Tyennan Orogeny. By the Silurian, the plate reorganisation associated with this collision caused the rifting off of much of the Early Paleozoic Gondwanan margin in western Queensland and western New South Wales and a U-shaped Banda Sea-style subduction zone began to form north of VanDieland. Through the Silurian and into the Devonian, this U-shaped subduction zone retreated south, outboard (east) of VanDieland.

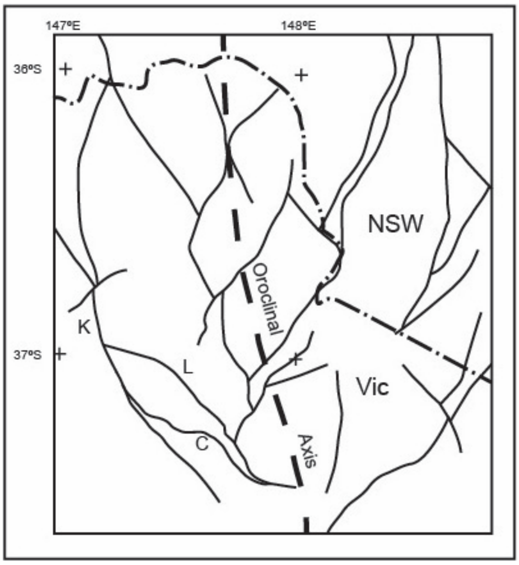


Figure 4-20. Faulting patterns in eastern Victoria and southeastern New South Wales in the vicinity of the suggested orocline. K marks the Kiewa Fault, which at its northern end strikes 350° but to the south the strain is taken up on the Lochiel (L) and Cassilis (C) faults that at their southern ends strike 280°. Most of the faults appear to have been formed at the same time, with northwest trending faults both truncating and being truncated by northeast trending faults. Fault patterns abstracted from Simons and Moore (1999) and Scheibner (1997).

The orocline in eastern Victoria and southeastern New South Wales is the remnant of this retreat, where the Ordovician turbidite rocks, initially formed along the eastern Gondwana margin, wrapped around VanDieland as they followed the retreating subduction zone (Moresi et al., 2014). For example, at the Tabberabbera-Omeo Zone boundary, the Kiewa Fault has a distinctive convex-south pattern. It changes from a strike of approximately 160° to 090°, with much of the strain taken up on the Cassilis and Lochiel faults (Figure 4-20). Further north, in the axis of the orocline outlined by Moresi et al. (2014), the Omeo and Deddick zones are characterised by an interlaced set of coeval northwest-trending and northeast-trending faults, and this pattern continues north into New South Wales (e.g. Scheibner, 1997). An implication of this geodynamic model is that similar structures might be present in the Eastern Tasmanian Zone. However, there is a narrower region of outcrop than on the mainland and most of it is covered by younger rocks or intruded by granite, making detection more difficult. Nevertheless, some similarities in character can be seen.

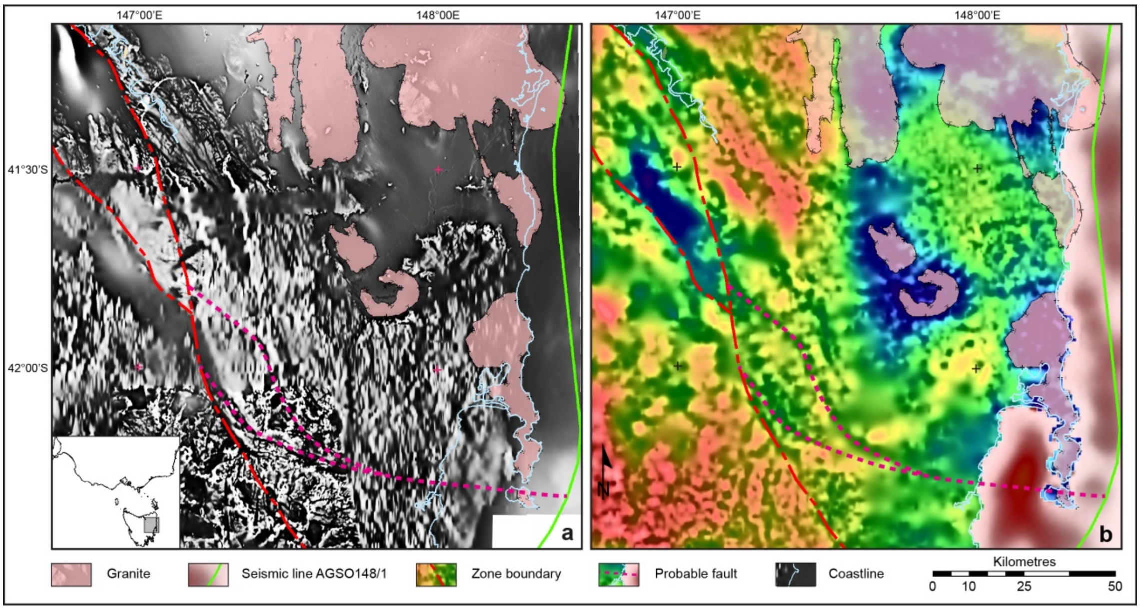


Figure 4-21. Basement structures concealed under younger cover that may be analagous to those seen in eastern Victoria. a. Magnetics, b, isostatic gravity onshore, free air offshore. Zone boundaries from Chapter 3. Magnetic and gravity data from Geoscience Australia.

Lows in the magnetic and gravity responses seen in Figure 4-21 curve away from the western boundary of the Eastern Tasmanian Zone in a convex-south shape, with the strike changing from approximately 140° to 100°, a similar pattern to that seen on the Kiewa Fault. These shapes are interpreted as faults in the basement that have been reactivated, perhaps at about 90 Ma in the breakup of Gondwana (Sdrolias et al., 2003; Berry, 2015), and so have created sites of increased alteration and weathering in the overlying Tasmanian Dolerite. In images of offshore seismic data, one of these zones correlates with what can be interpreted as a fault in AGSO seismic line 148/1 (Figure 4-22). Images of the seismic section suggest an interlaced pattern of faulting, with both north-dipping and south-dipping faults formed at the same time. This fault pattern is similar to the axial zone of the orocline on mainland Australia. While there is no conclusive proof of the development of an orocline in the Eastern Tasmanian Zone,

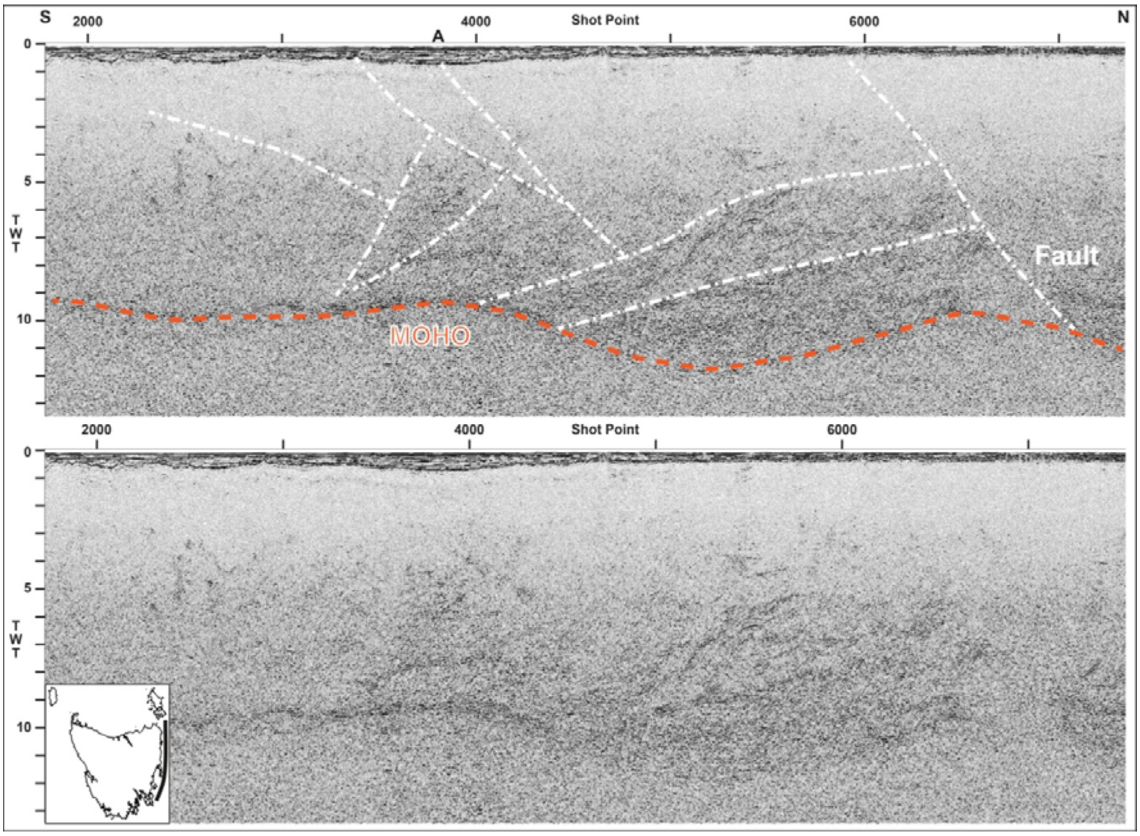


Figure 4-22. AGSO seismic line 148/01, north end. Upper panel interpreted, lower panel uninterpreted. The inset gives the location of the imaged segment. No vertical exaggeration. North- and south-dipping faults intersect each other, suggesting formation at the same time. The Fault at A is aligned with the curved fault in Figure 4-21.

the fault patterns seen there are consistent with this possibility. The detrital zircon populations in the Ordovician to Early Devonian Eastern Tasmanian Zone may also be significant. Black et al. (2004) noted the strong similarity of the detrital zircons there to those found elsewhere in the Lachlan Orogen. There was no evidence of zircons from the 496-508 Ma Mt Read Volcanics (Mortensen et al., 2015) or the distinctive 1400 Ma detrital zircons seen throughout VanDieland (Black et al., 2004) (Chapter 3). It is likely that none of the sedimentary rocks in the Eastern Tasmanian Zone were sourced from the now-adjacent VanDieland. One possible explanation for this is that the two regions that are now adjacent were not juxtaposed until after sedimentation was largely, if not completely, finished. The model of Moresi et al. (2014) suggests this possibility.

Figure 4-23 indicates that U/Pb zircon age dates on 'Tasmanian' granites have a distinctive spatial relationship. The oldest lie in the northeast corner, on Flinders Island

(401±4 Ma), and that they young in a clockwise direction, with the youngest on the eastern Tasmanian coast having an age of 374±3 Ma, one granite on the south coast has an age of 370±3 Ma, and the youngest 'Tasmanian' granite, the Sandblow Granite, lies in the northwest, on King Island (351±2 Ma) (Black et al., 2010). Black et al. (2010) suggested that the 390 to 400 Ma granites in eastern Tasmania were formed over an east-dipping subduction zone, and that the down-going slab broke off at about 380 Ma, generating the granites in western Tasmania. The model of Moresi et al. (2014) is consistent with this hypothesis, in that the east-dipping subduction zone of Black et al. (2010) may be the western part of the southwards-retreating U-shaped Banda Sea-style

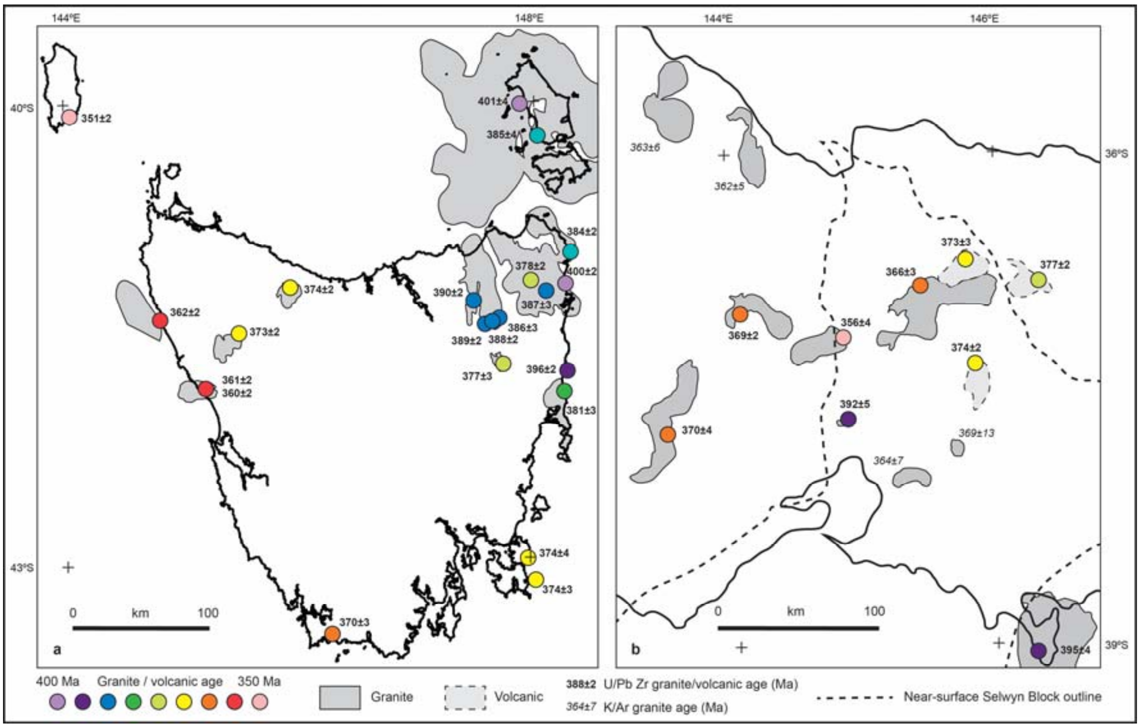


Figure 4-23. Tabberabberan age granites in Tasmania (a) and central Victoria (b), with ages that have been colour grouped into intervals of approximately 5 million years to highlight age trends. Only the granite complexes where SHRIMP or K/Ar on muscovite ages have been determined are shown. Tasmanian ages from Black et al. (2010), Victorian U/Pb zircon ages from Anderson (1997), Arne et al. (1998), Bierlein et al. (2001b), Clemens et al. (2014), Compston (2004), Elburg (1996b) and Kemp et al. (2006), and K/Ar ages on muscovite from Richards and Singleton (1981). The ages generally show clockwise younging, with the oldest granites in the northeast and the youngest in the northwest.

subduction system referred to above. This system may have initially generated granites on Flinders Island at 401±4 Ma while retreating south along the east Tasmanian coast. Shortly after 380 Ma, the system tore along its western margin and so allowed mafic under-plating of both eastern and western Tasmania, giving rise to further granite production. Under-plating began closest to the tearing, in southeastern Tasmania, at approximately 375 Ma, but later gave rise to granites as young as 360±2 Ma in western

Tasmania and ultimately to the 351±2 Ma granite on King Is, the most distant granite to the subduction system (Figure 4-23a). The Central Victorian Granite Province does not show a similar age pattern (Figure 4-23b). Rather, there appears to have been distinct magma pulses at approximately 395 Ma, 375 Ma (preserved as volcanic units) and at 368 Ma. K/Ar ages from the Lake Boga Granite (363±6 Ma) and the Pyramid Hill Granite (362±5 Ma) are close enough that they could be included in the 368 Ma grouping since the errors overlap and the closure temperature of the K/Ar muscovite system is significantly lower than that of the U/Pb zircon system.

4.5 CONCLUSIONS

The Proterozoic western Tasmanian crust continues north under Bass Strait to become the Selwyn Block. Its eastern boundary in Tasmania, the Tamar Fracture Zone, continues north and is equivalent to the Governor Fault in Victoria. The King Island, Rocky Cape and Burnie zones (Chapter 3) can be interpreted northward across Bass Strait, and are evident in the magnetic data in the Selwyn Block in Victoria. They are also evident in Victorian outcrops. For example, in Tasmania neither the Rocky Cape or King Island zones include mafic-ultramafic complexes or the Mt Read Volcanics. The present interpretation northward into Victoria is consistent with this; all fall inside the Burnie Zone, including the Cambrian volcanic rocks inferred in the northern part of seismic line 06GA-V4. In the western Melbourne Zone, the Ceres Metagabbro lies within the western part of the King Island Zone and is interpreted to have formed as part of the Neoproterozoic breakup sequence. The pre-Cenozoic basalt on Phillip Island is a northern extension of the 580 Ma mafic volcanic rocks on King Island.

Most of the Melbourne Zone is underlain by the Rocky Cape Zone. Most of seismic lines 06GA-V3 and 4 show a distinct layering that can be interpreted as reflecting the layering present in the Rocky Cape Zone—an uppermost Spinks Creek Volcanics, the lower Smithton Basin, the Rocky Cape Group, and an unseen 'Lower Rocky Cape Group' otherwise only known from a population of zircons in Tasmanian granites that is not present in the host rocks. Quartzite clasts recorded from a conglomerate in the southeastern Melbourne Zone have no known source in the Melbourne Zone, but are consistent with derivation from the Rocky Cape Zone.

Features in the granites also reflect the underlying rocks. Calc-silicate enclaves in the eastern parts of the Cobaw Complex and in the Warburton Granodiorite seem to have been sourced from the Smithton Basin equivalent rocks. The Mt Disappointment Granodiorite appears to have formed from the basaltic and metasedimentary rocks of the eastern King Island Zone. The $^{87}Sr/^{86}Sr_{370}$ ratios and many other features of the granites are compatible with formation from the uppermost Neoproterozoic rocks such as the Smithton Basin or from the eastern King Island rift sequences. In the western Melbourne Zone, both the $^{87}Sr/^{86}Sr_{370}$ ratios and the ε_{Nd} values indicate that the granite parent material becomes older to the north. The granite orientations appear to be controlled by the orientation of the mid-crustal fabrics as shown by the orientations of the boundaries of the Rocky Cape Zone.

The Smithton Basin is marked by a broad magnetic high from the Spinks Creek Volcanics, and this high generally extends north to the Melbourne Zone (Figure 4-3). If some of the Upper Devonian central Victorian granites were sourced from the Smithton Basin, then the upper units there would likely have been metamorphosed by the granites during or shortly after emplacement. The Spinks Creek Volcanics is the most likely unit to have been affected and to give a significant response, and is considered to be the cause of the broad highs seen in the Melbourne Zone.

The Eastern Tasmanian Zone continues north to become the Tabberabbera Zone in eastern Victoria. The eastern margin of the King Island Zone is defined in the isostatic gravity data. The Tyennan and Sorell-Badger Head Zones are restricted to regions close to Tasmania.

The effects of oroclinal development seen in eastern Victoria also appear to be reflected in eastern Tasmanian geology. The distinctive faulting patterns of both regions are quite similar. These are attributed to the formation of a U-shaped (Banda Sea style) retreating subduction system described by Moresi et al. (2014). As the development of this moved south, so did granite production in eastern Tasmania. When the subducting slab tore, shortly after 380 Ma, it allowed mafic underplating, which started in eastern Tasmania but spread west and ultimately north to King Is, forming the Sandblow Granite at 351±2 Ma. The timings of granite emplacement in the Central Victorian Granite Province suggest that slightly different processes have gone on there.

Thus the subdivisions seen in the outcropping part of VanDieland can be carried north into the buried Selwyn Block. These subdivisions can be carried to the northern end of the Selwyn Block, and together with those outlined in Chapter 3, give rise to a strato-tectonic map of VanDieland that stretches over 1700 km to the southern end of the east South Tasman Rise. As well, the effects of VanDieland colliding with and being embedded into the Lachlan Orogen appear to be present on both sides of Bass Strait.

4.6 ACKNOWLEDGEMENTS

The geophysical data used here were acquired by Geoscience Australia and the Geological Survey of Victoria, and the magnetic and gravity data were stitched by Geoscience Australia. Mapping by the Geological Survey of Victoria and Mineral Resources Tasmania is the basis for many of the comments on field relationships in those states.

Chapter 5

Day by day natural science accumulates new riches…. The true system of the world has been recognised, developed and perfected….. Everything has been discussed and analysed, or at least mentioned.

J. d'Alembert, 1759, Elements of Philosophy.

Conclusions

5.1 SUMMARY OF FINDINGS

This study of the basement (pre-Cretaceous) geology of the Bass Strait region and adjacent areas has supported the concept of the Selwyn Block model of VandenBerg et al. (2000) and Cayley et al. (2002). Geophysical interpretation, modeling, and a regional geological synthesis suggest that the Proterozoic and Cambrian cratonic crust of western Tasmania continues across Bass Strait and underlies the Melbourne Zone and adjacent areas in Victoria. The Tamar Fracture Zone in Tasmania can be correlated with the Governor Fault in Victoria, as proposed by Reed (2001). Other models that require alternative reconstructions (e.g. Gibson et al., 2011; Jacob and Dyment, 2014) are considered to be invalid.

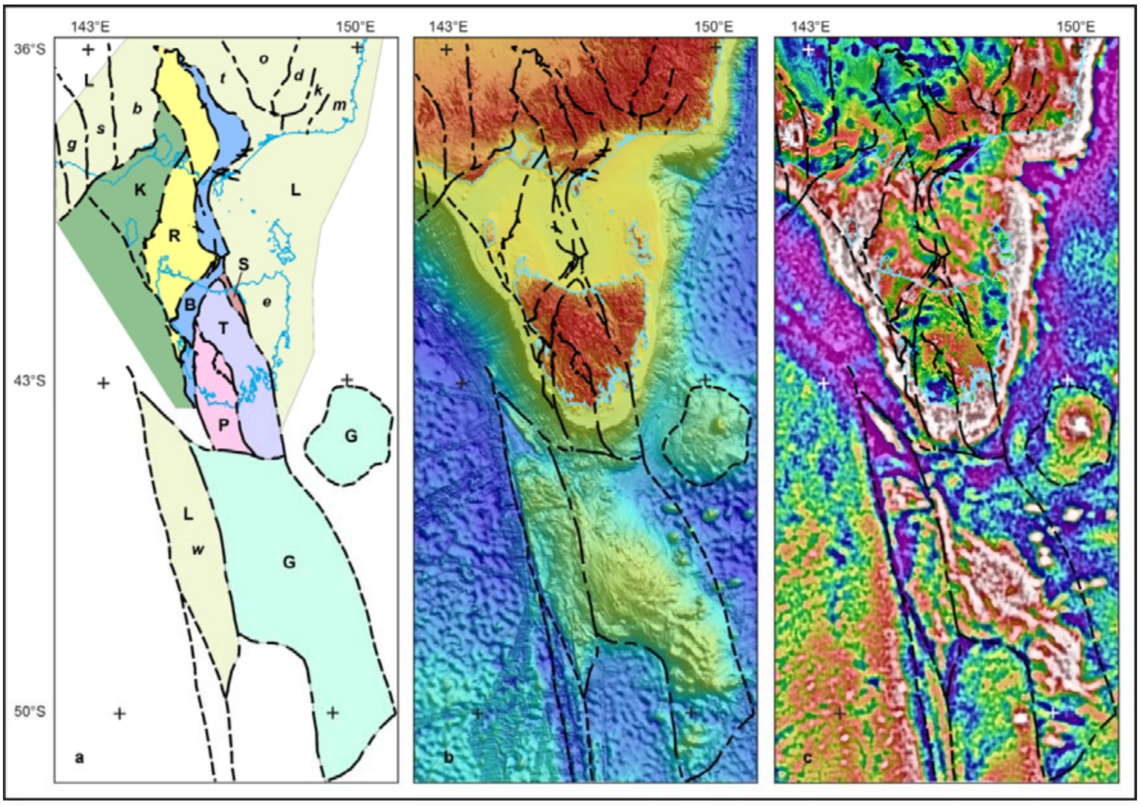


Figure 5-1. VanDieland. a. Zones in VanDieland. B Burnie, G Glomar, K King Island, P Pedder, R Rocky Cape, S Sorell-Badger Head, T Tyennan. VanDieland is surrounded by the Lachlan and Delamerian orogens L, subdivided into the following zones; *b* Bendigo, *d* Deddick,*e* East Tasmania, *g* Grampians-Stavely, *m* Mallacoota, *o* Omeo, *s* Stawell,and *t* Tabberabbera and *w* West South Tasman Rise. b. Zone boundaries with bathymetry and topography. c. Zone boundaries with isostatic gravity onshore and free air gravity offshore.

VanDieland (the Selwyn Block, western Tasmania, the east South Tasman Rise and the East Tasman Plateau; Cayley, 2011) has been subdivided into seven different zones

(Figure 5-1) and a geological history has been outlined for it that accounts for the major variations in each zone. From the west, these zones are the King Island, Rocky Cape, Pedder, Tyennan, and Sorell-Badger Head zones, with the Glomar Zone covering the east South Tasman Rise and the East Tasman Plateau. The west South Tasman Rise has been excluded from VanDieland as the detrital zircon populations suggest a closer correlation with the Delamerian Orogen of South Australia than the Paleozoic or Proterozoic crust in VanDieland. To the east of VanDieland, the East Tasmania Zone is correlated with the Tabberabbera Zone in Victoria.

The westernmost three zones, the King Island, Rocky Cape and Burnie zones, are interpreted to continue north under Bass Strait to form the Selwyn Block. This is evidenced by the magnetic and gravity interpretations that indicate continuity across Bass Strait. Outcrops in the Melbourne Zone in Victoria support this continuity. The Ceres Metagabbro is correlated with ?Neoproterozoic magnetic units west of King Island. The basalt that crops out on Phillip Island is correlated with the MORB and rift tholeiites on the east coast of King Island. The Mt Disappointment Granodiorite has features that suggest derivation from the mafic rocks in the King Island Zone. Both the Burnie Zone in Tasmania and its northern extension into Victoria include ultramafic and mafic rocks. In Tasmania, the Mt Read Volcanics are restricted to the Burnie Zone and zones further east. In Victoria, the coeval and geochemically identical Cambrian intermediate and rhyolitic volcanic rocks present in the eastern Melbourne Zone also lie in the northern extension of the Burnie Zone.

The Rocky Cape Zone does not crop out in Victoria, but the granites there suggest its presence, perhaps most clearly in the calc-silicate enclaves present in the Warburton Granodiorite and the eastern end of the Cobaw Complex. These are otherwise difficult to explain, since the Melbourne Zone is almost entirely comprised of clastic sedimentary rocks, but the Ediacaran-Cryogenian Smithton Basin of the Rocky Cape Zone includes the Black River and Smithton dolomites. Clasts of metaquartzite are present in sedimentary rocks in the southeastern Melbourne Zone and these too have no obvious source in the presently exposed sequences, but the lithology is well known in the Rocky Cape Zone in Tasmania. In the western Melbourne Zone, the transition from low $^{87}Sr/^{86}Sr$ ratios and high ε_{Nd} values in the south to high $^{87}Sr/^{86}Sr$ ratios and low ε_{Nd} values in the north can be explained by the transition from Upper Neoproterozoic King

Island Zone basement to Mesoproterozoic or older Rocky Cape Zone basement. As well, images from the Victorian 2006 deep seismic survey can be interpreted as indicating the same layering as is present in the Rocky Cape Zone in Tasmania. The Spinks Creek Volcanics, part of the Smithton Basin, are inferred to give rise to the broad magnetic highs present under the Strathbogie Complex, the Cerberean Cauldron and the Tynong Granite.

The oldest VanDieland crust is likely to have formed between the Mexico-Arizona border region in Laurentia and the central Transantarctic Mountains in East Antarctica. As Rodinia broke up, Laurentia and East Antarctica separated, and so by 580 Ma or shortly thereafter VanDieland was an aggregation of extended micro-continental fragments loosely held together by thinned continental crust. A modern analogue might be the Porcupine, Hatton and Rockall banks in the North Atlantic Ocean (e.g. Figure 1, Péron-Pinvidic and Manatschal, 2010). This aggregation then moved 'north' along the Terra Australis margin (Cawood, 2005), until by approximately 520 Ma it lay outboard of its present position with respect to Gondwana. VanDieland was then progressively shortened over the next 25 million years as it came under the influence of the circum-Pacific subduction cell that has operated throughout the Phanerozoic, as well as other more local subduction events. One of these local subduction events shortened the back arc basin between the King Island Zone and other zones further east, and as the subducting slab retreated, it gave rise to the economically important Cambrian Mt Read Volcanics. The Owen Conglomerate and correlates are an expression of post-collisional extension. From the end of the Ordovician, VanDieland was progressively accreted into Gondwana, with its present position finally being achieved in the Tabberabberan Orogeny through the mechanism outlined by Moresi et al. (2014). The faulting patterns seen in the East Tasmania Zone resemble those seen in the Tabberabbera, Deddick and Omeo zones in Victoria (Figure 5-1), suggesting that similar processes took place in eastern Tasmania. The general clockwise younging of granites in Tasmania, from as old as 401±4 Ma on Flinders Is through 374±3 Ma east of Hobart to 351±2 Ma on King Island is consistent with this hypothesis. It is suggested that a southwards-retreating Banda Sea-style subduction system to the east of VanDieland gave rise to the granites in eastern Tasmania that young to the south. After the downgoing slab tore, mafic underplating could take place and this thermal event spread west, away from the tear, reaching King Island last. This has not happened in central Victoria. Rather, there appears to have been distinct pulses of magma generation at approximately 395 Ma, 375 Ma and 368 Ma.

5.2 DIRECTIONS FOR FUTURE RESEARCH

This study has described one exotic fragment in the Terra Australis Orogen. More speculative possibilities include the Hay-Booligal Block (Glen, 2013) and the Beardmore micro-continent in Antarctica (Stump et al., 2003). However, if the commonly suggested analogue of this margin to the present southwestern Pacific (e.g. Crawford et al., 2003b) is valid, then other so far unrecognised fragments may lie along this margin. The initial recognition of the Selwyn Block came about by seeing that the magnetic units of western Tasmania crossed Bass Strait and continued into central Victoria. Other fragments may similarly be recognised by interpreting potential field geophysical responses in areas of cover. In eastern Australia, large areas of Paleozoic rocks are covered by the Murray Basin and the pre-Cenozoic geology is poorly known. Further north, in southern Queensland, the Paleozoic geology is even less well understood as it lies under the Eromanga, Cooper and Surat basins. Continental crustal fragments like the Lord Howe Rise would be difficult to recognise, as they would not crop out but be covered by Paleozoic and younger sedimentary rocks. The relationships of the Cambrian Western Province terranes in New Zealand (Adams et al., 2015) to the Lachlan Orogen and the rest of the Terra Australis margin is another poorly-resolved question.

The ubiquitous traces of 580 Ma zircons in the Paleozoic rocks of the Tasman Orogen can be attributed to igneous rocks formed in the last stages of regional rifting along the eastern Australian part of the Rodinian margin. But older populations are sometimes present, notably 600 to 700 Ma in the Delamerian Orogen of South Australia (Ireland et al., 1998), in the Anakie Metamorphic Group (Fergusson et al., 2001) and in Recent beach sand deposits in eastern Australia (Veevers, 2007). These have been considered to have come from Antarctica (Veevers, 2007), but it is possible that thinned crust now hidden under younger basins might also have contributed to this population. The 660 Ma granite in the Andersons Creek Ultramafic Complex (Chapter 3, Sorell-Badger Head Zone) is one example where the appropriate zircons have formed, although here the outcrop is too small to have made a significant contribution to the overall zircon budget of the region. A further possibility is that this population may have been derived from Laurentia (e.g. Balgord et al., 2013). If so, the data may help to constrain the timing of the breakup of Rodinia. Lu-Hf studies on the zircons could provide more constraints on their source.

As the basement to the Central Victorian Granite Province becomes better understood, this may open up opportunities for investigation. All intrusions are end products of their constituent starting materials and the P-T conditions prior to and during their crystallisation. Any additional controls on any of these factors should allow tighter constraints to be placed on other variables. In turn, this could lead to a better understanding of the processes that went on in the formation of the Central Victorian Granite Province, and perhaps also to a more generalised understanding of the formation of granites and the mineral deposits that granites generate. For example, what are the fundamental controls that drove the formation of the King Island scheelite deposits and the many tin deposits in the Burnie Zone in Tasmania and in the Omeo Zone in eastern Victoria, and why are these not present elsewhere in the region studied? Are these tin and tungsten deposits the indicators of other mineral deposits in the district, such as intrusion-related gold deposits (e.g. Thompson et al., 1999)?

This thesis has not examined any of the potential linkages of the Pre-Ordovician geology of Bass Strait to the Bass Basin. Yet the Cretaceous rifting in the Bass Basin appears to have been controlled by the pre-existing basement fault structures. The Chat Accommodation Zone (Figure 1, Briguglio et al., 2013) partitions the Durroon Sub-basin from the larger Cape Wickham Sub-basin. This partition overlies the western boundary of the East Tasmania Zone. The western end of the Cape Wickham Sub-basin parallels the eastern boundary of the King Island Zone and overlies a change in the magnetic character of the Rocky Cape Zone. It is possible that the new understanding of the major basement zones and fault patterns could lead to a better understanding of the structural evolution of the Bass Basin and so to more hydrocarbon discoveries. Ideally, this would be done in a 3-dimensional model that includes both the Bass Basin and the underlying basement. It would also be integrated with the 3-dimensional model developed by GeoScience Victoria (Rawling et al., 2011) and that currently being developed by Mineral Resources Tasmania.

References

Adams, C.J., Mortimer, N., Campbell, H.J., Griffin, W.L., 2015. Detrital zircon ages in Buller and Takaka terranes, New Zealand: constraints on early Zealandia history. New Zealand Journal of Geology and Geophysics, 1-26. doi:10.1080/00288306.2015.1025798.

Aitken, A.R.A., Young, D.A., Ferraccioli, F., Betts, P.G., Greenbaum, J.S., Richter, T.G., Roberts, J.L., Blankenship, D.D., Siegert, M.J., 2014. The subglacial geology of Wilkes Land, East Antarctica. Geophysical Research Letters 41, 2390-2400. doi:10.1002/2014GL059405.

Allibone, A., Wysoczanski, R., 2002. Initiation of magmatism during the Cambrian–Ordovician Ross orogeny in southern Victoria Land, Antarctica. Geological Society of America Bulletin 114, 1007-1018. doi:10.1130/0016-7606(2002)114<1007:iomdtc>2.0.co;2.

Andersen, T., 2014. The detrital zircon record: Supercontinents, parallel evolution—Or coincidence? Precambrian Research 244, 279-287. doi:http://dx.doi.org/10.1016/j.precamres.2013.10.013.

Anderson, J.A.C., 1997. A structural, metamorphic, isotopic and geochemical study of metasedimentary enclaves from the southern Lachlan Fold Belt. PhD, La Trobe, Melbourne.

Andréasson, P.-G., Svenningsen, O.M., Albrecht, L., 1998. Dawn of Phanerozoic orogeny in the North Atlantic tract; Evidence from the Seve-Kalak Superterrane, Scandinavian Caledonides. GFF 120, 159-172. doi:10.1080/11035899801202159.

Arne, D.C., Bierlein, F.P., McNaughton, N., Wilson, C.J.L., Morand, V.J., 1998. Timing of gold mineralisation in western and central Victoria, Australia: New constraints from SHRIMP II analysis of zircon grains from felsic intrusive rocks. Ore Geology Reviews 13, 251-273. doi:http://dx.doi.org/10.1016/S0169-1368(97)00021-8.

Baillie, P.W., Corbett, K.D., 1985. Sheet 57(7913N) Strahan Explanatory report, geological atlas 1:50,000 series. Geological Survey of Tasmania, Hobart, p. 75.

Balfour, J.C.M., 1968. Petroleum in Victoria - History and Present Status. APEA Journal 8, 16-20.

Balgord, E.A., Yonkee, W.A., Link, P.K., Fanning, C.M., 2013. Stratigraphic, geochronologic, and geochemical record of the Cryogenian Perry Canyon Formation, northern Utah: Implications for Rodinia rifting and snowball Earth glaciation. Geological Society of America Bulletin 125, 1442-1467. doi:10.1130/b30860.1.

Barton, T.J., 1999. Crustal structure of northern Tasmania. M.Sc., Monash University, Melbourne.

Berkman, D.A., 2001. Field geologists' manual, 4th ed. Australasian Institute of Mining and Metallurgy, Melbourne. 395 pages.

Bernecker, T., Moore, D.H., 2003. Linking basement and basin fill: implications for hydrocarbon prospectivity in the Otway Basin region. The APPEA Journal 43/1, 39-58.

Berry, R.F., 2015. Late Mesozoic strike-slip faulting in Tasmania. Australian Journal of Earth Sciences, 1-9. doi:10.1080/08120099.2015.1025837.

Berry, R.F., Bull, S.W., 2012. The pre-Carboniferous geology of Tasmania. Episodes International Geoscience Newsmagazine 35, 195-204.

Berry, R.F., Chmielowski, R.M., Steele, D.A., Meffre, S., 2007. Chemical U–Th–Pb monazite dating of the Cambrian Tyennan Orogeny, Tasmania. Australian Journal of Earth Sciences 54, 757-771. doi:10.1080/08120090701305269.

Berry, R.F., Crawford, A.J., 1988. The tectonic significance of Cambrian allochthonous mafic-ultramafic complexes in Tasmania. Australian Journal of Earth Sciences 35, 523-533. doi:10.1080/08120098808729467.

Berry, R.F., Gray, D.R., 2001. The structure of the coastal section from Goat Island to Ulverstone, northwestern Tasmania. Geological Society of Australia Specialist Group in Tectonics and Structural Geology Field Guide 10, Centre for Ore Deposit Research and School of Earth Sciences, University of Tasmania, and Mineral Resources Tasmania. 27 pages.

Berry, R.F., Holm, O.H., Steele, D.A., 2005. Chemical U-Th-Pb monazite dating and the Proterozoic history of King Island, southeast Australia. Australian Journal of Earth Sciences 52, 461-471. doi:10.1080/08120090500134514.

Berry, R.F., Jenner, G.A., Meffre, S., Tubrett, M.N., 2001. A North American provenance for Neoproterozoic to Cambrian sandstones in Tasmania? Earth and Planetary Science Letters 192, 207-222. doi:10.1016/S0012-821X(01)00436-8.

Berry, R.F., Meffre, S., Kreuzer, H., 1997. Metamorphic rocks from the southern margin of Tasmania and their tectonic significance. Australian Journal of Earth Sciences 44, 609-619. doi:10.1080/08120099708728340.

Berry, R.F., Steele, D.A., Meffre, S., 2008. Proterozoic metamorphism in Tasmania: implications for tectonic reconstructions. Precambrian Research 166, 387-396. doi:10.1016/j.precamres.2007.05.004.

Betts, P.G., Giles, D., Lister, G.S., Frick, L.R., 2002. Evolution of the Australian lithosphere. Australian Journal of Earth Sciences 49, 661-695. doi:http://dx.doi.org/10.1046/j.1440-0952.2002.00948.x.

Betts, P.G., Giles, D., Schaefer, B.F., 2008. Comparing 1800–1600 Ma accretionary and basin processes in Australia and Laurentia: possible geographic connections in Columbia. Precambrian Research 166, 81-92. doi:http://dx.doi.org/10.1016/j.precamres.2007.03.007.

Bierlein, F.P., Arne, D.C., Foster, D.A., Reynolds, P., 2001a. A geochronological framework for orogenic gold mineralisation in central Victoria. Mineralium Deposita 36, 741-746.

Bierlein, F.P., Arne, D.C., Keay, S.M., McNaughton, N.J., 2001b. Timing relationships between felsic magmatism and mineralisation in the central Victorian gold province, southeast Australia. Australian Journal of Earth Sciences 48, 883-899.

Bierlein, F.P., Foster, D.A., Gray, D.R., Davidson, G.J., 2005. Timing of orogenic gold mineralisation in northeastern Tasmania: implications for the tectonic and metallogenetic evolution of Palaeozoic SE Australia. Mineralium Deposita 39, 890-903.

Black, L.P., 2007. OZCHRON SHRIMP U-Pb Geochronology Interim Data Release, July 2007 edition. Geoscience Australia, Canberra.

Black, L.P., Calver, C.R., Seymour, D.B., Reed, A., 2004. SHRIMP U-Pb detrital zircon ages from Proterozoic and Early Palaeozoic sandstones and their bearing on the early geological evolution of Tasmania. Australian Journal of Earth Sciences 51, 885-900. doi:10.1111/j.1400-0952.2004.01091.x.

Black, L.P., Everard, J.L., McClenaghan, M.P., Korsch, R.J., Calver, C.R., Fioretti, A.M., Brown, A.V., Foudoulis, C., 2010. Controls on Devonian-Carboniferous magmatism in Tasmania, based on inherited zircon age patterns, Sr, Nd and Pb isotopes, and major and trace element geochemistry. Australian Journal of Earth Sciences 57, 933-968. doi:10.1080/08120099.2010.509407.

Black, L.P., McClenaghan, M.P., Korsch, R.J., Everard, J.L., Foudoulis, C., 2005. Significance of Devonian-Carboniferous igneous activity in Tasmania as derived from U-Pb SHRIMP dating of zircon. Australian Journal of Earth Sciences 52, 807-829. doi:10.1080/08120090500304232.

Black, L.P., Seymour, D.B., Corbett, K.D., Cox, S.E., Streit, J.E., Bottrill, R.S., Calver, C.R., Everard, J.L., Green, G.R., McClenaghan, M.P., Pemberton, J., Taheri, J., Turner, N.J., 1997. Dating Tasmania's oldest geological events, Record, 1997/15. Australian Geological Survey Organisation, Canberra, 57 pages.

Blackney, P.C.J., 1982. The petrology and conditions of metamorphism of the Fitzmaurice Bay to Stokes Point area, King Island. BSc(Hons), University of Tasmania, Hobart.

Blevin, J.E., Boreham, C.J., Lang, S.C., Mitchell, C.H., Nicholson, C.J., Trigg, K.R., Webster, M.A., 2003. Bass Basin GIS project, Western Tasmanian Minerals Program. Geoscience Australia and Mineral Resources Tasmania, Canberra.

Bodorkos, S., Clark, D.J., 2004. Evolution of a crustal-scale transpressive shear zone in the Albany–Fraser Orogen, SW Australia: 2. Tectonic history of the Coramup Gneiss and a kinematic framework for Mesoproterozoic collision of the West Australian and Mawson cratons. Journal of Metamorphic Geology 22, 713-731. doi:10.1111/j.1525-1314.2004.00544.x.

Bogdanova, S.V., Bingen, B., Gorbatschev, R., Kheraskova, T.N., Kozlov, V.I., Puchkov, V.N., Volozh, Y.A., 2008. The East European Craton (Baltica) before and during the assembly of Rodinia. Precambrian Research 160, 23-45. doi:http://dx.doi.org/10.1016/j.precamres.2007.04.024.

Boger, S.D., 2011. Antarctica — Before and after Gondwana. Gondwana Research 19, 335-371. doi:10.1016/j.gr.2010.09.003.

Boger, S.D., Miller, J.M., 2004. Terminal suturing of Gondwana and the onset of the Ross-Delamerian Orogeny: the cause and effect of an Early Cambrian reconfiguration of plate motions. Earth and Planetary Science Letters 219, 35-48. doi:10.1016/S0012-821X(03)00692-7.

Borg, S.G., DePaolo, D.J., 1991. A tectonic model of the Antarctic Gondwana margin with implications for southeastern Australia: isotopic and geochemical evidence. Tectonophysics 196, 339-358. doi:http://dx.doi.org/10.1016/0040-1951(91)90329-Q.

Borg, S.G., DePaolo, D.J., 1994. Laurentia, Australia, and Antarctica as a Late Proterozoic supercontinent: Constraints from isotopic mapping. Geology 22, 307-310. doi:10.1130/0091-7613(1994)022<0307:LAAAAA>2.3.CO;2.

Bottrill, R.S., Taheri, J., 2007. Petrology of the host rocks, including mineralisation and adjacent rock sequences, from the Savage River mine, Record, 2007/05. Tasmanian Geological Survey, Hobart, 54 pages.

Boyer, S.E., Elliott, D., 1982. Thrust systems. American Association of Petroleum Geologists Bulletin 66, 1196-1230.

Bright, R.M., Amato, J.M., Denyszyn, S.W., Ernst, R.E., 2014. U-Pb geochronology of 1.1 Ga diabase in the southwestern United States: Testing models for the origin of a post-Grenville large igneous province. Lithosphere 6, 135-156. doi:10.1130/l335.1.

Briguglio, D., Kowalczyk, J., Stilwell, J.D., Hall, M., Coffa, A., 2013. Detailed paleogeographic evolution of the Bass Basin: Late Cretaceous to present. Australian Journal of Earth Sciences, 1-16. doi:10.1080/08120099.2013.826282.

Brown, A.V., 1986. Geology of the Dundas-Mt Lindsay-Mt Youngbuck region, Bulletin, 62. Tasmania Department of Mines, Hobart, 221 pages.

Brown, A.V., 2011. Geological atlas 1:50 000 series. Sheet 78 (7912S). Montgomery, Explanatory Report. Geological Survey Tasmania, Hobart, 107 pages.

Brown, A.V., Calver, C.R., Clarke, M.J., Corbett, K.D., Everard, J.L., Forsyth, S.M., Goscombe, B.A., Green, D.C., Green, G.R., McClenaghan, M.P., Pemberton, J., Seymour, D.B., Vicary, M., Woolward, I.R., Worthing, M.A., 2012. Geology of Tasmania. Geological Atlas 1:500,000 digital series, 2012.1 ed. Mineral Resources Tasmania.

Brown, A.V., McClenaghan, M.P., Turner, N.J., Baillie, P.W., McClenaghan, J., Calver, C.R., 1989. Geological atlas 1:50 000 series. Sheet 73 (8112N). Huntley. Geological Survey, Tasmania Department of Mines, Hobart, 109 pages.

Burns, K.L., 1964. Geological Survey explanatory report, one mile geological map series, k55-6-29, Devonport. Government Printer, Tasmania, Hobart. 266 pages.

Burrett, C.F., Berry, R.F., 2001. Proterozoic Australia-Western United States (AUSWUS) fit between Laurentia and Australia. Geology 28, 103-106. doi:10.1130/0091-7613(2000)28<103:PAUSAF>2.0.CO;2.

Burrett, C.F., Martin, E.L., 1989. Geology and mineral resources of Tasmania. Geological Society of Australia Inc. Special Publication 15. 574 pages.

Bushby, Y., 2002. The petrology of the Cambrian greenstones on Phillip Island. BSc(Hons), University of Melbourne, Melbourne.

Calver, C.R., 1998. Isotope stratigraphy of the Neoproterozoic Togari Group, Tasmania. Australian Journal of Earth Sciences 45, 865-874. doi:10.1080/08120099808728441.

Calver, C.R., 2004. Geology of King Island. Geological Society of Australia, Field Guide, 17th Australian Geological Convention, Hobart, B3. 25 pages.

Calver, C.R., 2007. Some notes on the geology of King Island, Record, 2007/02. Mineral Resources Tasmania, Hobart, 20 pages.

Calver, C.R., 2008. Digital Geological Atlas Sheet 2456. Grassy. 1:25 000 Scale Series. Mineral Resources Tasmania, Hobart.

Calver, C.R., 2009. Setting and geochronology of terminal Cryogenian glacials, Grassy Group, King Island, In: Direen, N.G., Calver, C.R., McNeill, A.W. (Eds.), Tungsten, fire and ice in the realm of the ancient king. Geological Society of Australia & Mineral Resources Tasmania, Hobart, pp. 15-17.

Calver, C.R., 2011. Chapter 64 Neoproterozoic glacial deposits of Tasmania. Geological Society, London, Memoirs 36, 649-658. doi:10.1144/m36.64.

Calver, C.R., 2012. Explanatory report for the Grassy and Naracoopa geological map sheets, Explanatory Report 1:25 000 Scale Digital Geological Map Series. Mineral Resources Tasmania, Hobart, 71 pages.

Calver, C.R., Black, L.P., Everard, J.L., Seymour, D.B., 2004. U-Pb zircon age constraints on late Neoproterozoic glaciation in Tasmania. Geology 32, 893-896. doi:10.1130/G20713.1.

Calver, C.R., Crowley, J.L., Wingate, M.T.D., Evans, D.A.D., Raub, T.D., Schmitz, M.D., 2013a. Globally synchronous Marinoan deglaciation indicated by U-Pb geochronology of the Cottons Breccia, Tasmania, Australia. Geology 41, 1127-1130. doi:10.1130/g34568.1.

Calver, C.R., Everard, J.L., 2014. The Surprise Bay Formation (~1300 Ma) and related rocks of western King Island, Record, 2014/01. Tasmanian Geological Survey, Hobart Tasmania, 33 pages.

Calver, C.R., Everard, J.L., Berry, R.F., Bottrill, R.S., Seymour, D.B., 2014. Proterozoic Tasmania, In: Corbett, K.D., Quilty, P.G., Calver, C.R. (Eds.), Geological evolution of Tasmania. Geological Society of Australia Special Publication 24, pp. 34-94.

Calver, C.R., Forsyth, S.M., Everard, J.L., 2006. Geology of the Maydena, Skeleton, Nevada, Weld and Picton 1:25 000 scale map sheets, Tasmanian Geological Survey Record, 2006/04. Mineral Resources Tasmania, Hobart Tas., 126 pages.

Calver, C.R., Grey, K., Laan, M., 2010. The 'string of beads' fossil (Horodyskia) in the mid-Proterozoic of Tasmania. Precambrian Research 180, 18-25. doi:10.1016/j.precamres.2010.02.005.

Calver, C.R., Meffre, S., Everard, J.L., 2013b. Felsic porphyry sills in Surprise Bay Formation near Currie, King Island, dated at ~775 Ma (LA-ICPMS, U-Pb on zircon), Record, 2013/04. Tasmanian Geological Survey, Hobart Tas., 10 pages.

Calver, C.R., Reed, A., 2001. Port Sorell Formation, In: Reed, A.R. (Ed.), Structure and setting of Proterozoic and Palaeozoic rocks in the Tamar region, northern Tasmania. Geological Society of Australia Specialist Group in Tectonics and Structural Geology Field Guide 9, Centre for Ore Deposit Research and the School of Earth Sciences University of Tasmania, and Mineral Resources Tasmania, pp. 91-97.

Calver, C.R., Walter, M.R., 2000. The Late Neoproterozoic Grassy Group of King Island, Tasmania: correlation and palaeogeographic significance. Precambrian Research 100, 299-312. doi:10.1016/S0301-9268(99)00078-9.

Casquet, C., Rapela, C.W., Pankhurst, R.J., Baldo, E.G., Galindo, C., Fanning, C.M., Dahlquist, J.A., Saavedra, J., 2012. A history of Proterozoic terranes in southern South America: From Rodinia to Gondwana. Geoscience Frontiers 3, 137-145. doi:10.1016/j.gsf.2011.11.004.

Cawood, P.A., 2005. Terra Australis Orogen: Rodinia breakup and development of the Pacific and Iapetus margins of Gondwana during the Neoproterozoic and Paleozoic. Earth Science Reviews 69, 249-279. doi:10.1016/j.earscirev.2004.09.001.

Cawood, P.A., Buchan, C., 2007. Linking accretionary orogenesis with supercontinent assembly. Earth-Science Reviews 82, 217-256. doi:http://dx.doi.org/10.1016/j.earscirev.2007.03.003.

Cawood, P.A., Kröner, A., Collins, W.J., Kusky, T.M., Mooney, W.D., Windley, B.F., 2009. Accretionary orogens through Earth history, In: Cawood, P.A., Kröner, A. (Eds.), Earth Accretionary Systems in Space and Time. The Geological Society, London, Special Publication 318, pp. 1–36. doi:10.1144/SP318.1.

Cawood, P.A., Wang, Y., Xu, Y., Zhao, G., 2013. Locating South China in Rodinia and Gondwana: a fragment of greater India lithosphere? Geology 41, 903-906. doi:10.1130/g34395.1.

Cayley, R.A., 2011. Exotic crustal block accretion to the eastern Gondwanaland margin in the Late Cambrian–Tasmania, the Selwyn Block, and implications for the Cambrian–Silurian evolution of the Ross, Delamerian and Lachlan orogens. Gondwana Research 19, 628-649. doi:10.1016/j.gr.2010.11.013.

Cayley, R.A., 2012. Oroclinal folding in the Lachlan Fold Belt: consequence of southeast-directed Siluro-Devonian subduction rollback superimposed on an accreted Ordovician arc assemblage in eastern Australia, In: White, S. (Ed.), The great east-Australian arm-wave of 2012: tectonics, modern analogues/events, mineralisation (GAGA 2012). Selwyn Symposium of the GSA Victoria Division, Geological Society of Australia Extended Abstracts No 103, pp. 34-43.

Cayley, R.A., Korsch, R.J., Moore, D.H., Costelloe, R.D., Nakamura, A., Willman, C.E., Rawling, T.J., Morand, V.J., Skladzien, P.B., O'Shea, P.J., 2011. Crustal architecture of Central Victoria: results from the 2006 deep crustal reflection seismic survey. Australian Journal of Earth Sciences 58, 113-156. doi:10.1080/08120099.2011.543151.

Cayley, R.A., Taylor, D.H., 2001. Ararat 1:100,000 map area geological report, Report, 115. Geological Survey of Victoria, Melbourne Vic., 324 pages.

Cayley, R.A., Taylor, D.H., VandenBerg, A.H.M., Moore, D.H., 2002. Proterozoic – Early Palaeozoic rocks and the Tyennan Orogeny in central Victoria: the Selwyn Block and its tectonic implications. Australian Journal of Earth Sciences 49, 225–254. doi:10.1046/j.1440-0952.2002.00921.x.

Champion, D.C., 2013. Neodymium depleted mantle model age map of Australia: explanatory notes and user guide, Record, 2013/44. Geoscience Australia, Canberra, 209 pages. doi:10.11636/Record.2013.044.

Champion, D.C., Kositcin, N., Huston, D.L., Mathews, E., Brown, C., 2009. Geodynamic synthesis of the Phanerozoic of eastern Australia and implications for metallogeny, Record, 2009/18. Geoscience Australia, Canberra A.C.T., 255 pages.

Chappell, B., White, A., 1974. Two contrasting granite types, Pacific Geology, pp. 173-174.

Chappell, B.W., White, A.J.R., Hine, R., 1988. Granite provinces and basement terranes in the Lachlan Fold Belt, southeastern Australia. Australian Journal of Earth Sciences 35, 505-521. doi:10.1080/08120098808729466.

Chmielowski, R., Berry, R.F., 2012. The Cambrian metamorphic history of Tasmania: the metapelites. Australian Journal of Earth Sciences 59, 1007-1019. doi:10.1080/08120099.2012.724449.

Chmielowski, R.M., 2009. The Cambrian metamorphic history of Tasmania. thesis, University of Tasmania, Hobart, 162 pages.

Clarke, D.A., Emerson, D.W., 1991. Notes on rock magnetization characteristics in applied geophysical studies. Exploration Geophysics 22, 547-555. doi:10.1071/EG991547

Clarke, M.J., Farmer, N., 1983. A diamond drill hole at The Quoin. south-east of Ross, Unpublished report, 1983/18. Mineral Resources Tasmania, Hobart, 5 pages.

Clemens, J.D., Benn, K., 2010. Anatomy, emplacement and evolution of a shallow-level, post-tectonic laccolith: the Mt Disappointment pluton, SE Australia. Journal of the Geological Society, London 167, 915-941. doi:http://dx.doi.org/10.1144/0016-76492009-120.

Clemens, J.D., Bezuidenhout, A., 2014. Origins of co-existing diverse magmas in a felsic pluton: the Lysterfield Granodiorite, Australia. Contributions to Mineralogy and Petrology 167, 1-23. doi:10.1007/s00410-014-0991-9.

Clemens, J.D., Birch, W.D., 2012. Assembly of a zoned volcanic magma chamber from multiple magma batches: The Cerberean Cauldron, Marysville Igneous Complex, Australia. Lithos 155, 272-288. doi:http://dx.org/10.1016/j.lithos.2012.09.007.

Clemens, J.D., Birch, W.D., Dudley, R.J., 2011. S-type ignimbrites with polybaric crystallisation histories: the Tolmie Igneous Complex, Central Victoria, Australia. Contributions to Mineralogy and Petrology 162, 1315-1337. doi:http://dx.doi.org/10.1007/s00410-011-0652-1.

Clemens, J.D., Frei, D., Finger, F., 2014. A new precise date for the Tolmie Igneous Complex in northeastern Victoria. Australian Journal of Earth Sciences 61, 951-958. doi:10.1080/08120099.2014.955058.

Clemens, J.D., Phillips, G.N., 2014. Inferring a deep-crustal source terrane from a high-level granitic pluton: the Strathbogie Batholith, Australia. Contributions to Mineralogy and Petrology 168, 1-22. doi:10.1007/s00410-014-1070-y.

Collins, W.J., 2002. Hot orogens, tectonic switching, and creation of continental crust. Geology 30, 535-538. doi:10.1130/0091-7613(2002)030<0535:HOTSAC>2.0.CO;2.

Collins, W.J., 2003. Slab pull, mantle convection, and Pangaean assembly and dispersal. Earth and Planetary Science Letters 205, 225-237. doi:10.1016/S0012-821X(02)01043-9.

Collins, W.J., Vernon, R.H., 1992. Palaeozoic arc growth, deformation and migration across the Lachlan Fold Belt, southeastern Australia. Tectonophysics 214, 381-400. http://dx.doi.org/10.1016/0040-1951(92)90206-L.

Compston, W., 2004. SIMS U-Pb zircon ages for the Upper Devonian Snobs Creek and Cerberean Volcanics from Victoria, with age uncertainty based on UO_2/UO v. UO/U precision. Journal of the Geological Society of London 161, 223-228. doi:10.1144/0016-764902-121.

Condie, K.C., Belousova, E., Griffin, W.L., Sircombe, K.N., 2009. Granitoid events in space and time: constraints from igneous and detrital zircon age spectra. Gondwana Research 15, 228-242. doi:http://dx.doi.org/10.1016/j.gr.2008.06.001.

Coney, P.J., 1989. Structural aspects of suspect terranes and accretionary tectonics in western North America. Journal of Structural Geology 11, 107-125. doi:10.1016/0191-8141(89)90038-2.

Coney, P.J., 1992. The Lachlan belt of eastern Australia and Circum-Pacific tectonic evolution. Tectonophysics 214, 1-24. doi:10.1016/0040-1951(92)90187-B.

Coney, P.J., Edwards, A., Hine, R., Morrison, F., Windrim, D., 1990. The regional tectonics of the Tasman orogenic system, eastern Australia. Journal of Structural Geology 12, 519-543. doi:10.1016/0191-8141(90)90071-6.

Cooper, A.F., Maas, R., Scott, J.M., Barber, A.J.W., 2011. Dating of volcanism and sedimentation in the Skelton Group, Transantarctic Mountains: implications for the Rodinia-Gondwana transition in southern Victoria Land, Antarctica. Geological Society of America Bulletin 123, 681-702. doi:10.1130/B30237.1.

Cooper, G.T., Hill, K.C., 1997. Cross-section balancing and thermochronological analysis of the Mesozoic development of the Eastern Otway Basin. APPEA Journal, 390-414.

Corbett, K.D., 1992. Stratigraphic-volcanic setting of massive sulfide deposits in the Cambrian Mount Read Volcanics, Tasmania. Economic Geology 87, 564-586. doi:0361-0128/92/1326/56.

Corbett, K.D., 2003. A review of geology and exploration in the Macquarie Harbour–Elliott Bay area, South West Tasmania, Record, 2003/4. Tasmanian Geological Survey, Hobart Tas., 34 pages.

Corbett, K.D., Quilty, P.G., Calver, C.R.e., 2014. Geological Evolution of Tasmania. Geological Society of Australia (Tasmania Division). 639 pages.

Crawford, A.J., Berry, R.F., 1992. Tectonic implications of Late Proterozoic-Early Palaeozoic igneous rock associations in western Tasmania. Tectonophysics 214, 37-56. doi:10.1016/0040-1951(92)90189-D.

Crawford, A.J., Cameron, W.E., Keays, R.R., 1984. The association boninite low-Ti andesite-tholeiite in the Heathcote Greenstone Belt, Victoria; ensimatic setting for the early Lachlan Fold Belt. Australian Journal of Earth Sciences 31, 161-175. doi:10.1080/08120098408729290.

Crawford, A.J., Cayley, R.A., Taylor, D.H., Morand, V.J., Gray, C.M., Kemp, A.I.S., Wohlt, K.E., VandenBerg, A.H.M., Moore, D.H., Maher, S., Direen, N.G., Edwards, J., Donaghy, A.G., Anderson, J.A., Black, L.P., 2003a. Neoproterozoic and Cambrian, In: Birch, W.D. (Ed.), Geology of Victoria, 3rd ed. Geological Society of Australia (Victoria Division), Melbourne, pp. 73-93.

Crawford, A.J., Everard, J.L., Bottrill, R.S., 2014. Mafic-ultramafic complexes and associated rocks, In: Corbett, K.D., Quilty, P.G., Calver, C.R. (Eds.), Geological evolution of Tasmania. Geological Society of Australia (Tasmania Division), Hobart, Tasmania, pp. 120-141.

Crawford, A.J., Meffre, S., Symonds, P.A., 2003b. 120 to 0 Ma tectonic evolution of the southwest Pacific and analogous geological evolution of the 600 to 220 Ma Tasman Fold Belt System, In: Hillis, R.R., Müeller, D.R. (Eds.), The evolution and dynamics of the Australian Plate. Geological Society of Australia Special Publication 22 and Geological Society of America Special Paper 372, Sydney N.S.W., pp. 383-403. doi:10.1130/0-8137-2372-8.383.

Crook, K.A.W., 1979. Tectonic implications of some field relations of the Adelaidean Cooee Dolerite, Tasmania. Australian Journal of Earth Sciences 26, 353-361. doi:10.1080/00167617908729102.

Dalziel, I.W.D., 1997. Neoproterozoic-Paleozoic geography and tectonics: review, hypothesis, environmental speculation. Geological Society of America Bulletin 109, 16-42. doi:10.1130/0016-7606(1997)109<0016:ONPGAT>2.3.CO;2.

David, T.W.E., 1950. The Geology of the Commonwealth of Australia (Edited and Much Supplemented by W.R. Browne)(3 vols). Edward Arnold and Co, London.

Dentith, M., Mudge, S.T., 2014. Geophysics for the mineral exploration geologist. Cambridge University Press, Cambridge, UK. 438 pages.

Direen, N.G., Crawford, A.J., 2003a. Fossil seaward-dipping reflector sequences preserved in southeastern Australia: a 600 Ma volcanic passive margin in eastern Gondwanaland. Journal of the Geological Society, London 160, 985-990. doi:10.1144/0016-764903-010

Direen, N.G., Crawford, A.J., 2003b. The Tasman Line: where is it, what is it, and is it Australia's Rodinian breakup boundary? Australian Journal of Earth Sciences 50, 491-502. doi:10.1046/j.1440-0952.2003.01005.x.

Direen, N.G., Stagg, H.M.J., Symonds, P.A., Norton, I.O., 2012. Variations in rift symmetry: cautionary examples from the Southern Rift System (Australia–Antarctica), In: Mohriak, W.U., Danforth, A., Post, P.J., Brown, D.E., Tari, G.C., Nemčok, M., Sinha, S.T. (Eds.), Geological Society, London, Special Publication 369. doi: 10.1144/sp369.4

Driscoll, J., 2006. Geothermal prospectivity of onshore Victoria, 1st ed. Department of Primary Industries, Melbourne. 48 pages.

Drummond, B.J., Barton, T.J., Korsch, R.J., Rawlinson, N., Yeates, A.N., Collins, C.D.N., Brown, A.V., 2000. Evidence for crustal extension and inversion in eastern Tasmania, Australia, during the Neoproterozoic and Early Palaeozoic. Tectonophysics 329, 1-21. doi:10.1016/S0040-1951(00)00185-2.

Duddy, I., 2003. Mesozoic - a time of change in tectonic regime, In: Birch, W.D. (Ed.), Geology of Victoria, 3rd ed. Geological Society of Australia (Victoria Division), Melbourne, pp. 239-287.

Elburg, M.A., 1996a. Evidence of isotopic equilibration between microgranitoid enclaves and host granodiorite, Warburton Granodiorite, Lachlan Fold Belt, Australia. Lithos 38, 1-22. doi:http://dx.doi.org/10.1016/0024-4937(96)00003-5.

Elburg, M.A., 1996b. U-Pb ages and morphologies of zircon in microgranitoid enclaves and peraluminous host granite: evidence for magma mingling. Contributions to Mineralogy and Petrology 123, 177-189. doi:10.1007/s004100050149.

Elliott, C.G., Gray, D.R., 1992. Correlations between Tasmania and the Tasman-Transantarctic orogen: evidence for easterly derivation of Tasmania relative to mainland Australia. Geology 20, 621-625. doi:10.1130/0091-7613(1992)020<0621:CBTATT>2.3.CO;2.

Elliott, C.G., Woodward, N.B., Gray, D.R., 1993. Complex regional fault history of the Badger Head region, northern Tasmania. Australian Journal of Earth Sciences 40, 155-168. doi:10.1080/08120099308728072.

Encarnación, J., Grunow, A., 1996. Changing magmatic and tectonic styles along the paleo-Pacific margin of Gondwana and the onset of early Paleozoic magmatism in Antarctica. Tectonics 15, 1325-1341. doi:10.1029/96TC01484.

Eshaghi, E., M. Reading, A., Roach, M., J. Cracknell, M., Bombardieri, D., Duffett, M., 2015. Insights into the continental structure of southeast Australia and Tasmania from passive seismic and magnetic datasets. ASEG Extended Abstracts 2015, 1-4. doi:http://dx.doi.org/10.1071/ASEG2015ab203.

Esso Exploration Austraila Inc., 1967. Bass 3 Tasmania well completion report, Unpublished report, 330001. Mineral Resources Tasmania, 131 pages.

Esso Exploration and Production Australia Inc., 1970. Whelk No 1, raw data report, Unpublished company report. Department of Infrastructure, Energy and Resources, 101 pages.

Evans, D.A.D., 2009. The palaeomagnetically viable, long-lived and all-inclusive Rodinia supercontinent reconstruction. Geological Society, London, Special Publications 327, 371-404. doi:http://dx.doi.org/10.1144/SP327.16.

Evans, D.A.D., 2013. Reconstructing pre-Pangean supercontinents. Geological Society of America Bulletin 125, 1735-1751. doi:10.1130/b30950.1.

Everard, J.L., Calver, C.R., 2014. Early Cambrian allochthonous and parautochthonous sequences, In: Corbett, K.D., Quilty, P.G., Calver, C.R. (Eds.), Geological evolution of Tasmania. Geological Society of Australia (Tasmania Division) Special Publication 24, pp. 110-120.

Everard, J.L., Calver, C.R., Pemberton, J., Taheri, J., Dixon, G., Quilty, P.G., 1997. Geology of the islands of southwestern Bass Strait, Record, 1997/03. Hobart, 62 pages.

Everard, J.L., Seymour, D.B., Reed, A.R., McClenaghan, M.P., Green, D.C., Calver, C.R., 2007. Regional geology of the southern Smithton Synclinorium explanatory report for the Roger, Sumac and Dempster 1:25,000 scale geological map sheets, far northwestern Tasmania, 1:25,000 Scale Digital Geological Map Series Explanatory Report, 2. Department of Infrastructure, Energy and Resources, Rosny Park, 237 pages.

Exon, N.F., Berry, R.F., Crawford, A.J., Hill, P.J., 1997a. Geological evolution of the East Tasman Plateau, a continental fragment southeast of Tasmania. Australian Journal of Earth Sciences 44, 597-608. doi:10.1080/08120099708728339.

Exon, N.F., Marshall, J.F., McCorkle, D.C., Alcock, M., Chaproniere, G.C.H., Connell, R., Dutton, S.J., Elmes, M., Findlay, C., Robertson, L., Rollet, N., Samson, C., Shafik, S., Whitmore, G.P., 1995. AGSO Cruise 147 report — Tasman Rises geological sampling cruise of Rig Seismic: stratigraphy, tectonic history and palaeoclimate of the offshore Tasmanian region, AGSO Record 1995/56. Australian Geological Survey Organisation, Canberra, 159 pages.

Exon, N.F., Moore, A.M.G., Hill, P.J., 1997b. Geological framework of the South Tasman Rise, south of Tasmania, and its sedimentary basins. Australian Journal of Earth Sciences 44, 561-577. doi:10.1080/08120099708728337.

Fanning, C.M., Morand, V.J., 2002. Results of some SHRIMP U-Pb zircon dating of rocks from Victoria, Technical Record, 2002/1. Geological Survey of Victoria, 73 pages.

Federico, L., Capponi, G., Crispini, L., 2006. The Ross orogeny of the Transantarctic Mountains: a northern Victoria Land perspective. International Journal of Earth Sciences : Geologische Rundschau 95, 759-770. doi:10.1007/s00531-005-00.

Fergusson, C.L., 1987. Early Palaeozoic back-arc deformation in the Lachlan Fold Belt, southeastern Australia: implications for terrane translations in eastern Gondwanaland, In: Leitch, E.C., Scheibner, E. (Eds.), Terrane Accretion and Orogenic Belts. American Geophysical Union, pp. 39-56.

Fergusson, C.L., Carr, P.F., Fanning, C.M., Green, T.J., 2001. Proterozoic-Cambrian detrital zircon and monazite ages from the Anakie Inlier, central Queensland: Grenville and Pacific-Gondwana signatures. Australian Journal of Earth Sciences 48, 857-866. doi:http://dx.doi.org/10.1046/j.1440-0952.2001.00904.x.

Fergusson, C.L., Gray, D.R., Cas, R.A.F., 1986. Overthrust terranes in the Lachlan fold belt, southeastern Australia. Geology 14, 519-522. doi:http://dx.doi.org/10.1130/0091-7613(1986)14<519:OTITLF>2.0.CO;2.

Fergusson, C.L., Nutman, A.P., Kamiichi, T., Hidaka, H., 2013. Evolution of a Cambrian active continental margin: the Delamerian–Lachlan connection in southeastern Australia from a zircon perspective. Gondwana Research 24, 1051-1066. doi:http://dx.doi.org/10.1016/j.gr.2013.03.006.

Finlayson, D.M., Johnstone, D.W., Owen, A.J., Wake-Dyster, K.D., 1996. Deep seismic images and the tectonic framework of early rifting in the Otway Basin, Australian southern margin. Tectonophysics 264, 137-152. doi:10.1016/S0040-1951(96)00123-0.

Finn, C., Moore, D.H., Damaske, D., Mackey, T., 1999. Aeromagnetic legacy of early Paleozoic subduction along the Pacific margin of Gondwana. Geology 27, 1087-1090. doi:10.1130/0091-7613(1999)027<1087:ALOEPS>2.3.CO;2.

Finn, C.A., Goodge, J.W., Damaske, D., Fanning, C.M., 2006. Scouting craton's edge in Paleo-Pacific Gondwana, In: Fütterer, D.K., Damaske, D., Kleinschmidt, G., Miller, H., Tessensohn, F. (Eds.), Antarctica Contributions to Global Earth Sciences Proceedings of the IX International Symposium of Antarctic Earth Sciences Potsdam 2003. Springer, Heildelberg Berlin, pp. 165-174.

Fioretti, A.M., Black, L.P., Foden, J., Visonà, D., 2005a. Grenville-age magmatism at the South Tasman Rise (Australia): a new piercing point for the reconstruction of Rodinia. Geology 33, 769-772. doi:10.1130/G21671.1.

Fioretti, A.M., Capponi, G., Black, L.P., Varne, R., Visona, D., 2005b. Surgeon island granite SHRIMP zircon ages: a clue for the Cambrian tectonic setting and evolution of the Palaeopacific margin of Gondwana (northern Victoria Land, Antarctica). Terra Nova 17, 242-249. doi:10.1111/j.1365-3121.2005.00606.x.

Fitzsimons, I.C.W., 2003. Proterozoic basement provinces of southern and southwestern Australia, and their correlation with Antarctica, In: Yoshida, M., Windley, B.E., Dasgupta, S. (Eds.), Proterozoic East Gondwana: Supercontinent Assembly and Breakup. The Geological Society of London, pp. 93-130. doi: 10.1144/GSL.SP.2003.206.01.07

Flöttmann, T., Haines, P., Jago, J., James, P., Belperio, A., Gum, J., 1998. Formation and reactivation of the Cambrian Kanmantoo Trough, SE Australia: implications for early Palaeozoic tectonics at eastern Gondwana's plate margin. Journal of the Geological Society of London 155, 101-115. doi:10.1144/gsjgs.155.3.0525.

Foden, J., Elburg, M.A., Dougherty-Page, J., Burtt, A., 2006. The timing and duration of the Delamerian Orogeny: correlation with the Ross Orogen and implications for Gondwana assembly. The Journal of Geology 114, 189-210. doi:10.1086/499570.

Foden, J., Sandiford, M., Dougherty-Page, J., Williams, I., 1999. Geochemistry and geochronology of the Rathjen Gneiss: implications for the early tectonic evolution of the Delamerian Orogen. Australian Journal of Earth Sciences 46, 377-389. doi:10.1046/j.1440-0952.1999.00712.x.

Foster, D.A., Gray, D.R., 2007. Strain rate in Paleozoic thrust sheets, the western Lachlan Orogen, Australia: strain analysis and fabric geochronology, In: Sears, J.W., Harms, T.A., Evenchick, C.A. (Eds.), Whence the mountains? Inquiries into the evolution of orogenic systems: a volume in honor of Raymond A. Price. Geological Society Of America Special Paper, pp. 349–368. doi:10.1130/2007.2433(17).

Foster, D.A., Gray, D.R., Bucher, M., 1999. Chronology of deformation within the turbidite-dominated, Lachlan orogen: implications for the tectonic evolution of eastern Australia and Gondwana. Tectonics 18, 452-485. doi:http://dx.doi.org/0278-7407/99/1998TC900031.

Foster, D.A., Gray, D.R., Spaggiari, C., 2005. Timing of subduction and exhumation along the Cambrian East Gondwana margin, and the formation of Paleozoic backarc basins. Geological Society of America Bulletin 117, 105-116. doi:10.1130/B25481.1.

Fu, B., Fairmaid, A., Phillips, D., 2009. New $^{40}Ar/^{39}Ar$ geochronology of gold deposits in central Victoria: final report, Gold Undercover Report, 19. Department of Primary Industries, Victoria, 29 pages.

Furr, B.C., 1986. Chat No. 1 well completion report; Bridge Oil Ltd, Unpublished open file report. Mineral Resources Tasmania, Hobart, Tas, 284 pages.

Gee, R.D., Legge, P.J., 1979. Geological Atlas 1:63 360 series, sheet 30 (8215N) Beaconsfield map and explanatory notes. 2nd ed. Tasmania Department of Mines, Hobart.

Gibson, G.M., Morse, M.P., Ireland, T.R., Nayak, G.K., 2011. Arc–continent collision and orogenesis in western Tasmanides: insights from reactivated basement structures and formation of an ocean–continent transform boundary off western Tasmania. Gondwana Research 19, 608-627. doi:http://dx.doi.org/10.1016/j.gr.2010.11.020.

Gibson, G.M., Totterdell, J.M., White, L.T., Mitchell, C.H., Stacey, A.R., Morse, M.P., Whitaker, A., 2013. Pre-existing basement structure and its influence on continental rifting and fracture zone development along Australia's southern rifted margin. Journal of the Geological Society 170, 365-377. doi:10.1144/jgs2012-040.

Glen, R.A., 2005. The Tasmanides of eastern Australia, In: Vaughan, A.P.M., Leat, P.Y., Pankhurst, R.J. (Eds.), Terrane Processes at the Margins of Gondwana. Geological Society, London Special Publication 246, London U.K., pp. 23-96. doi: 10.1144/GSL.SP.2005.246.01.02.

Glen, R.A., 2013. Refining accretionary orogen models for the Tasmanides of eastern Australia. Australian Journal of Earth Sciences 60, 315-370. doi:10.1080/08120099.2013.772537.

Godard, G., Palmeri, R., 2013. High-pressure metamorphism in Antarctica from the Proterozoic to the Cenozoic: a review and geodynamic implications. Gondwana Research 23, 844-864. doi:10.1016/j.gr.2012.07.012.

Goodge, J.W., Fanning, C.M., 2010. Composition and age of the East Antarctic Shield in eastern Wilkes Land determined by proxy from Oligocene-Pleistocene glaciomarine sediment and Beacon Supergroup sandstones, Antarctica. Geological Society of America Bulletin 122, 1135-1159. doi:http://dx.doi.org/10.1130/B30079.1.

Goodge, J.W., Fanning, C.M., Bennett, V.C., 2001. U–Pb evidence of ~1.7 Ga crustal tectonism during the Nimrod Orogeny in the Transantarctic Mountains, Antarctica: implications for Proterozoic plate reconstructions. Precambrian Research 112, 261-288. doi:http://dx.doi.org/10.1016/S0301-9268(01)00193-0.

Goodge, J.W., Fanning, C.M., Brecke, D.M., 2010. Continuation of the Laurentian Grenville province across the Ross Sea margin of East Antarctica. Journal of Geology 118, 601-619. doi:10.1086/656385.

Goodge, J.W., Fanning, C.M., Norman, M.D., Bennett, V.C., 2012. Temporal, isotopic and spatial relations of Early Paleozoic Gondwana-margin arc magmatism, central Transantarctic Mountains, Antarctica. Journal of Petrology 53, 2027-2065. doi:10.1093/petrology/egs043.

Goodge, J.W., Myrow, P., Williams, I.S., Bowring, S.A., 2002. Age and provenance of the Beardmore Group, Antarctica: constraints on Rodinia Supercontinent breakup. The Journal of Geology 110, 393-406. doi:10.1086/340629.

Goodge, J.W., Vervoort, J.D., Fanning, C.M., Brecke, D.M., Farmer, G.L., Williams, I.S., Myrow, P.M., DePaolo, D.J., 2008. A positive test of East Antarctica–Laurentia juxtaposition within the Rodinia Supercontinent. Science 321, 235-240. doi:10.1126/science.1159189.

Goodge, J.W., Walker, N.W., Hansen, V.L., 1993. Neoproterozoic-Cambrian basement-involved orogenesis within the Antarctic margin of Gondwana. Geology 21, 37-40. doi:10.1130/0091-7613(1993)021<0037:ncbiow>2.3.co;2.

Goodge, J.W., Williams, I.S., Myrow, P.M., 2004. Provenance of Neoproterozoic and lower Paleozoic siliciclastic rocks of the active-margin sedimentation central Ross orogen, Antarctica: detrital record of rift-, passive-, and active-margin sedimentation. Geological Society of America Bulletin 116, 1253-1279. doi:10.1130/B25347.1.

Goss, B., Lenard, M., 2006. EL 3706 Foster project, Victoria for the period ending 31 March 2006. Expired Exploration Licence Report, Expired Exploration Licence Report File. Minerals and Petroleum Victoria.

Graeber, F.M., Houseman, G.A., Greenhalgh, S.A., 2002. Regional teleseismic tomography of the western Lachlan Orogen and the Newer Volcanic Province, southeast Australia. Geophysical Journal International 149, 249-266. doi:10.1046/j.1365-246X.2002.01598.x.

Grant, A.J., 2002. Petrophysics compilation for Victoria, Unpublished Report. Geological Survey of Victoria, Melbourne, 23 pages.

Gray, C.M., 1990. A strontium isotope traverse across the granitic rocks of southeastern Australia: petrographic and tectonic implications. Australian Journal of Earth Sciences 37, 331-349. doi:10.1080/08120099008727931.

Gray, D.R., Foster, D.A., 1998. Character and kinematics of faults within the turbidite-dominated Lachlan Orogen: implications for tectonic evolution of eastern Australia. Journal of Structural Geology 20, 1691-1720. doi:10.1016/S0191-8141(98)00089-3.

Gray, D.R., Foster, D.A., 2004. Tectonic evolution of the Lachlan Orogen, southeast Australia: historical review, data synthesis and modern perspectives. Australian Journal of Earth Sciences 51, 773-817. doi:10.1111/j.1400-0952.2004.01092.x.

Gray, D.R., Foster, D.A., Gray, C., Cull, J., Gibson, G., 1998. Lithospheric Structure of the Southeast Australian Lachlan Orogen along the Victorian Global Geoscience Transect. International Geology Review 40, 1088-1117. doi:10.1080/00206819809465256.

Gray, D.R., Foster, D.A., Morand, V.J., Willman, C.E., Cayley, R.A., Spaggiari, C.V., Taylor, D.H., Gray, C.M., VandenBerg, A.H.M., Hendrickx, M.A., Wilson, C.J.L., 2003. Structure, metamorphism, geochronology and tectonics of Palaeozoic rocks interpreting a complex, long-lived orogenic system, In: Birch, W.D. (Ed.), The geology of Victoria, Third ed. Geological Society of Australia (Victoria Division), Melbourne, pp. 15-71.

Gray, D.R., Mortimer, L., 1996. Implications of overprinting deformations and fold interference patterns in the Melbourne Zone, Lachlan Fold Belt. Australian Journal of Earth Sciences 43, 103-114. doi:10.1080/08120099608728240.

Gray, D.R., Willman, C.E., 1991. Deformation in the Ballarat Slate Belt, central Victoria, and implications for the crustal structure across southeast Australia. Australian Journal of Earth Sciences 38, 171-201. doi:10.1080/08120099108727964.

Gray, D.R., Willman, C.E., Foster, D.A., 2006. Crust restoration for the western Lachlan Orogen using the strain-reversal, area-balancing technique: implications for crustal components and original thicknesses. Australian Journal of Earth Sciences 53, 329-341. doi:10.1080/08120090500499305.

Greenfield, J.E., Musgrave, R.J., Bruce, M.C., Gilmore, P.J., Mills, K.J., 2011. The Mount Wright Arc: a Cambrian subduction system developed on the continental margin of East Gondwana, Koonenberry Belt, eastern Australia. Gondwana Research 19, 650-669. doi:10.1016/j.gr.2010.11.017.

Hall, M., 1998. The structural history of northern Tasmania and the Bass Strait connection, In: Findlayson, D.M., Jones, L.E.A. (Eds.), Mineral systems and the crust - upper mantle of south eastern Australia, AGSO Record 1998/2, pp. 93-97.

Hall, M., 2001. Structural geology of Hunter, Walker and Robbins Islands and Woolnorth Peninsula, northwest Tasmania, In: Brown, A.V., Drummond, B.J., Korsch, R.J. (Eds.), The geological framework of Tasmania; a workshop to finalise the TASGO and TASMAP Projects under the National Geoscience Mapping Accord. Australian Geological Survey Organisation and Mineral Resources Tasmania, Hobart, 13 pages.

Hall, R., 2002. Cenozoic geological and plate tectonic evolution of SE Asia and the SW Pacific: computer-based reconstructions, model and animations. Journal of Asian Earth Sciences 20, 353-431. doi:http://dx.doi.org/10.1016/S1367-9120(01)00069-4.

Hally, L.A., Paterson, J.R., 2014. Biodiversity, biofacies and biogeography of middle Cambrian (Series 3) arthropods (Trilobita and Agnostida) on the East Gondwana margin. Gondwana Research 26, 654-674. doi:http://dx.doi.org/10.1016/j.gr.2014.04.002.

Halpin, J.A., Jensen, T., McGoldrick, P., Meffre, S., Berry, R.F., Everard, J.L., Calver, C.R., Thompson, J., Goemann, K., Whittaker, J.M., 2014. Authigenic monazite and detrital zircon dating from the Proterozoic Rocky Cape Group, Tasmania: Links to the Belt-Purcell Supergroup, North America. Precambrian Research 250, 50-67. doi:http://dx.doi.org/10.1016/j.precamres.2014.05.025.

Handler, M.R., Bennett, V.C., 2001. Constraining continental structure by integrating Os isotopic ages of lithospheric mantle with geophysical and crustal data: an example from southeastern Australia. Tectonics 20, 177-188. doi:http://dx.doi.org/0278-7407/01/2000TC001196.

Handler, M.R., Bennett, V.C., Esat, T., 1997. The persistence of off-cratonic lithospheric mantle: Os isotopic systematics of variably metasomatised southeast Australian xenoliths. Earth and Planetary Science Letters 151, 61-75. doi:10.1016/S0012-821X(97)00118-0.

Harlan, S.S., Heaman, L., LeCheminant, A.N., Premo, W.R., 2003. Gunbarrel mafic magmatic event: a key 780 Ma time marker for Rodinia plate reconstructions. Geology 31, 1053-1056. doi:10.1130/g19944.1.

Henjes-Kunst, F., Schüssler, U., 2003. Metasedimentary units of the Cambro-Ordovician Ross orogen in northern Victoria Land and Oates Land: Implications for their provenance and geotectonic setting from geochemical and Nd-Sr isotope data. Terra Antartica 10, 105-128.

Henry, D.A., Birch, W.D., 1992. Cambrian greenstone on Phillip Island, Victoria. Australian Journal of Earth Sciences 39, 567-575. doi:10.1080/08120099208728050.

Hergt, J.M., Brauns, C.M., 2001. On the origin of Tasmanian dolerites. Australian Journal of Earth Sciences 48, 543-549. doi:10.1046/j.1440-0952.2001.00875.x.

Hill, K.C., Hill, K.A., Cooper, G.T., O'Sullivan, A.J., O'Sullivan, P.B., Richardson, M.J., 1995. Inversion around the Bass Basin, SE Australia. Geological Society, London, Special Publications 88, 525-547. doi:10.1144/gsl.sp.1995.088.01.27.

Hill, P.J., Webber, K., 1995. Deep crustal seismic survey, circum-Tasmania and south Tasman rise: AGSO Survey 148/159 post-cruise report by P.J. Hill, K. Webber & Survey 148/159 Shipboard Party, AGSO Record, 1995/27. Bureau of Mineral Resources, 25 pages.

Hoffman, P.F., Calver, C.R., Halverson, G.P., 2009. Cottons Breccia of King Island, Tasmania: Glacial or non-glacial, Cryogenian or Ediacaran? Precambrian Research 172, 311-322. doi:10.1016/j.precamres.2009.06.003.

Holm, O.H., Berry, R.F., 2002. Structural history of the Arthur Lineament, northwest Tasmania: an analysis of critical outcrops. Australian Journal of Earth Sciences 49, 167-185. doi:10.1046/j.1440-0952.2002.00918.x.

Holm, O.H., Crawford, A.J., Berry, R.F., 2003. Geochemistry and tectonic settings of meta-igneous rocks in the Arthur Lineament and surrounding area, northwest Tasmania. Australian Journal of Earth Sciences 50, 903-918. doi:10.1111/j.1400-0952.2003.01033.x.

Ireland, T.R., Flöttmann, T., Fanning, C.M., Gibson, G.M., Preiss, W.V., 1998. Development of the early Paleozoic Pacific margin of Gondwana from detrital-zircon ages across the Delamerian orogen. Geology 26, 243-246. doi:0.1130/0091-7613(1998)026<0243:DOTEPP>2.3.CO;2.

Ireland, T.R., Morand, V.J., Gibson, G.M., 2002. Results of some recent SHRIMP U-Pb zircon dating of rocks from the Glenelg Zone of western Victoria, Technical Record, 2002/2. Geological Survey of Victoria, Melbourne, 23 pages.

Isles, D.J., Rankin, L.R., 2013. Geological interpretation of aeromagnetic data. Australian Society of Exploration Geophysicists, Perth, 357 pages.

Jacob, J., Dyment, J., 2014. Early opening of Australia and Antarctica: New inferences and regional consequences. Tectonophysics 636, 244-256. doi:http://dx.doi.org/10.1016/j.tecto.2014.08.020.

Jago, J.B., Brown, A.V., 1989. Middle to Upper Cambrian fossiliferous sedimentary rocks In: Burrett, C.F., Martin, E.L. (Eds.), Geology and mineral resources of Tasmania. Geological Society of Australia, pp. 74-82.

Jenkins, R.J.F., Cooper, J.A., Compston, W., 2002. Age and biostratigraphy of Early Cambrian tuffs from SE Australia and southern China. Journal of the Geological Society of London 159, 645-658. doi:10.1144/0016-764901-127.

Jennings, I.B., 1963. One mile geological map K/55-6-45 Middlesex, Explanatory report. Tasmanian Geological Survey, 161 pages.

Jones, J.V., Connelly, J.N., Karlstrom, K.E., Williams, M.L., Doe, M.F., 2009. Age, provenance, and tectonic setting of Paleoproterozoic quartzite successions in the southwestern United States. Geological Society of America Bulletin 121, 247-264. doi:10.1130/b26351.1.

Jones, J.V., Daniel, C.G., Frei, D., Thrane, K., 2011. Revised regional correlations and tectonic implications of Paleoproterozoic and Mesoproterozoic metasedimentary rocks in northern New Mexico, USA: New findings from detrital zircon studies of the Hondo Group, Vadito Group, and Marqueñas Formation. Geosphere 7, 974-991. doi:10.1130/ges00614.1.

Karlstrom, K.E., Åhäll, K.I., Harlan, S.S., Williams, M.L., McLelland, J., Geissman, J.W., 2001. Long-lived (1.8-1.0 Ga) convergent orogen in southern Laurentia, its extensions to Australia and Baltica, and implications for refining Rodinia. Precambrian Research 111, 5-30. doi:10.1016/S0301-9268(01)00154-1.

Keay, S., Collins, W.J., McCulloch, M.T., 1997. A three-component Sr-Nd isotopic mixing model for granitoid genesis, Lachlan fold belt, eastern Australia. Geology 25, 307-310. doi:10.1130/0091-7613(1997)025<0307:atcsni>2.3.co;2.

Kemp, A.I.S., Hawkesworth, C.J., Paterson, B.A., Foster, G.L., Kinny, P.D., Whitehouse, M.J., Maas, R., EIMF, 2006. Exploring the plutonic–volcanic link: a zircon U–Pb, Lu–Hf and O isotope study of paired volcanic and granitic units from southeastern Australia. Earth and Environmental Science Transactions of the Royal Society of Edinburgh 97, 337-355. doi:doi:10.1017/S0263593300001498.

Kennett, B.L.N., Salmon, M., 2012. AuSREM: Australian Seismological Reference Model. Australian Journal of Earth Sciences 59, 1091-1103. doi:10.1080/08120099.2012.736406.

Kheraskova, T., Volozh, Y.A., Vorontsov, A., Pevzner, L., Sychkin, N., 2002. Sedimentation Conditions at the Central East European Platform in Riphean and Early Vendian. Lithology and Mineral Resources 37, 68-81. doi:10.1023/A:1013639623107.

Kirkland, C.L., Smithies, R.H., Woodhouse, A.J., Howard, H.M., Wingate, M.T.D., Belousova, E.A., Cliff, J.B., Murphy, R.C., Spaggiari, C.V., 2013. Constraints and deception in the isotopic record; the crustal evolution of the west Musgrave Province, central Australia. Gondwana Research 23, 759-781. doi:http://dx.doi.org/10.1016/j.gr.2012.06.001.

Knight, L.A., McDonald, P.A., Frankel, E., Moore, D.H., 1995. A preliminary appraisal of the pre-Tertiary infrabasins beneath the Murray Basin, Victorian Initiative for Minerals and Petroleum Report, 16. Department of Agriculture, Energy and Minerals, Melbourne, 137 pages.

Korsch, R.J., Barton, T.J., Gray, D.R., Owen, A.J., Foster, D.A., 2002. Geological interpretation of a deep seismic reflection transect across the boundary between the Delamerian and Lachlan orogens, in the vicinity of The Grampians, Western Victoria. Australian Journal of Earth Sciences 49, 1057-1075. doi:10.1046/j.1440-0952.2002.00963.x.

Large, R., Doyle, M., Raymond, O., Cooke, D., Jones, A., Heasman, L., 1996. Evaluation of the role of Cambrian granites in the genesis of world class VHMS deposits in Tasmania. Ore Geology Reviews 10, 215-230. doi:10.1016/0169-1368(95)00024-0.

Leaman, D.E., Webster, S.S., 2002. Quantitative interpretation of magnetic and gravity data for the Western Tasmanian Regional Minerals Program, Tasmanian Geological Survey Record, 2002/15. Mineral Resources Tasmania, Hobart Tas., 91 pages.

Li, X.-H., Li, Z.-X., Li, W.-X., 2014. Detrital zircon U–Pb age and Hf isotope constrains on the generation and reworking of Precambrian continental crust in the Cathaysia Block, South China: A synthesis. Gondwana Research 25, 1202-1215. doi:http://dx.doi.org/10.1016/j.gr.2014.01.003.

Li, Z.-X., Evans, D.A.D., Halverson, G.P., 2013. Neoproterozoic glaciations in a revised global palaeogeography from the breakup of Rodinia to the assembly of Gondwanaland. Sedimentary Geology 294, 219-232. doi:http://dx.doi.org/10.1016/j.sedgeo.2013.05.016.

Li, Z.-X., Li, X.-H., Li, W.-X., Ding, S., 2008a. Was Cathaysia part of Proterozoic Laurentia? – new data from Hainan Island, south China. Terra Nova 20, 154-164. doi:10.1111/j.1365-3121.2008.00802.x.

Li, Z.X., 2001. Understanding the Precambrian tectonic events in Tasmania: clues from south China, In: Davidson, G., Pongratz, J. (Eds.), 2001: a structural odessey. Specialist Group in Tectonics & Structural Geology, Geological Society of Australia, Ulverstone, Tasmania, Abstracts 64, pp. 110-111.

Li, Z.X., Baillie, P.W., Powell, C.M., 1997. Relationship between northwestern Tasmania and East Gondwanaland in the Late Cambrian/Early Ordovician: paleomagnetic evidence. Tectonics 16, 161-171. doi:10.1029/96TC02729.

Li, Z.X., Bogdanova, S.V., Collins, A.S., Davidson, A., Waele, B.D., Ernst, R.E., Fitzsimons, I.C.W., Fuck, R.A., Gladkochub, D.P., Jacobs, J., Karlstrom, K.E., Lul, S., Natapovm, L.M., Pease, V., Pisarevsky, S.A., Thrane, K., Vernikovsky, V., 2008b. Assembly, configuration, and break-up history of Rodinia: a synthesis. Precambrian Research 160, 179-210. doi:10.1016/j.precamres.2007.04.021.

Li, Z.X., Evans, D.A.D., 2011. Late Neoproterozoic 40° intraplate rotation within Australia allows for a tighter-fitting and longer-lasting Rodinia. Geology 39, 39–42. doi:10.1130/G31461.1.

Lisker, F., Läufer, A.L., 2013. The Mesozoic Victoria Basin: Vanished link between Antarctica and Australia. Geology 41, 1043-1046. doi:10.1130/g33409.1.

Lister, G.S., Etheridge, M.A., Symonds, P.A., 1991. Detachment models for the formation of passive continental margins. Tectonics 10, 1038-1064. doi:0278-7407/91/90TC-01007.

Lunt, C.K., 1970. Clam 1 well completion report; Esso Exploration & Production Australia Inc., Unpublished open file report. Mineral Resources Tasmania.

Maas, R., Nicholls, I., 2012. A Proterozoic basement block in SE Australia: Sr-Nd-O isotope tracing of the Selwyn Block, Victoria, 34th International Geological Congress. Australian Geoscience Council, Brisbane, Qld.

Macdonald, F.A., Prave, A.R., Petterson, R., Smith, E.F., Pruss, S.B., Oates, K., Waechter, F., Trotzuk, D., Fallick, A.E., 2013. The Laurentian record of Neoproterozoic glaciation, tectonism, and eukaryotic evolution in Death Valley, California. Geological Society of America Bulletin 125, 1203-1223. doi:10.1130/b30789.1.

Madsen, A.O., 2002. The structural development of the Torquay Embayment, Eastern Otway Basin, Victoria, Australia. M.Sc., University of Copenhagen.

Maher, S., Moore, D.H., Crawford, A.J., Twyford, R., Fanning, C.M., 1997. Test drilling on the southern margin of the Murray Basin, Victorian Initiative for Minerals and Petroleum Report, 52. Department of Natural Resources and Environment, Melbourne, 275 pages.

McClenaghan, M.P., Vicary, M., 2005. Sheet 4442, Railton. Digital Geological Atlas 1:25 000 series. Mineral Resources Tasmania, Hobart.

McDougall, I., Leggo, P.J., 1965. Isotopic age determinations on granitic rocks from Tasmania. Journal of the Geological Society of Australia 12, 295-332. doi:10.1080/00167616508728598.

McFarlane, H.B., 2011. The structural architecture and tectonic evolution of the Avebury-Trial Harbour area, western Tasmania. BSc(Hons), Monash University Clayton, Vic, 40 pages.

McLean, M.A., Morand, V.J., Cayley, R.A., 2010. Gravity and magnetic modelling of crustal structure in central Victoria: what lies under the Melbourne Zone? Australian Journal of Earth Sciences 57, 153-173. doi:10.1080/08120090903416245.

McNeill, A.W., Mortensen, J.K., Gemmell, J.B., 2012. High precision U-Pb chronostratigraphy of the Mount Read Volcanics, Tasmania: implications for mineral exploration and tectonic reconstructions, In: S.White (Ed.), The Great East-Australian Geology Arm Wave of 2012: tectonics, modern analogues/events, mineralisation. (GAGA 2012) Selwyn Symposium of the Geological Society of Australia Victoria Division. Geological Society of Australia, Melbourne, 70.

McWilliams, M., Schmidt, P., 2003. Paleomagnetism of Grassy Group rocks, King Island, Tasmania, EGS - AGU - EUG Joint Assembly, Abstracts from the meeting held in Nice, France, 6 - 11 April 2003, abstract #12406, Nice.

Medig, K.P.R., Thorkelson, D.J., Davis, W.J., Rainbird, R.H., Gibson, H.D., Turner, E.C., Marshall, D.D., 2014. Pinning northeastern Australia to northwestern Laurentia in the Mesoproterozoic. Precambrian Research 249, 88-99. doi:http://dx.doi.org/10.1016/j.precamres.2014.04.018.

Meffre, S., Berry, R.F., Hall, M., 2000. Cambrian metamorphic complexes in Tasmania: tectonic implications. Australian Journal of Earth Sciences 47, 971-985. doi:10.1046/j.1440-0952.2000.00825.x.

Meffre, S., Direen, N.G., Crawford, A.J., Kamenetsky, V., 2004. Mafic volcanic rocks on King Island, Tasmania: evidence for 579 Ma break-up in east Gondwana. Precambrian Research 135, 177-191. doi:10.1016/j.precamres.2004.08.004.

Metcalfe, I., 2011. Tectonic framework and Phanerozoic evolution of Sundaland. Gondwana Research 19, 3-21. doi:10.1016/j.gr.2010.02.016.

Metcalfe, I., 2013. Gondwana dispersion and Asian accretion: Tectonic and palaeogeographic evolution of eastern Tethys. Journal of Asian Earth Sciences 66, 1-33. doi:http://dx.doi.org/10.1016/j.jseaes.2012.12.020.

Miller, J.M., Gray, D.R., 2007. Subduction-related deformation and the Narooma anticlinorium, Eastern Lachlan Fold Belt, Southeastern New South Wales. Australian Journal of Earth Sciences 44, 237-251. doi:http://dx.doi.org/10.1080/08120099708728307.

Miller, J.M., Phillips, D., Wilson, C.J.L., Dugdale, L.J., 2005. Evolution of a reworked orogenic zone: the boundary between the Delamerian and Lachlan Fold Belts, southeastern Australia. Australian Journal of Earth Sciences 52, 921-940. doi:10.1080/08120090500304265.

Milligan, P.R., Franklin, R., Minty, B.R.S., Richardson, L.M., Percival, P.J., 2010. Magnetic anomaly map of Australia. Fifth ed. Geoscience Australia, Department of Resources, Energy and Tourism, Canberra ACT.

Mineral Resources Tasmania Data Management Group, 2011. Tasmania 1:25,000 digital geology. Mineral Resources Tasmania, Department of Infrastructure, Energy and Resources, Hobart.

Moore, D.H., 2002. Basement-basin relationships in the Otway Basin, Victoria, southeast Australia, Victorian Initiative for Minerals and Petroleum Report, 78. Department of Natural Resources and Environment, Melbourne, 48 pages.

Moore, D.H., 2006. The boundary of the Rodinian craton in southeastern Australia, In: Denham, D. (Ed.), 2006 Australian Earth Sciences Convention Geological Society of Australia and Australian Society of Exploration Geophysicists, Melbourne Vic.

Moore, D.H., Betts, P., Aillères, L., 2012a. Resolving the Tyennan-Delamerian subduction paradox, In: Betts, P., Wilson, C., Rawling, T., Venn, C. (Eds.), Specialist Group in Tectonics and Structural Geology conference 'Cause and effects of deformation in thhe lithosphere'. Geological Society of Australia, Waratah Bay, Victoria, 91 pages.

Moore, D.H., Betts, P.G., Hall, M., 2012b. Western Tasmania came together in the Cambrian, In: Plavsa, D., Collins, A. (Eds.), 2012 International Association for Gondwana Research Annual Convention and 9th International Symposium on Gondwana to Asia, Adelaide, pp. 9-11.

Moore, D.H., Betts, P.G., Hall, M., 2015. Fragmented Tasmania: the transition from Rodinia to Gondwana. Australian Journal of Earth Sciences 62, 1-35. doi:10.1080/08120099.2014.966757.

Moore, D.H., Betts, P.G., Hall, W.D.M., 2013. Towards understanding the early Gondwanan margin in southeastern Australia. Gondwana Research 23, 1581-1598. doi:10.1016/j.gr.2012.08.006.

Moore, D.H., McLean, M.A., 2009. The northern Bendigo Zone: interpretation of the Bendigo-Mitiamo gravity survey, including depths to basement, Gold Undercover Report 11. Geoscience Victoria, Melbourne, 17 pages.

Moore, D.H., Wong, D., 2002. Eastern and central Gippsland Basin, southest Australia; basement interpretation and basement links, Victorian Initiative for Minerals and Petroleum Report, 69. Department of Natural Resources and Environment, Melbourne, 52 pages.

Morand, V.J., 1995. The Barrabool Hills Metagabbro: a curious piece of the Cambrian jig-saw puzzle in Victoria, In: Hill, E. (Ed.), Specialist Group in Tectonics and Structural Geology Conference. Geological Society of Australia Abstracts 40, Clare Valley, pp. 111-112.

Morand, V.J., Fanning, C.M., 2006. SHRIMP zircon dating results for various rocks from Victoria, 2006, Technical Record, 2006/2. Geological Survey of Victoria, Melbourne, 70 pages.

Morand, V.J., Fanning, C.M., 2009. SHRIMP zircon dating results for various rocks from Victoria, 2009, Technical Record, 2009/2. Melbourne, 38 pages.

Morand, V.J., McKnight, S., 2006. Metamorphic style of the Tabberabbera Zone, Lachlan Fold Belt, In: Denham, D. (Ed.), Australian Earth Sciences Convention 2006. Geological Society of Australia and Australian Society of Exploration Geophysicists, Melbourne.

Morand, V.J., Wohlt, K.E., Taylor, D.H., Cayley, R.A., Kemp, A.I.S., Magart, A.P.M., Simons, B.A., 2004. Glenelg Special Map area, Geological Report, 123. Geological Survey of Victoria, 256 pages.

Moresi, L., Betts, P.G., Miller, M.S., Cayley, R.A., 2014. Dynamics of continental accretion. Nature 508, 245-248. doi:10.1038/nature13033.

Morey, A.A., Bierlein, F.P., Cherry, D.P., Turner, G., 2002. Genesis of greenstone-hosted Cu–Au mineralisation at Hill 800, Mt Useful Slate Belt, eastern Victoria. Australian Journal of Earth Sciences 49, 787-799. doi:10.1046/j.1440-0952.2002.00954.x.

Morse, M., Gibson, G., Mitchell, C., 2009. Basement constraints on offshore basin architecture as determined by new aeromagnetic data acquired over Bass Strait and western margin of Tasmania, ASEG 20th Geophysical Conference & Exhibition extended abstracts. Australian Society of Exploration Geophysicists.

Mortensen, J.K., Gemmell, J.B., McNeill, A.W., Friedman, R.M., 2015. High-precision U-Pb zircon chronostratigraphy of the Mount Read Volcanic Belt in western Tasmania, Australia: implications for VHMS deposit formation. Economic Geology 110, 445-468.

Münker, C., Crawford, A.J., 2000. Cambrian arc evolution along the SE Gondwana active margin: a synthesis from Tasmania-New Zealand-Australia-Antarctica correlations. Tectonics 19, 415-432. doi:http://dx.doi.org/0278-7407/00/2000TC900002512.

Nakamura, A., Bacchin, M., Milligan, P.R., Tracey, R., 2011. Residual gravity anomaly map of onshore Australia. First ed. Geoscience Australia, Department of Resources, Energy and Tourism, Canberra.

Nelson, D.R., 2006a. 178070: amphibolite, Haig Cave; geochronology dataset 596, in Compilation of geochronology data, June 2006 update. Western Australian Geological Survey, Perth, 4 pages.

Nelson, D.R., 2006b. 178072: tonalitic gneiss, Haig Cave; geochonology dataset 598, in Compilation of geochronology data, June 2006 update. Western Australia Geological Survey, 4 pages.

Noll, C.A., Hall, M., 2005. Structural architecture of the Owen Conglomerate, West Coast Range, western Tasmania: field evidence for Late Cambrian extension. Australian Journal of Earth Sciences 52, 411-426. doi:10.1080/08120090500134563.

Norvick, M.S., Smith, M.S., 2001. Mapping the plate tectonic reconstruction of southern and southeastern Australia and implications for petroleum systems. APPEA Journal 41/1, 15-35.

O'Halloran, G.J., Johnstone, E.M., 2001. Late Cretaceous Rift Volcanics of the Gippsland Basin, SE Australia - New Insights from 3D Seismic, In: Hill, K.C., Bernecker, T. (Eds.), Eastern Australian Basins Symposium 1. Petroleum Exploration Society of Australia, Melbourne, pp. 353-362.

O'Neill, W.J., 1995. North Mining Ltd. E.L. 3285, Dimboola, Victoria. Final report January, 1995, Expired minerals exploration reports file (unpubl.). Department of Agriculture, Energy and Minerals, Melbourne.

Offler, R., Miller, J.M., Gray, D.R., Foster, D.A., Bale, R., 1998. Crystallinity and b_0 spacing of K-white micas in a Paleozoic accretionary complex, eastern Australia: metamorphism, paleogeotherms, and structural style of an underplated sequence. Journal of Geology 106, 495-509. doi:http://dx.doi.org/0022-1376/98/10604-0006.

Orth, K., VandenBerg, A.H.M., Nott, R.J., Simons, B.A., 1995. Murrindal 1: 100,000 map geological report, Report, 100. Geological Survey of Victoria, 236 pages.

Patison, N.L., Berry, R.F., Davidson, G.J., Taylor, B.P., Bottrill, R.S., Manzi, B., Ryba, J., R.E.Shepherd, 2001. Regional metamorphism of the Mathinna Group, northeast Tasmania. Australian Journal of Earth Sciences 48, 281-292. doi:10.1046/j.1440-0952.2001.00862.x.

Patriat, P., Achache, J., 1984. India-Eurasia collision chronology has implications for crustal shortening and driving mechanism of plates. Nature 311, 615-621. doi:10.1038/311615a0.

Paulick, H., McPhie, J., 1999. Facies architecture of the felsic lava-dominated host sequence to the Thalanga massive sulfide deposit, Lower Ordovician, northern Queensland. Australian Journal of Earth Sciences 46, 391-405. doi:10.1046/j.1440-0952.1999.00713.x.

Payne, J.L., Hand, M., Barovich, K.M., Reid, A., Evans, D.A.D., 2009. Correlations and reconstruction models for the 2500–1500 Ma evolution of the Mawson Continent, In: Reddy, S.M., Mazumder, R., Evans, D.A.D., Collins, A.S. (Eds.), Palaeoproterozoic supercontinents and global evolution. Special Publication 323. Geological Society of London, London, pp. 319–355.

Péron-Pinvidic, G., Manatschal, G., 2010. From microcontinents to extensional allochthons: witnesses of how continents rift and break apart? Petroleum Geoscience 16, 189-197. doi:10.1144/1354-079309-903.

Peucat, J.J., Capdevila, R., Fanning, C.M., Ménot, R.P., Pécora, L., Testut, L., 2002. 1.60 Ga felsic volcanic blocks in the moraines of the Terre Adélie Craton, Antarctica: comparisons with the Gawler Range Volcanics, South Australia. Australian Journal of Earth Sciences 49, 831-845. doi:10.1046/j.1440-0952.2002.00956.x.

Phillips, G., Miller, J.M., Wilson, C.J.L., 2002. Structural and metamorphic evolution of the Moornambool Metamorphic Complex, western Lachlan Fold Belt, southeastern Australia. Australian Journal of Earth Sciences 49, 891-913. doi:10.1046/j.1440-0952.2002.00958.x.

Pilia, S., Rawlinson, N., Direen, N.G., Reading, A.M., Cayley, R., Pryer, L., Arroucau, P., Duffett, M., 2015. Linking mainland Australia and Tasmania using ambient seismic noise tomography: Implications for the tectonic evolution of the east Gondwana margin. Gondwana Research 28, 1212-1227. doi:http://dx.doi.org/10.1016/j.gr.2014.09.014.

Powell, C.M., Baillie, P.W., 1992. Tectonic affinity of the Mathinna Group in the Lachlan Fold Belt. Tectonophysics 214, 193-209. doi:10.1016/0040-1951(92)90197-E.

Powell, C.M., Baillie, P.W., Conaghan, P.J., Turner, N.J., 1993. The mid-Palaeozoic turbiditic Mathinna Group, northeast Tasmania. Australian Journal of Earth Sciences 40, 169-196. doi:10.1080/08120099308728073.

Powell, C.M., Baillie, P.W., VandenBerg, A.H.M., 2003. Silurian to mid-Devonian basin development of the Melbourne Zone, Lachlan Fold Belt, southeastern Australia. Tectonophysics 375, 9-36. doi:10.1080/08120099308728073.

Preiss, W., 2001. Neoproterozoic extension and Cambrian plate convergence in South Australia; relationships with Tasmania, In: Davidson, G., Pongratz, J. (Eds.), 2001: a structural odyssey. Specialist Group in Tectonics & Structural Geology, Geological Society of Australia, Ulverstone, Tas., pp. 142-143.

Preiss, W.V., 1995. Metamorphism, In: Drexel, J.F., Preiss, W.V. (Eds.), The geology of South Australia. Volume 2, The Phanerozoic. Bulletin 54, South Australia Geological Survey, Adelaide, pp. 53-54.

Preiss, W.V., 2000. The Adelaide Geosyncline of South Australia and its significance in Neoproterozoic continental reconstruction. Precambrian Research 100, 21-63. doi:10.1016/S0301-9268(99)00068-6.

Preiss, W.V., Belperio, A., Cowley, W.M., Rankin, L.R., 1993. Neoproterozoic, In: Drexel, J.F., Preiss, W.V., Parker, A.J. (Eds.), The geology of South Australia. Volume 1, The Precambrian. Bulletin 54, Geological Survey of South Australia, Adelaide, pp. 170-203.

Premier Oil (Australasia), Ltd, 1999. Well completion report White Ibis 1, Unpublished open file report. Mineral Resources Tasmania, 470 pages.

Prendergast, E., Offler, R., 2012. Underplated seamount in the Narooma accretionary complex, NSW, Australia. Lithos 154, 224-234. doi:http://dx.doi.org/10.1016/j.lithos.2012.07.010.

Price, R.C., Gray, C.M., Frey, F.A., 1997. Strontium isotopic and trace element heterogeneity in the plains basalts of the Newer Volcanic Province, Victoria, Australia. Geochimica et Cosmochimica Acta 61, 171-192. doi:http://dx.doi.org/0016-7037/97.

Price, R.C., Nicholls, I.A., Day, A., 2014. Lithospheric influences on magma compositions of late Mesozoic and Cenozoic intraplate basalts (the Older Volcanics) of Victoria, south-eastern Australia. Lithos 206–207, 179-200. doi:http://dx.doi.org/10.1016/j.lithos.2014.07.027.

Price, R.C., Nicholls, I.A., Gray, C.M., 2003. Cainozoic igneous activity, In: Birch, W.D. (Ed.), Geology of Victoria, 3rd ed. Geological Society of Australia (Victorian Division), Melbourne, Vic., pp. 361-375.

Purss, M.B.J., Cull, J., 2001. Heat-flow data in western Victoria. Australian Journal of Earth Sciences 48, 1-4. doi:10.1046/j.1440-0952.2001.00840.x.

Quilty, P.G., Hill, P.J., Exon, N.E., Baillie, P.W., Everard, J.L., Forsyth, S.M., Calver, C.R., Bottrill, R.S., Taheri, J., Bayer, E.E., Doyle, R.B., O'Reilly, S.Y., Zhang, M., Macphail, M., Hill, R.S., Partridge, A.D., Jordan, G.J., Lewis, S., 2014. Cretaceous-Neogene evolution of Tasmania, In: Corbett, K.D., Quilty, P.G., Calver, C.R. (Eds.), Geological evolution of Tasmania. Geological Society of Australia (Tasmania Division), Hobart, Tas., pp. 409-509.

Rawling, T.J., Osborne, C.R., Mclean, M.A., Skladzien, P.S., Cayley, R.A., Wllliams, B., 2011. 3D Victoria Final Report, 3D Victoria Report 14, 14. Department of Primary Industries, Victoria, Melbourne, p. 96. doi:978-1-74264-977-1.

Rawlinson, N., Kennett, B.L.N., 2008. Teleseismic tomography of the upper mantle beneath the southern Lachlan Orogen, Australia. Physics of the Earth and Planetary Interiors 167, 84-97. doi:http://dx.doi.org/10.1016/j.pepi.2008.02.007.

Rawlinson, N., Salmon, M., Kennett, B.L.N., 2014. Transportable seismic array tomography in southeast Australia: Illuminating the transition from Proterozoic to Phanerozoic lithosphere. Lithos 189, 65-76. doi:http://dx.doi.org/10.1016/j.lithos.2013.06.001.

Rawlinson, N., Tkalčić, H., Reading, A.M., 2010. Structure of the Tasmanian lithosphere from 3D seismic tomography. Australian Journal of Earth Sciences 57, 381-394. doi:10.1080/08120099.2010.481325.

Rawlinson, N., Urvoy, M., 2006. Simultaneous inversion of active and passive source datasets for 3-D seismic structure with application to Tasmania. Geophysical Research Letters 33, L24313, 5. doi:10.1029/2006GL028105.

Reed, A., 2001. Pre-Tabberabberan deformation in eastern Tasmania: a southern extension of the Benambran Orogeny. Australian Journal of Earth Sciences 48, 785-796. doi:10.1046/j.1440-0952.2001.00900.x.

Reed, A., Calver, C.R., Bottrill, R.S., 2002. Palaeozoic suturing of eastern and western Tasmania in the west Tamar region: implications for the tectonic evolution of southeast Australia. Australian Journal of Earth Sciences 49, 809-830. doi:10.1046/j.1440-0952.2002.00955.x.

Reed, A., Vicary, M.J., 2005. Sheet 4843 Beaconsfield. Digital Geological Atlas 1:25 000 series. Mineral Resources Tasmania.

Reed, A.R., Zengerer, M., Roach, M.J., 2001. Andersons Creek Ultramafic Complex and Simmonds Hill Metamorphics, In: Reed, A.R. (Ed.), Structure and setting of Proterozoic and Palaeozoic rocks in the Tamar region, northern Tasmania. Field Guide 9, Specialist Group in Tectonics and Structural Geology, Geological Society of Australia. Mineral Resources Tasmania and Centre for Ore Deposit Research and School of Earth Sciences University of Tasmania, Hobart, pp. 43-55.

Reid, C., Chester, A., Stacey, A., Burrett, C., 2003. Stratigraphic results of diamond drilling of the Hunterston Dome, Tasmania: implications for palaeogeography and hydrocarbon potential. Papers and proceedings of the Royal Society of Tasmania 137, 87-94.

Richards, J.R., Singleton, O.P., 1981. Palaeozoic Victoria, Australia: Igneous rocks, ages and their interpretation. Journal of the Geological Society of Australia 28, 395-421. doi:10.1080/00167618108729178.

Rickards, R.B., Hills, P.B., Banks, M.R., MacDonald, G., 2002. The significance of a new locality for *Monograptus Thomasi* (Early Devonian) southwest of Beaconsfield, Tasmania. Papers and proceedings of the Royal Society of Tasmania 136, 7-16.

Roach, M.J., Leaman, D.E., 1996. The regional structure of northeast Tasmania - a geophysical perspective, Abstracts 13th Australian Geological Convention; Canberra, February 1996. Geological Society of Australia, Canberra, p. 366.

Ross, H.E., Blakely, R.J., Zoback, M.D., 2006. Testing the use of aeromagnetic data for the determination of Curie depth in California. Geophysics 71, L51-L59. doi:10.1190/1.2335572.

Rossetti, F., Vignaroli, G., Vincenzo, G.D., Gerdes, A., Ghezzo, C., Theye, T., Balsamo, F., 2011. Long-lived orogenic construction along the paleo-Pacific margin of Gondwana (Deep Freeze Range, North Victoria Land, Antarctica). Tectonics 30, TC4008, 1-27. doi:10.1029/2010TC002804.

Rossiter, A., Gray, C.M., 2008. Barium contents of Victorian granites: key to understanding crustal architecture in the southern Lachlan Fold Belt? Australian Journal of Earth Sciences 55, 433-448. doi:http://dx.doi.org/10.1080/08120090801888586

Royer, J.-Y., Rollet, N., 1997. Plate-tectonic setting of the Tasmanian region. Australian Journal of Earth Sciences 44, 543-560. doi:10.1080/08120099708728336.

Scheibner, E., 1997. Stratotectonic map of New South Wales scale 1:1 000 000. Geological Survey of New South Wales, Sydney.

Scheibner, E., Veevers, J.J., 2000. Tasman Fold Belt System, In: Veevers, J.J. (Ed.), Billion-year earth history of Australia and neighbours in Gondwanaland. GEMOC Press, Sydney, pp. 154-234.

Schellart, W.P., 2008. Overriding plate shortening and extension above subduction zones: A parametric study to explain formation of the Andes Mountains. Geological Society of America Bulletin 120, 1441-1454. doi:10.1130/b26360.1.

Schneider, C.L., 2005. The 3D structural evolution of the Shipwreck Trough, Otway Basin, southeastern Australia: architecture of an oblique rift margin. PhD, Monash University, 219 pages.

Scholte, T., 2010. Physical volcanology, composition & emplacement environment of the Flinders Province Basalt, Flinders, Victoria. B.Sc.(Hons), Monash University, 38 pages.

Sdrolias, M., Muller, R.D., Gaina, C., 2003. Tectonic evolution of the southwest Pacific using constraints from backarc basins. Geological Society of America Special Papers 372, 343-359. doi:10.1130/0-8137-2372-8.343.

Setiawan, A., 2000. The evolution of the Bass Basin. M.Sc., Monash University, 179 pages.

Seymour, D.B., Calver, C.R., 1995. Explanatory notes for the time–space diagram and stratotectonic elements map of Tasmania, Tasmanian Geological Survey Record, 1995/01. Tasmanian Department of Development and Resources and the Australian Geological Survey Organisation, Hobart, p. 62.

Seymour, D.B., Calver, C.R., 1998. Time-space diagram for Tasmania. NGMA TASGO Project: sub-project 1 - geological synthesis, Second ed. Mineral Resources Tasmania and Australian Geological Survey Organisation, Hobart Tas.

Seymour, D.B., Green, G.R., Calver, C.R., 2007. The geology and mineral deposits of Tasmania: a summary, Bulletin, 72. Mineral Resources Tasmania, Hobart Tas., 29 pages.

Seymour, D.B., Green, G.R., Calver, C.R., 2013. The geology and mineral deposits of Tasmania: a summary, Bulletin, 72. Third ed. Mineral Resources Tasmania, Hobart, 29 pages.

Seymour, D.B., Vicary, M.J., 2010. A review of Cambrian megabreccias in the Penguin–Ulverstone area, central northern Tasmania, Record, 2010/02. Mineral Resources Tasmania, Department of Infrastructure, Energy and Resources, Hobart Tas., 23 pages.

Seymour, D.B., Woolward, I.R., McClenaghan, M.P., Bottrill, R.S., 2011. Stratigraphic revision and re-mapping of the Mathinna Supergroup between the River Tamar and the Scottsdale Batholith, northeast Tasmania, Explanatory Report 1:25 000 Scale Digital Geological Map Series, 4. Mineral Resources Tasmania, Hobart, 82 pages.

Shaw, R.D., Wellman, P., Gunn, P.J., Whitaker, A.J., Tarlowski, C., Morse, M.P., 1996. Guide to using the Australian crustal elements map, Record 1996/30, Australian Geological Survey Organisation Canberra.

Simmons, J.M., Cas, R.A.F., Nicholls, I.A., 2014. The volcanological and petrogenetic origins of the basaltic Flinders Volcanic Province (49–39 Ma), Older Volcanics Provinces, southeast Australia. Australian Journal of Earth Sciences 61, 719-733. doi:10.1080/08120099.2014.920725.

Simons, B.A., Moore, D.H., 1999. Victoria 1:1 000 000 Pre-Permian geology. First ed. Geological Survey of Victoria, Melbourne.

Sircombe, K.N., Hazelton, M.L., 2004. Comparison of detrital zircon age distributions by kernel functional estimation. Sedimentary Geology 171, 91-111. doi:http://dx.doi.org/10.1016/j.sedgeo.2004.05.012.

Skladzien, P.B., 2007. Compilation of rock densities for Victoria, GeoScience Victoria Gold Undercover Report, 5. Department of Primary Industries, Melbourne, 64 pages.

Soesoo, A., Bons, P.D., Elburg, M.A., 1999. Freestone dykes—an alkali-rich Jurassic dyke population in eastern Victoria. Australian Journal of Earth Sciences 46, 1-9. doi:10.1046/j.1440-0952.1999.00682.x.

Spaggiari, C.V., Gray, D.R., Foster, D.A., 2002. Blueschist metamorphism during accretion in the Lachlan Orogen, south-eastern Australia. Journal of Metamorphic Geology 20, 711-726. doi:10.1046/j.1525-1314.2002.00405.x.

Spaggiari, C.V., Gray, D.R., Foster, D.A., 2003a. Formation and emplacement of the Dolodrook serpentinite body, Lachlan Orogen, Victoria. Australian Journal of Earth Sciences 50, 709-723. doi:http://dx.doi.org/10.1111/j.1440-0952.2003.01021.x.

Spaggiari, C.V., Gray, D.R., Foster, D.A., 2003b. Tethyan- and Cordilleran-type ophiolites of eastern Australia: implications for the evolution of the Tasmanides, In: Dilek, V., Robinson, P.T. (Eds.), Ophiolites in Earth history. Special Publication 218. The Geological Society of London, London, pp. 517-539.

Spaggiari, C.V., Gray, D.R., Foster, D.A., 2004. Ophiolite accretion in the Lachlan Orogen, southeastern Australia. Journal of Structural Geology 26, 87-112. doi:http://dx.doi.org/10.1016/S0191-8141(03)00084-1.

Spaggiari, C.V., Gray, D.R., Foster, D.A., McKnight, S., 2003c. Evolution of the boundary between the western and central Lachlan Orogen: implications for Tasmanide tectonics. Australian Journal of Earth Sciences 50, 725-749. doi:10.1111/j.1440-0952.2003.01022.x.

Squire, R.J., Wilson, C.J.L., 2005. Interaction between collisional orogenesis and convergent-margin processes: evolution of the Cambrian proto-Pacific margin of East Gondwana. Journal of the Geological Society 162, 749-761. doi:10.1144/0016-764904-087.

Squire, R.J., Wilson, C.J.L., Dugdale, L.J., Jupp, B.J., Kaufman, A.L., 2006. Cambrian backarc-basin basalt in western Victoria related to evolution of a continent-dipping subduction zone. Australian Journal of Earth Sciences 53, 707-719. doi:http://dx.doi.org/10.1080/08120090600827405.

Stacey, A.R., Berry, R.F., 2004. The structural history of Tasmania: a review for petroleum explorers, PESA Eastern Australian Basins Symposium II. Petroleum Exploration Society of Australia, Adelaide, South Australia, pp. 151-162.

Stern, R.J., Reagan, M., Ishizuka, O., Ohara, Y., Whattam, S., 2012. To understand subduction initiation, study forearc crust: to understand forearc crust, study ophiolites. Lithosphere 4, 469-483. doi:10.1130/l183.1.

Stewart, J.R., Betts, P.G., 2010. Implications for Proterozoic plate margin evolution from geophysical analysis and crustal-scale modeling within the western Gawler Craton, Australia. Tectonophysics 483, 151-177. doi:http://dx.doi.org/10.1016/j.tecto.2009.11.016.

Stuart-Smith, P.G., Black, L.P., 1999. Willaura Sheet 7422 Victoria 1:100 000 map geological report, Record 1999/38. Australian Geological Survey Organisation, 49 pages.

Stump, E., Gootee, B., Talarico, F., 2003. Tectonic model for development of the Byrd Glacier discontinuity and surrounding regions of the Transantarctic Mountains during the Neoproterozoic - Early Paleozoic, In: Fütterer, D.K., Damaske, D., Kleinschmidt, G., Miller, H., Tessensohn, F. (Eds.), Antarctica: contributions to global earth sciences Proceedings of the IX international symposium of Antarctic earth sciences Potsdam 2003. Springer, Heidelberg, Berlin, pp. 181-190.

Stump, E., Gootee, B., Talarico, F., 2006. Tectonic model for development of the Byrd Glacier discontinuity and surrounding regions of the Transantarctic Mountains during the Neoproterozoic - Early Paleozoic, In: Fütterer, D.K., Damaske, D., Kleinschmidt, G., Miller, H., Tessensohn, F. (Eds.), Antarctica Contributions to Global Earth Sciences Proceedings of the IX International Symposium of Antarctic Earth Sciences Potsdam 2003. Springer, Heildelberg Berlin, pp. 181-190.

Stump, E., White, A.J.R., Borg, S.G., 1986. Reconstruction of Australia and Antarctica: evidence from granites and recent mapping. Earth and Planetary Science Letters, 79, 348-360. doi:10.1016/0012-821X(86)90191-3.

Taylor, D.H., Cayley, R.A., 2000. Character and kinematics of faults within the turbidite-dominated Lachlan Orogen: implications for tectonic evolution of eastern Australia: discussion. Journal of Structural Geology 22, 523-528. doi:10.1016/0012-821X(86)90191-3.

Taylor, D.H., Whitehead, M.L., Olshina, A., Leonard, J.G., 1996. Ballarat 1:100 000 map geological report, Report, 101. Geological Survey of Victoria, 117 pages.

Teasdale, J.P., Pryer, L.L., Stuart-Smith, P.G., Romine, K.K., Etheridge, M.A., Loutit, T.S., Kyan, D.M., 2003. Structural framework and basin evolution of Australia's southern margin. APPEA Journal 43, 13-37.

Tessensohn, F., Henjes-Kunst, F., 2005. Northern Victoria Land terranes, Antarctica: far-travelled or local products?, In: Vaughan, A.P.M., Leat, P.T., Pankhurst, R.J. (Eds.), Terrane Processes at the Margins of Gondwana. Special Publication 246. Geological Society, London U.K., pp. 275-291.

Thompson, J.F.H., Sillitoe, R.H., Baker, T., Lang, J.R., Mortensen, J.K., 1999. Intrusion-related gold deposits associated with tungsten-tin provinces. Mineralium Deposita 34, 323-334. doi:10.1007/s001260050207.

Tickell, S.J., 1990. Colac and part Beech Forest. 1:50,000 geological map, First ed. Victorian Department of Manufacturing & Industry Development, Melbourne.

Timmons, J.M., Karlstrom, K.E., Dehler, C.M., Geissman, J.W., Heizler, M.T., 2001. Proterozoic multistage (ca. 1.1 and 0.8 Ga) extension recorded in the Grand Canyon Supergroup and establishment of northwest- and north-trending tectonic grains in the southwestern United States. Geological Society of America Bulletin 113, 163-181. doi:10.1130/0016-7606(2001)113<0163:pmcage>2.0.co;2.

Trigg, K.R., Blevin, J.E., Boreham, C.J., 2003. An audit of petroleum exploration wells in the Bass Basin 1965 - 1999, Record, 2003/11. Geoscience Australia, Canberra, p. 281.

Turner, N.J., Black, L.P., Kamperman, M., 1998. Dating of Neoproterozoic and Cambrian orogenies in Tasmania. Australian Journal of Earth Sciences 45, 789-806. doi:10.1080/08120099808728433.

Turner, N.J., Bottrill, R.S., 2001. Blue amphibole, Arthur Metamorphic Complex, Tasmania: composition and regional tectonic setting. Australian Journal of Earth Sciences 48, 167-181. doi:10.1046/j.1440-0952.2001.00847.x.

Turner, S., Haines, P., Foster, D., Powell, R., Sandiford, M., Offler, R., 2009. Did the Delamerian Orogeny Start in the Neoproterozoic? Journal of Geology 117, 575-583. doi:10.1086/600866.

Turner, S.P., Adams, C.J., Flöttmann, T., Foden, J.D., 1993. Geochemical and geochronological constraints on the Glenelg River Complex, western Victoria. Australian Journal of Earth Sciences 40, 275-292. doi:10.1080/08120099308728080.

VandenBerg, A.H.M., 1991. Kilmore 1:50,000 map geological report, Report, 91. Geological Survey of Victoria, 85 pages.

VandenBerg, A.H.M., Cayley, R.A., Willman, C.E., Morand, V.J., Seymon, A.R., Osborne, C.R., Taylor, D.H., Haydon, S.J., McLean, M., Quinn, C., Jackson, P., Sandford, A.C., 2006. Walhalla-Woods Point-Tallangallook special area geological report, Report, 127. Geological Survey of Victoria, Melbourne, 448 pages.

VandenBerg, A.H.M., Willman, C.E., Hendrickx, M., Bush, M.D., Sands, B.C., 1995. The geology and prospectivity of the 1993 Mount Wellington airborne survey area, Victorian Initiative for Minerals and Petroleum Report, 2. Geological Survey of Victoria, Melbourne, 152 pages.

VandenBerg, A.H.M., Willman, C.E., Maher, S., Simons, B.A., Cayley, R.A., Taylor, D.H., Morand, V.J., Moore, D.H., Radojkovic, A., 2000. The Tasman Fold Belt System in Victoria, Special Publication. Geological Survey of Victoria, 463 pages.

VandenBerg, A.H.M., Willman, C.E., Morand, V.J., McHaffie, I.W., Simons, B.A., Quinn, C., Westcott, A., 2004. Buffalo 1:100 000 map area geological report, Report, 124. Geological Survey of Victoria, Melbourne, 205 pages.

Veevers, J.J., 2007. Pan-Gondwanaland post-collisional extension marked by 650–500 Ma alkaline rocks and carbonatites and related detrital zircons: a review. Earth Science Reviews 83, 1-83. doi:http://dx.doi.org/10.1016/j.earscirev.2007.03.001.

Veevers, J.J., 2012. Reconstructions before rifting and drifting reveal the geological connections between Antarctica and its conjugates in Gondwanaland. Earth Science Reviews 111, 249-318. doi:10.1016/j.earscirev.2011.11.009.

Veevers, J.J., Eittreim, S.L., 1988. Reconstruction of Antarctica and Australia at breakup (95 ± 5 Ma) and before rifting (160 Ma). Australian Journal of Earth Sciences 35, 355-362. doi:10.1080/08120098808729453.

Vicary, M., 2006. Re-interpretation of geological relationships in the Castra–Kindred area, Tasmanian Geological Survey Record, 2006/1. Hobart, 39 pages.

Vicary, M.J., 2004. Burnie 4045. 1:25 000 Geological Map. Mineral Resources Tasmania.

Vicary, M.J., Calver, C.R., Everard, J.L., Seymour, D.B., 2008. Ulverstone 4244 1:25 000 geological map. Digital Geological Atlas 1:25 000 series. Mineral Resources Tasmania.

Welsh, S.I., Higgins, D.V., Callaway, G.A.e., 2011. Surface Geology of Victoria 1 : 250 000. Geological Survey of Victoria, Department of Primary Industries, Melbourne. 116 pages.

White, A.J.R., 2002. Central Victorian granites — low oxidation states, near-surface intrusions and possible sources of salt, In: Phillips, G.N., Ely, K.S. (Eds.), Victoria Undercover: Benalla 2002 Conference proceedings and field guide: collaborative geoscience in northern Victoria. CSIRO Publishing, Melbourne, pp. 51-53.

Whitmeyer, S.J., Karlstrom, K.E., 2007. Tectonic model for the Proterozoic growth of North America. Geosphere 3, 220-259. doi:10.1130/ges00055.1.

Wilhem, C., Windley, B.F., Stampfli, G.M., 2012. The Altaids of Central Asia: A tectonic and evolutionary innovative review. Earth-Science Reviews 113, 303-341. doi:http://dx.doi.org/10.1016/j.earscirev.2012.04.001.

Williams, H., Stewart, J., Betts, P., 2009. Imposition of a Proterozoic salient on a Palaeozoic orogen at the eastern margin of Gondwana. Gondwana Research 16, 669-686. doi:10.1016/j.gr.2009.06.006.

Williams, S.E., Whittaker, J., Muller, R.D., 2012. Full-fit reconstructions of the southern Australian margin and Antarctica – implications for correlating geology between Australia and Antarctica, Proceedings of the Eastern Australasian Basins Symposium IV, Brisbane Qld, pp. 1-8.

Williams, S.E., Whittaker, J.M., Müller, R.D., 2011. Full-fit, palinspastic reconstruction of the conjugate Australian-Antarctic margins. Tectonics 30, TC6012. doi:10.1029/2011TC002912.

Willman, C.E., Cayley, R.A., Vandenberg, A.H.M., Haydon, S.J., Osborne, C.R., Seymon, A.R., Thom J.L., 2005. Dargo 1:100 000 map area geological report, Report, 126. Geological Survey of Victoria, 189 pages.

Wilson, C.J.L., McKnight, S.W., Dugdale, A.L., Rawling, T.J., Farrar, A.D., McKenzie, M.J., Melling, W.D., 2009. Illite crystallinity and the b-spacing values of white micas and their implications for gold mineralisation in the Lachlan Orogen. Australian Journal of Earth Sciences 56, 1143-1164. doi:http://dx.doi.org/10.1080/08120090903246238.

Wingate, M.T.D., Campbell, I.H., Compston, W., Gibson, G.M., 1998. Ion microprobe U-Pb ages for Neoproterozoic basaltic magmatism in south-central Australia and implications for the breakup of Rodinia. Precambrian Research 87, 135-159. doi:10.1016/S0301-9268(97)00072-7.

Wingate, M.T.D., Evans, D.A.D., 2003. Palaeomagnetic constraints on the Proterozoic tectonic evolution of Australia. Geological Society, London, Special Publication 206, 77-91. doi:10.1144/gsl.sp.2003.206.01.06.

Woof, C.H., 2006. The possible extension of northwestern Tasmanian, Neoproterozoic-Cambrian lithosphere into southern Victoria. BSc(Hons), Monash University, 40 pages.

Worthing, M.A., Woolward, I.R., 2010. Explanatory Report for the Dublin Town (5840), Brilliant (5841), Falmouth (6040) and Beaumaris (6041) geological map sheets, Explanatory Report 1:25 000 Scale Digital Geological Map Series, 3. Mineral Resources Tasmania, Hobart, 75 pages.

Yonkee, W.A., Dehler, C.D., Link, P.K., Balgord, E.A., Keeley, J.A., Hayes, D.S., Wells, M.L., Fanning, C.M., Johnston, S.M., 2014. Tectono-stratigraphic framework of Neoproterozoic to Cambrian strata, west-central U.S.: Protracted rifting, glaciation, and evolution of the North American Cordilleran margin. Earth-Science Reviews 136, 59-95. doi:http://dx.doi.org/10.1016/j.earscirev.2014.05.004.

Young, M.K., Rawlinson, N., Arroucau, P., Reading, A.M., Tkalčić, H., 2011. High-frequency ambient noise tomography of southeast Australia: new constraints on Tasmania's tectonic past. Geophysical Reseach Letters 38, L13313, 1-6. doi:10.1029/2011GL047971.

Yumul, G.P., Dimalanta, C.B., Tamayo, R.A., Maury, R.C., 2003. Collision, subduction and accretion events in the Philippines: a synthesis. Island Arc 12, 77-91. doi:10.1046/j.1440-1738.2003.00382.x.

Zengerer, M., 1999. Gravity and magnetics of the West Tamar district, northern Tasmania. BSc.(Hons), Unversity of Tasmania, Hobart.

Zhao, G., Cawood, P.A., 2012. Precambrian geology of China. Precambrian Research 222–223, 13-54. doi:http://dx.doi.org/10.1016/j.precamres.2012.09.017.

Zhao, J.-H., Zhou, M.-F., Yan, D.-P., Zheng, J.-P., Li, J.-W., 2011. Reappraisal of the ages of Neoproterozoic strata in South China: no connection with the Grenvillian orogeny. Geology 39, 299-302. doi:10.1130/g31701.1.

Zheng, J.P., Griffin, W.L., Li, L.S., O'Reilly, S.Y., Pearson, N.J., Tang, H.Y., Liu, G.L., Zhao, J.H., Yu, C.M., Su, Y.P., 2011. Highly evolved Archean basement beneath the western Cathaysia Block, South China. Geochimica et Cosmochimica Acta 75, 242-255. doi:http://dx.doi.org/10.1016/j.gca.2010.09.035.

Appendix 1

Z. R. DENNIS, D. H. MOORE & J. P. CULL. 2011 MAGNETOTELLURIC SURVEY FOR UNDERCOVER STRUCTURAL MAPPING, CENTRAL VICTORIA.

Australian Journal of Earth Sciences, 58: 1, 33 — 47.

Australian Journal of Earth Sciences
Publication details, including instructions for authors and subscription information:
http://www.informaworld.com/smpp/title~content=t716100753

Magnetotelluric survey for undercover structural mapping, Central Victoria

Z. R. Dennis[a]; D. H. Moore[a]; J. P. Cull[a]
[a] School of Geoscience, Monash University, Vic, Australia

Online publication date: 22 January 2011

To cite this Article Dennis, Z. R. , Moore, D. H. and Cull, J. P.(2011) 'Magnetotelluric survey for undercover structural mapping, Central Victoria', Australian Journal of Earth Sciences, 58: 1, 33 — 47
To link to this Article: DOI: 10.1080/08120099.2011.534817
URL: http://dx.doi.org/10.1080/08120099.2011.534817

Australian Journal of Earth Sciences (2011) **58**, (33–47)

Magnetotelluric survey for undercover structural mapping, Central Victoria

Z. R. DENNIS*, D. H. MOORE AND J. P. CULL

School of Geoscience, Monash University, Vic. 3800, Australia.

Over 2% of total world gold production has come from the Victorian gold fields; however, most of the shallow deposits have now been discovered, and new methods are required for exploration beneath the thick sedimentary cover of the Murray Basin. Consequently, as part of the GeoScience Victoria Gold Undercover Initiative, new magnetotelluric (MT) data have now been obtained to provide a greater understanding of potential gold bearing structures extending to mid-crustal and greater depths. Coinciding with part of the 2006 Central Victorian Seismic Survey, a 170 km long MT transect lies within the Bendigo and Melbourne Zones, two of the fault bounded subdivisions of the western Lachlan Fold Belt. Additional transient electromagnetic (TEM) data have now been obtained at each of the MT sites allowing further processing, including compensation for static shifts that are significant in this area. The revised MT section provides an image consistent with the seismic section. The major fault zones align with several dipping conductive slabs supporting the orientation and depth extent revealed by the seismic profile. There is some evidence in the MT data that near-surface geological structures can be extrapolated to mantle depths.

KEY WORDS: magnetotellurics, resistivity, conductivity, static shifts, transient electromagnetic, gold undercover, Murray Basin, Western Lachlan Orogen.

INTRODUCTION

Since the first discoveries of shallow alluvial gold in 1851 (Ramsay *et al.* 1998), there has been a significant drive for exploration of the gold-bearing slate belts of Central Victoria. In particular, the Bendigo Zone has been identified as the most richly endowed metallogenic belt in the Victorian gold province (Lisitsin *et al.* 2007) making up most of the 2% of total world production coming from the state (Phillips & Hughes 1996). However, as most of the shallow ore deposits have now been discovered (Hough *et al.* 2007), there is considerable incentive to develop and use new methods to detect buried mineralisation zones, particularly to the north of the current major gold regions, where the geological structure is obscured by sedimentary deposits from the Murray Basin (Smith & Frankcombe 2006).

As a part of the Gold Undercover initiative outlined by GeoScience Victoria (GSV), a magnetotelluric (MT) survey was conducted north of the town of Bendigo, during winter 2007, with the aim of providing a first look at the deep conductivity structure in this region (Cull *et al.* 2008). This was followed in 2008 by a second parallel line, offset by approximately 50 km to the north and terminating at the eastern end close to the town of Echuca. The initial models were presented in Cull *et al.* (2008) and Dennis *et al.* (2009) and these have indicated strong north–south trends are present, consistent with the location of the known faults in the area. The location of the 2007 MT transect coincides with east–west sections of the earlier seismic reflection survey conducted by GSV in 2006. The results of both the MT and seismic surveys, when combined with the results of several early profiles (e.g. Gray *et al.* 1991; Korsch *et al.* 2002), are designed to define the geometry of the major faults in the region in addition to providing some primary constraints for the deep structure in exposed parts of the western Lachlan Fold Belt (ANSIR 2008; O'Shea 2008; Willman *et al.* in press). The similarity in the resolvable depths of MT and seismic reflection profiles thus presents an opportunity for a comparison of results to further strengthen any geological implications that can be drawn from the geophysical results.

The original (KIGAM) 2-D inversion model for the major MT section (line 2) has been superimposed with the major fault lines identified in the seismic reflection profile in Dennis *et al.* (2009; figure 9). This preliminary overlap shows that significant correlation exists between areas of high seismic reflectivity and contrasting electrical response to the regional average. Galvanic distortion or static shift of MT data however, is a well documented problem (Sternberg *et al.* 1988; Pellerin & Hohmann 1990; Meju 1996; Simpson & Bahr 2005); thus, after the completion of a coinciding near surface Transient Electromagnetic (TEM) profile (Dennis *et al.* 2010), static shift corrections were applied to the (Bendigo line 2) MT dataset, allowing a more accurate conclusion to be drawn from the resulting section.

*Corresponding author: zara.dennis@sci.monash.edu.au

ISSN 0812-0099 print/ISSN 1440-0952 online © 2011 Geological Society of Australia
DOI: 10.1080/08120099.2011.534817

GEOLOGICAL SETTING

The MT lines were located over parts of the Bendigo and Melbourne Zones which, together with the Stawell Zone further west, comprise the Western Lachlan Orogen (see Figure 1) (Glen 2005); a world class gold province that has produced over 2500 tonnes (80 million ounces) of gold (Phillips & Hughes 1996). The surface exposures in all three are almost entirely marine turbidites (VandenBerg *et al.* 2000); dominantly Cambrian to Early Ordovician in the Stawell Zone, Early to Middle Ordovician in the Bendigo Zone and Silurian to Middle Devonian in the Melbourne Zone.

The lowest part of the Bendigo Zone is exposed in the east, in the Heathcote Fault Zone (HFZ). Here the outcrop includes significant quantities of Cambrian oceanic mafic tholeiites and associated deep marine sediments, including black shale. On the 2006 seismic data, the HFZ seems to continue to the MOHO (Korsch *et al.* 2008; Cayley *et al.* in press). The seismic data also suggest that the midcrust of the Bendigo Zone comprises stacked layers of tholeiite (Cayley *et al.* in press). Elsewhere, the surface exposures are of generally north-striking turbidites that have been simply chevron folded with interlimb angles of 35° to 50°, giving an overall shortening of about 60% (Gray & Willman 1991). As well, mostly west-dipping north-striking sub-vertical faults about 10 to 20 km apart cut the sequence. However, the overall surface form is sub-horizontal, with rises in the hangingwalls compensated for by fault movements of the order of 1 km. The seismic data show that the faults flatten at depth, from about 75° near the surface to about 15° in the tholeiites (Cayley *et al.* in press).

This folding and faulting largely took place in the Benambran Orogeny, from 455 to 440 Ma. How much of it was episodic and how much was continuous has been the subject of debate, with Gray & Foster (1998) and VandenBerg (1999) summarising the arguments for and against continuous deformation, respectively. It now seems possible that both arguments have some validity, with deformation being more pronounced in the period 455 to 440 Ma, but continuing outside this time bracket.

Except in the easternmost parts, where some of the mineralisation is younger, this deformation event coincided with the main gold-producing event in the Bendigo Zone. Willman *et al.* (in press) showed that the gold-bearing fluids were generated by dewatering of the midcrust during burial, deformation and amphibolite-facies metamorphism. Fluids then migrated upwards along the deep shallow-dipping faults until they reached the brittle–ductile transition where they rose in the brittle zones, into the hangingwalls of the faults to deposit gold where it is seen today.

Following the Benambran Orogeny, granites were intruded into the western Bendigo Zone at about 400 Ma. In the east, the granites are somewhat younger, typically with ages of 360 to 375 Ma and were intruded after the 385 to 380 Ma Tabberabberan Orogeny.

The Melbourne Zone, bounded by the Mt William Fault in the west and the Governor Fault in the north and east, contrasts strongly with the zones further west. At the surface, the carbonaceous Early to Middle Ordovician Mt Easton Shale is the oldest sedimentary unit exposed and is overlain by the Murrindindi Supergroup, a Silurian to Middle Devonian turbidite sequence that generally reflects a shallowing of the depositional environment. The paleocurrent directions indicate that these rocks were mostly sourced from the south and west; the relative mineralogical maturity of the sandstones suggests that they were derived from the older sedimentary rocks of the Bendigo Zone and the Mathinna Supergroup (Powell *et al.* 2003). Only in the southeastern part of the Melbourne Zone, the Mount Easton Province, were the Middle Devonian sediments sourced from the now adjacent Wagga-Omeo Complex. The youngest sedimentary rocks in the western part of the

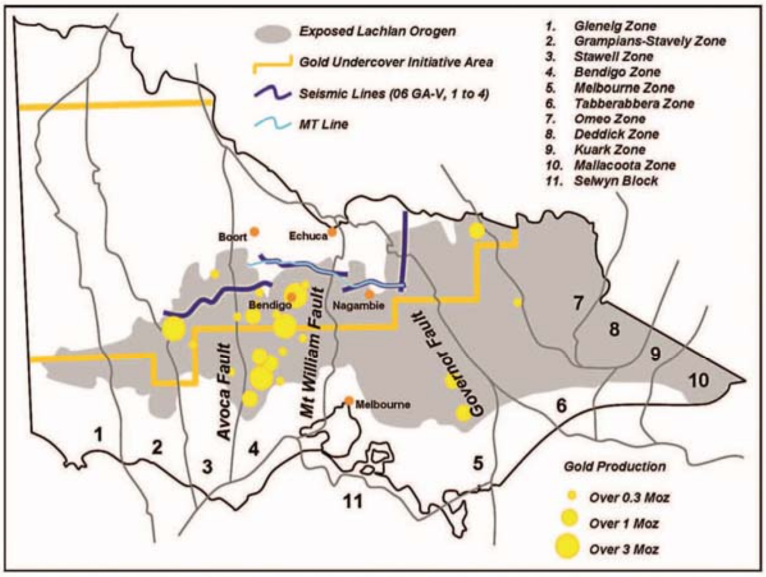


Figure 1 Map of Victoria showing the location of the survey lines. Also defined are the major structural zones of the western Lachlan Orogen and the faults crossed by the traverses. The yellow line encompasses the key initiative area for gold undercover exploration as defined by GSV with current gold production totals also indicated (after Birch & VandenBerg 2003; Department of Primary Industries, Victoria 2008).

Melbourne Zone were deposited in shallow marine or terrestrial environments, while in the east deep water conditions continued. In the west, the stratigraphic thickness reaches about 10 km, while in the east only about 3 km is exposed (VandenBerg *et al.* 2000).

Sedimentation in the Melbourne Zone was terminated by the Upper Devonian Tabberabbean Orogeny. Except in the east, the resultant folding was much more open in style than the chevron folds of the Bendigo Zone, with interlimb angles of 70° to 100° and shortening of about 25% (Gray 1995). In the easternmost Melbourne Zone, shortening increases to more than 70% in tight isoclinally folded rocks (Gray 1995). Two generations of open folds are present in the northern Melbourne Zone; an older east–west set is overprinted by a younger northwest–southeast-trending set, giving the outcrops a dome and basin pattern (Gray & Mortimer 1996).

A major period of igneous activity followed the Tabberabberan Orogeny. The earliest event was the intrusion of the mantle-derived Woods Point Dyke swarm, probably at about 377 Ma (Bierlein *et al.* 2001). Subsequently high level granites were intruded and calderas formed, particularly in the central Melbourne Zone but extending into the adjacent Bendigo and Tabberabbera Zones (e.g. VandenBerg *et al.* 2000). These granites are geochemically distinctive, with higher Ba contents, lower initial $^{87}Sr/^{86}Sr$ ratios and lower oxygen fugacities than the adjacent granites, suggesting a basement more like that in Tasmania (Chappell *et al.* 1988; Gray 1990; Rossiter & Gray 2008). The gold of the Woods Point-Walhalla region was introduced at about this time, after the intrusion of the dykes but before the granites.

Subsequently, the basins of the Howitt Province formed over the eastern edge of the Melbourne Zone, the Governor Fault. These basins are structural remnants of a wider sedimentary and volcanic sequence at least 3 km thick that was deposited before and during the Early Carboniferous Kanimblan Orogeny (VandenBerg *et al.* 2000).

The 2006 seismic survey showed that the crust below the Melbourne Zone is significantly different to that below the Bendigo Zone (Korsch *et al.* 2008; Cayley *et al.* in press). The Bendigo Zone shows steeply dipping upper crustal features while the midcrust has a strong, shallowly west-dipping character and a poorly defined MOHO. By contrast, the Melbourne Zone has strongly defined sub-horizontal layers at ~5 s two-way time, about 15 km deep, and 7 to 8 s two-way time, about 21 to 24 km deep. As well, it has a well-defined MOHO at ~12 s two-way time, about 36 km deep. These changes in character take place at the down-dip extension of the HFZ.

Interpretation of the 2006 deep seismic survey suggests that the Cambrian felsic volcanic rocks that crop out in the southeastern part of the Melbourne Zone formed the upper parts of this deep crustal block (Cayley *et al.* in press). This deep crustal unit has previously been named the Selwyn Block by Cayley *et al.* (2002). The Selwyn Block is believed to be the northwards extension of the pre-Ordovician Tasmanian cratonic basement. In Tasmania, this basement includes Cambrian felsic and ultramafic volcanic rocks and Neoproterozoic and Mesoproterozoic metasedimentary and granitic rocks (Seymour *et al.* 2007).

Given the geometry in the Tabberabberan Orogeny of 25 to 70% shortening of the Melbourne Zone rocks over the rigid Mesoproterozoic to Cambrian Selwyn Block, it seems highly likely that a décollement formed near the contact between the two packages. The Mount Easton Shale may have been involved; if so it is likely that it has significant thickness variations, having acted as a ductile layer infilling gaps between and under more competent strata.

A few hundred metres of Upper Carboniferous to Permian glacial deposits are preserved in the deeper paleovalleys over the Bendigo and Melbourne Zones (Moore & McLean 2009). Subsequently almost all of the northern parts of the Bendigo and Melbourne Zones were covered by up to 300 m of Cenozoic Murray Basin sediments.

THEORETICAL CONCEPTS OF THE MT METHOD

Magnetotellurics is a passive geophysical technique used for determining the electrical conductivity structure of the Earth's crust and upper mantle by measurement of the time-varying electric (E) and magnetic (H) fields at the surface (Cagniard 1952; Vozoff 1972, 1991; Simpson & Bahr 2005). The method has been successful for applications both in crustal exploration (identification of conductive anomalies associated with petroleum and ore bodies), the mapping of shallow geological structure (fault recognition), and in inferring deep electrical conductivity distribution and temperature regimes in the upper mantle. The basic principle of the MT method is based upon the diffusion of the EM fields from the relatively non-conductive air into the comparatively high conductive Earth. These naturally occurring fields are due to telluric currents (electrical currents which flow beneath the ground) and the fluctuating magnetic fields which induce them. The typical frequency range of an MT sounding is 10^{-3} to 20 000 Hz (Keller 1970). Above 1 Hz, worldwide thunderstorms in the lower atmosphere generate radio signals from lightening strikes known as sferics. Frequencies below 1 Hz are primarily due to solar activity interacting with the magnetosphere. The amplitude and phase relationships of the fields at the surface are dependent on the electrical properties of the subsurface within which the EM fields propagate. Therefore, by monitoring the fluctuating EM fields at the surface we can investigate the underlying conductivity structure. Depth of investigation is frequency dependent with longer periods penetrating to greater depths (Vozoff 1972).

The mathematical treatment of the MT method (fundamentally built upon Maxwell's Laws of electromagnetic induction) aims to reconstruct the subsurface conductivity from surface impedance data, measured over a range of frequencies. Time-varying, uniform, horizontal sheets of electrical current are present above the Earth's surface; these are commonly referred to in the literature as plane waves, normally incident to the surface of the Earth. The displacement currents are, however, negligible compared with the time-varying conduction currents, thus resulting in a quasi-static induction field governed by the equations of diffusion rather than wave propagation (Simpson & Bahr 2005).

The apparent resistivity (ρ) at the surface for a homogeneous half-space can be written in terms of the EM field vectors **E** and **H**:

$$\rho = \frac{1}{\omega\mu}\left|\frac{\mathrm{E}}{\mathrm{H}}\right|^2 \tag{1}$$

where $\omega = 2\pi f$, and thus defines the frequency of the propagating waves, and μ is the magnetic permeability of the Earth. The ratio of the field vectors is termed the characteristic impedance; a measure of the opposition experienced by an alternating current flowing within a circuit or conductive medium, it is related to the resistivity of the Earth at any depth. If the Earth's electrical properties are seen to vary with depth, then the observed values of apparent resistivity will vary with frequency. We can equate the relationship between resistivity (ρ, unit's Ωm), skin-depth (δ, in metres) and frequency (f, in Hz) by the expression:

$$\delta = \sqrt{\frac{\rho}{\pi f \mu}} \tag{2}$$

This, therefore, defines the skin-depth of the EM signal propagating within a section. It is an indication of the maximum depth of penetration, defined as the depth at which the field has fallen to e^{-1} (approx 37%) of its original magnitude. Directly associated with the skin-depth, the depth of penetration, is defined as the maximum depth at which a recognisable signal maybe resolved, approximated by Spies (1989) as being in the order of 1.5 skin depths.

The final Earth response parameter required to completely define the MT response, is the phase difference (θ) between **E** and **H**. Normally for a 1-D Earth, there is a phase difference of 45° between **E** and **H**. More complex, multidimensional, properties generate a corresponding rotation in relative phase at each frequency. Various algorithms are applied to calculate these Earth responses from the field measurements, which in turn can be used to determine the conductivity structure of the underlying section. MT data processing routines typically include Fourier transformation of the time-series field data to generate apparent resistivity and phase curves, and noise suppression, via robust processing with the inclusion of a remote reference site to overcome local noise.

DATA ACQUISITION AND PROCESSING

Measurements of **E** (E_x, E_y) and **H** (H_x, H_y) were made in orthogonal horizontal directions at the surface with a fifth component H_z (where positive-z is vertically downwards) recording the vertical magnetic field. The electric field was recorded via two sets of non-polarisable porous pot electrodes containing a $PbCl_2$ solution, connected via a 50 m cable and orientated in a N–S/E–W cross. With the two horizontal coils parallel with the alignment of the electrodes, the magnetic field was measured via three orthogonal coils (consisting of a ferrite core 1.5 m in length, wound with wire), which were buried to eliminate noise from wind motion.

Five component MTU-5A systems manufactured by Phoenix Geophysics, Ltd were installed at several (usually six or seven) sites simultaneously, with an operating frequency band ranging from 320 to approximately 0.00034 Hz. Measurements were taken over a 15 h period each day, starting at 17:00 and finishing at 08:00 the next day in local time. The remote reference site was situated near Swan Hill; approximately 120 km from the nearest sounding site, it was operated continuously throughout the survey and was synchronised with the field sites via inbuilt GPS units.

Inversion Routines

The inversion of MT data is generally non-unique and any model resolved is only one of many possible models that fit the data. The aim of the inversion process is therefore to produce a simple and smooth model that can be confidently interpreted alongside *a-priori* geological information (Constable *et al.* 1987). Consequently, the preliminary analysis of this dataset (as presented in the Gold Undercover Reports) was conducted using two independent algorithms in order to provide two comparative models and increase confidence in the interpretation of the section. Based on a linearised least-squares fit with smoothness regularisation for 2-D inversions, both algorithms incorporate the finite-element method for forward modelling in the inversion stage, but adopt two different schemes to determine the regularisation parameters, these are: Akaike Basian Information Criteria (Uchida 1993) and Active Constraint balancing (Yi *et al.* 2003; Lee *et al.* 2009). Following the completion of the TEM transect (Dennis *et al.* 2010) and applying statics corrections to the dataset, the original 2-D sections were used as an a priori guide for generating new models. Produced using the WinGLink® software package for MT data (distributed by Geosystem SRL) comprehensive details of the inversion routine implemented are available in Rodi & Mackie (2001).

STATIC SHIFTS

The complication presented by static shifts on an MT dataset has been a subject of discussion by many workers, as the galvanic distortion it represents can have significant detrimental effects on inversion results. Its estimation and removal is therefore a critical step in the MT inversion process, as it can significantly affect the resistivities and depths resolved by the model (Tournerie *et al.* 2004). Numerous correction schemes have been presented with Simpson & Bahr (2005) classifying the various methods into three broad groups: (i) averaging (statistical) techniques, (ii) long-period corrections relying on assumed deep structure or long-period magnetic transfer functions, and (iii) short-period corrections which rely upon active near-surface measurements. The latter of these methods, the use of additional short-period data to resolve the near surface structure, is the scheme which we have implemented here; specifically using data from the overlaying TEM transect (TEM results presented in Dennis *et al.* 2010).

The phenomenon of the static offset occurs in situations where there exists a surficial heterogeneity situated in the vicinity of an MT sounding site at depths less than approximately 1.5 skin depths of the highest frequency (Spies 1989). These small-scale anomalous bodies (where small is defined by Sternberg *et al.* (1985) as a body whose dimensions are much less than the skin depth at the highest recorded frequency) cannot be resolved directly by the MT method, as they are too shallow and thus outside the available signal range. The typical sounding depth for the TEM method, however, resides within the top few hundred meters, the region in which bodies imposing galvanic distortions in MT are located. A similar complication also arises if the survey region is host to a surface layer that is significantly greater in resistivity or conductivity than predicted (Sternberg *et al.* 1988). In this situation the entire transect (or a large section of it) may have undergone a shift, which due to the consistency of the offset and the shape of the curves remaining unaffected, is unlikely to be accounted for without supplementary data.

The distortion is caused by the buildup of a local electric field, generated by static charges produced by the background field, which accumulate at the rock interface of shallow heterogeneities (Pellerin & Hohmann 1990; MacNae *et al.* 1998). The result is that the apparent resistivity curve of the sounding site becomes shifted along the ρ-axis on a log-log plot of apparent resistivity (Ωm) *vs* frequency (*Hz*). A frequency independent effect (see Figure 2), this means that the entire curve is offset equally and does not undergo distortion, making it almost impossible to identify the presence of static shifts in MT data alone. The amount by which the curve is shifted is commonly referred to as the static shift factor (s) for a single sounding site. If not accounted for then the determined apparent resistivity will be out by a factor of s and the depth at which a body resides incorrect by the square root of s, i.e. layer thicknesses are multiplied by $s^{1/2}$ in the 1-D model (Simpson & Bahr 2005). In flat sedimentary environments, free from faults and intrusions, the displacement of the MT sounding curves is not usually large; however, in hard rock environments, as typically encountered in mineral exploration, shift factors commonly exceed a factor of 10 (MacNae *et al.* 1998).

As the TEM method relies upon the measurement of an artificially induced magnetic field it is virtually uninfluenced by the galvanic distortion caused by the charge build up. The time-domain TEM decay curve can be easily converted into an equivalent frequency-domain apparent resistivity curve, by dividing the transient time-scale (given in ms) by 194 and then comparing the resulting curve to the MT apparent resistivity plot (time-scale in seconds) (Sternberg *et al.* 1988). By using this simple conversion factor the equivalent TEM apparent resistivity curves can be directly compared with the corresponding MT curves

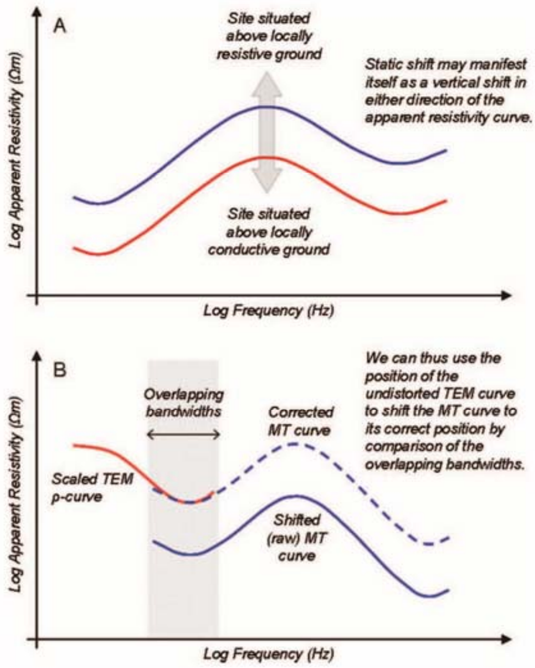


Figure 2 MT apparent resistivity curves demonstrating influence of a static offset. (a) Direction of the offset induced by locally conductive/resistive ground; (b) illustration of how TEM data with an equivalent overlapping bandwidth can be used to correct the location of the MT apparent resistivity curve (adapted from results presented in Sternberg *et al.* 1988).

(as illustrated by Figure 2), and the shift factors can subsequently be determined by consideration of the vertical offset between the TEM curve and the two separate MT modes.

Determination of Shift Factors

Coincident-loop TEM data were collected concurrent with all 53 MT sites identified as Bendigo line 2 in the MT08 report (Dennis *et al.* 2009) (also see Figure 1; MT and ground TEM lines are coincident). In areas of uncomplicated geology both central-loop and coincident-loop TEM soundings generate results, that when converted to frequency domain equivalents, can easily be compared with MT datasets (Meju 1998). Shift factors were subsequently determined individually for each of the MT modes at all sites. The value of s determined is a multiplication factor for the MT curves; a value of $s = 1$ corresponds to no static shift identified, a value of <1 indicates a resistive anomaly, and a value >1 corresponds to a conductive heterogeneity. One would assume therefore that over a large series of readings, shift factors of greater or less than one would occur with equal probability. Approximately three-quarters of the shift factors determined for the line are greater than or equal to 1; this is consistent for an area where conductive overburden is present, in this case, Murray Basin sediments.

Figure 3 presents a selection of the apparent resistivity curves for the dataset, showing plots for both the MT and TEM methods. The sites depicted are representative of four distinct distortions which may occur, while highlighting the variability of the magnitude and direction of a static offset even within a single transect. As illustrated, site 2, situated at the foot of a large granite outcrop, shows a large upward shift of both MT curves. This result is typical for a sounding made where a surficial resistive body is present, in this case an isolated igneous intrusion. The locally resistive ground at this site has also resulted in no overlap of the bandwidths available; this is due to the rapid decay of the TEM signal in highly resistive Earth. The raw apparent resistivity curves generated for site 10 show an excellent overlay without the requirement for shifting. The two MT modes also both coincide well, depicting an area that is predominantly 1-D in structure. This is in sharp contrast to the plot for site 28 where the high frequency splitting of the MT modes is indicative of near surface anisotropy or localised multi-dimensionality. The final site shown in Figure 3, site 35, shows both modes to have undergone a significant downward shift, a response consistent with the presence of a surficial conductor, or in this region, the more likely consequence of sedimentary conductive overburden. It can also be seen here that the gradients of the raw curves do not present an ideal match; further investigation by the primary author at this site and others with a similar response, has revealed superparamagnetic (SPM) effects (Buselli 1982; Nabighian & MacNae 1991) to have influenced the late-time TEM decays. If SPM effects are not removed then the resulting frequency equivalent TEM curve is seen to tend downwards in the plot, i.e. showing more conductive ground as shown in Figure 3.

INDUCTION VECTORS

By employing a five component recording system as described, measurement of the vertical magnetic field component was also made. The vertical (z-axis) magnetic field comes as a consequence of lateral conductivity variations present in the nearby Earth. These gradients can be pictorially represented by the use of induction vectors (also known as Parkinson arrows or tippers); depicted by arrows they originate at the sounding site and following the Parkinson convention, the real component of the vector points towards the high conductivity side of a local conductivity contrast (Hoffmann-Rothe *et al.* 2001). The length of the arrow is seen to decrease in length with horizontal distance from the anomaly, with factors such as depth of contrast, and its response to varying frequencies also an influence, the length is in general an indication of the magnitude of the absolute change in conductivity/resistivity, emphasising the source of a specific inhomogeneity (Murphy *et al.* 2008).

Induction arrows for the Bendigo transect are shown in Figure 4 for a number of frequencies, with the longer periods relating to greater depths. As illustrated, the high frequency vectors (above ~0.70 Hz) point in random directions, representing the inhomogeneities of the near surface. Below this the arrows can be seen to align, with two distinct trends appearing. To the east of ~300 km-E (approximate easting of the HFZ) the arrows appear to rotate inward towards the centre of the

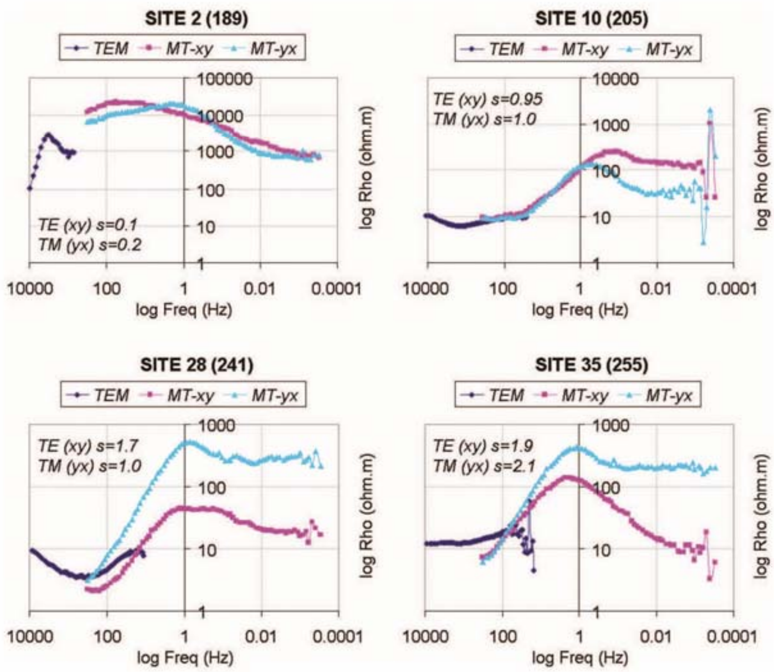


Figure 3 Apparent resistivity curves for MT and TEM data showing four distinct distortions. Site 2 shows an upward shift, typical in the presence of a surficial resistive anomaly; site 10 shows no significant offset; site 28 shows splitting of the MT mode characteristic of near surface anisotropy with the TE mode only significantly affected; and site 35, showing a downward shift of both modes indicative of conductive overburden. The number given in parentheses is the site number as given in the 2007 MT Report (Cull *et al.* 2008).

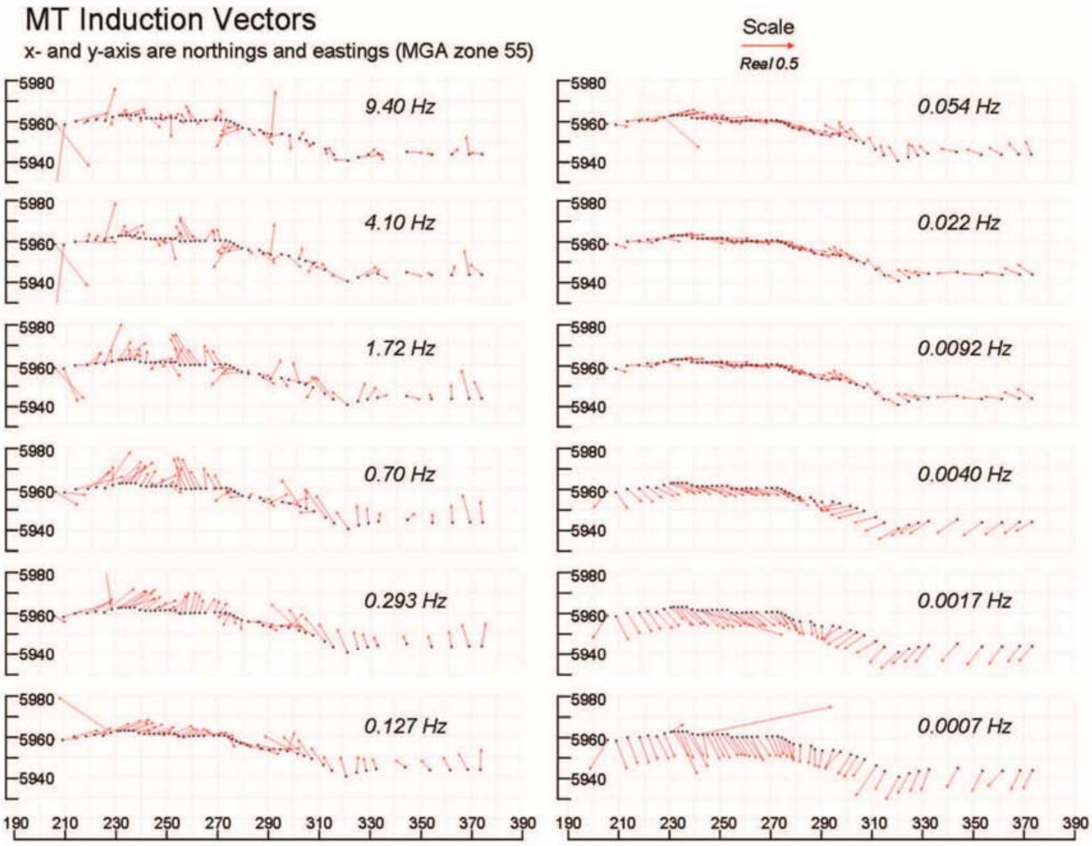


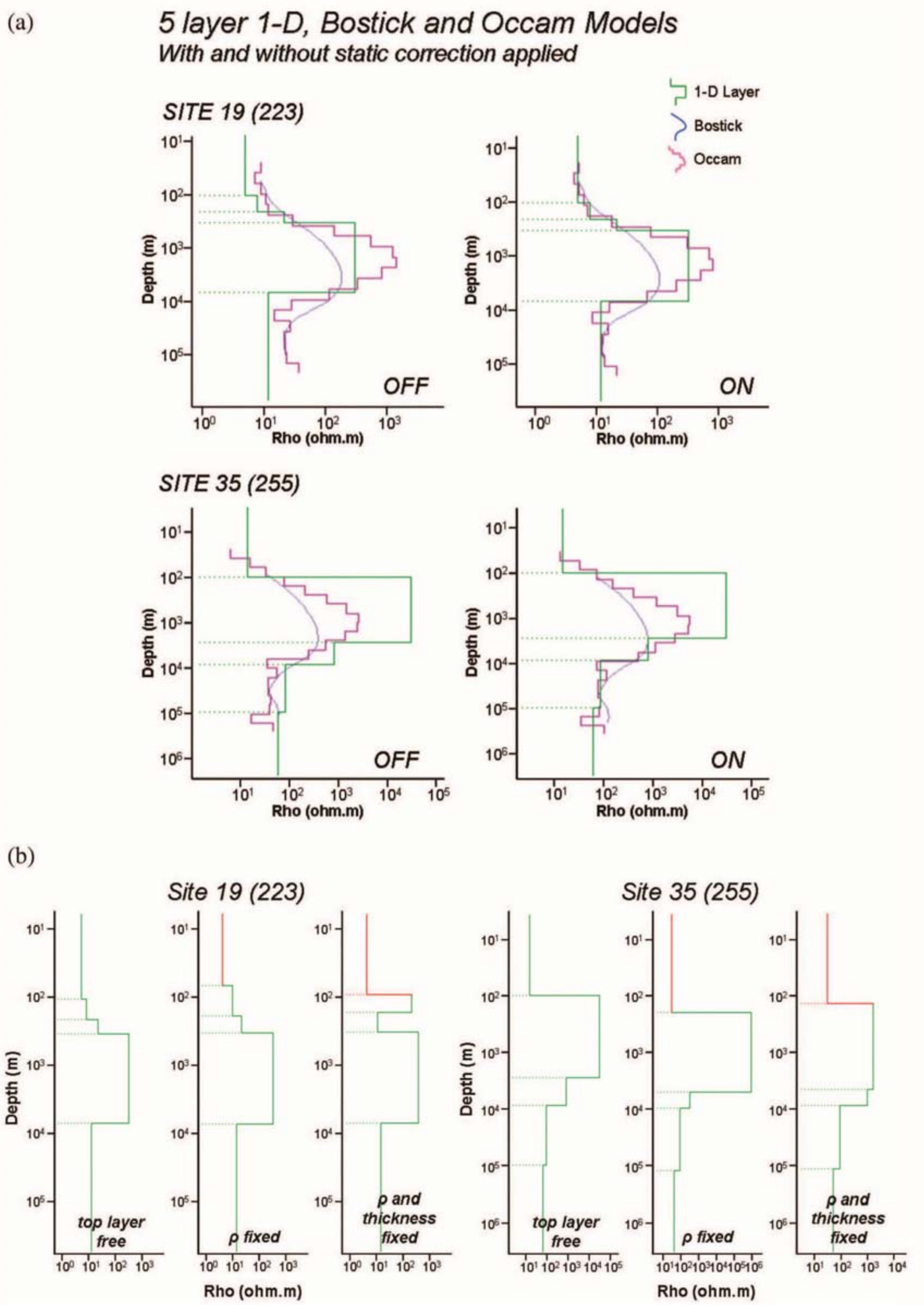
Figure 4 Induction vectors for various frequencies. Following the Parkinson convention, the induction arrows point towards regions of high conductivity, with each sounding site attributed to and marking the origin of an individual vector. The range of the induction-vector method (both in depth and horizontally) increases for long periods.

transect, an effect that is mirrored to the west of 300 km-E. With a strong northerly tendency this pattern becomes evident at around 0.239 Hz and can be defined to a varying extent at all lower frequencies depicted; swinging around from a northerly to southerly trend the arrows at 0.022 Hz are almost all along an east–west line, pointing at the HFZ. This trend consequently indicates that the geological boundary between the Bendigo and Melbourne Zones is one of high conductivity with considerable depth extent. As we move to periods of around 0.004 Hz and longer, all of the induction vectors (excluding two anomalous responses at 0.0007 Hz) are seen to take a southerly direction, presenting the occurrence of an off-line conductor situated at depth to the south of the line. A response which demonstrates the 'coast effect' described by Lilley (1976), it is consistent with and adds detail to the same observation made therein from earlier magnetometer array studies (Figure 2; Lilley 1976). In summary, therefore, from a randomised surficial response, the induction arrows can be seen to take on a northerly trend, which progressively rotates clockwise west of, and anti-clockwise to the east of ~300 km-E, to a final southerly direction (due to the conductive mass imposed by sea water) for long periods.

Other more localised responses can also be defined including a second conductive focus at short periods (1.72 to 0.70 Hz), off-line to the north of the transect and centred about 240 to 250 km-E, and a resistive anomaly with the arrows for the first two sites at the western end of the transect pointing away from each other over a wide frequency range (0.127 Hz to the lower limit). The location of the northern conductive focus coincides with the Whitelaw Fault region and thus may represent a shallow feature not dissimilar to the more prominent and further penetrating zone boundary at ~300 km-E. Whereas the resistive focus at around 205 km-E coincides with the location of a surficial granite body as identified by the annotated seismic reflection profile.

1-D MODELS

The 1-D interpretation of MT data is often used as a means to define a reasonable first approximation of the true conductivity-depth distribution beneath a sounding site (Jones 1983). Generated using WinGLink®, presented in Figure 5a are the 1-D models for two sites (19 and 35) along line 2 of the Bendigo line, for both shifted and non-shifted data. In addition to the 1-D layers,


(a)
5 layer 1-D, Bostick and Occam Models
With and without static correction applied
1-D Layer
Bostick
Occam
SITE 19 (223)
Depth (m)
Rho (ohm.m)
OFF
ON
SITE 35 (255)
OFF
ON
(b)
Site 19 (223)
Site 35 (255)
top layer free
ρ fixed
ρ and thickness fixed
Depth (m)
Rho (ohm.m)

Occam and Bostick curves are also shown and their relationship with the layered model provides an insight into how the statics are influencing the inversion results.

The Occam inversion (Constable *et al.* 1987) is a real multi-layer smoothness inversion, while the Bostick (Bostick 1977) curve is a simple transformation of the apparent resistivity curve to an apparent depth relation. It can be clearly seen that the statics corrections applied do not have a dramatic influence on the layer depths and thicknesses (also given in Table 1); however, the clear shifting of the Occam and Bostick curves that occurs in relation to the layered model (a common observation) demonstrates that the majority of the statics seen here are a near surface effect in this environment. As this surficial layer is effectively transparent to the MT sounding, static shifts are by nature of greater relevance to near surface structure and it is the shape (gradient) of the apparent resistivity curve rather than its absolute value that will have the most significant influence on the inversion results.

We can therefore conclude from the 1-D models that it is a high-frequency perturbation that is causing the shift; a near surface effect that does not significantly alter the gross layering approximation. This emphasises the subtle effect of the static shift on the deep inversion results, whereas constraints on the near surface, including structural correlations are much more evident. Additionally if we look at the Bostick curves (for site 35 in particular) the steep gradient of the model at high frequencies means that the surface layer may not actually be defined by the field results, rather it is an artefact of the modelling program when data are not available at these depths. Apparent resistivity curves which are not asymptotic reduce the validity of shift factors derived from the overlapping curves (Meju 1996) as the correct final level remains ambiguous. Comparing the upper most layer of the MT inversion to three-layer models derived for the TEM alone (using the GRENDL algorithm based on work by Raiche *et al.* 1985) we can confirm that the resistivity magnitude of this top layer is supported by the transient results, with the basement layer in the TEM indicating $\rho = 14.54$ Ωm at 79.93 m depth and $\rho = 25.46$ Ωm at 38.21 m, for sites 19 and 35 respectively.

We can subsequently demonstrate the sensitivity of the deep layered inversion relevant to this top layer by manually fixing the resistivity and thickness of the surficial layer to values determined from the TEM and then varying the restrictions applied to the parameters to assess the alteration induced at depth. Again using the 1-D inversion results for sites 19 and 35, Figure 5b shows the layered models that result for three different settings; with (a) the top layer unrestricted, with (b) the apparent resistivity of the top layer fixed and finally with (c) both ρ and the thickness (t) of this surficial layer held constant. The values at which to fix the surface layer are determined directly from the offset derived by comparison of the MT and TEM results. The shift factor (s) (also referred to as the scaling factor) is then used to recalculate the values from those determined from the unperturbed inversion. Table 2 subsequently shows the fixed (top-layer) values calculated from the shift factor along with the respective values for the deeper layers. We can see from these tabulated results that a change to the near surface apparent resistivity does in fact induce a change at all depths, as expected due to the frequency independent nature of the phenomenon.

Table 1 Five-layer 1-D inversion models with and without static shift applied. Apparent resistivities and layer thicknesses given for sites 19 and 35; the site number shown in parentheses is the site number as given in Cull *et al.* (2008).

Site	Static shift off		Static shift on	
	ρ (Ωm)	t (m)	ρ (Ωm)	t (m)
19 (223)	8.45	94.66	4.86	98.05
	10.16	104.31	7.54	95.90
	22.95	179.30	20.60	123.74
	515.29	8281.29	296.50	6225.92
	19.83	–	11.33	–
35 (255)	6.80	66.68	13.73	94.74
	20 294	1785.28	28 474	2559.57
	376.68	4045.88	759.04	5660.65
	39.41	73 760	80.72	84 441
	21.03	–	56.28	–

2-D INVERSION RESULTS

Figure 6 shows two inversion models produced for the dataset, one with and one without statics corrections applied. The individual shift factors derived for each independent mode have been used and the second (shifted) inversion executed with all other parameters maintained.

Considering the shifted model we can break down the regional electrical response into four broad sections. With an average penetration depth of ~120 m resolved by the TEM soundings, this thin conductive surface layer is present across the majority of the survey line and is subsequently underlain by a series of resistive regions that are intersected by a series of west dipping conductive slabs. At a depth of around 20 km the crustal depth resistive heterogeneities give way to a conductive layer. Most prominent to the western end of the survey line a disrupted conductive band can be seen to continue

◀

Figure 5 (a) 1-D inversion, Bostick and Occam models for invariant mode data. Five-layer 1-D inversion models generated and plotted alongside their corresponding Bostick and Occam results for comparison. Results shown for two sites where noise distortion was minimal and where reasonable large shift factors were displayed; site 19, $xy = 1.1$, $yx = 0.3$ and site 35, $xy = 1.9$, $yx = 2.1$. (b) 1-D layered models with varying top layer parameters. From left to right, the models show the inversion results for unshifted data with no layers fixed, apparent resistivity fixed to shifted value and with both the apparent resistivity and top layer thickness held to values after static corrections have been applied.

Table 2 Apparent resistivities and thicknesses for the five-layer, 1-D inversion models. The unfixed values are as determined directly from the inversion of the results with the static offset applied and where the held values (bold) have subsequently been calculated from this and the corresponding shift factors. The site number shown in parentheses is the site number as given in Cull *et al.* (2008).

Site	Unfixed		ρ held		ρ and t held	
	ρ (Ωm)	t (m)	ρ (Ωm)	t (m)	ρ (Ωm)	t (m)
19 (223)	4.86	98.05	**3.40**	62.28	**3.40**	**82.03**
	7.54	95.90	7.89	107.44	170.23	68.46
	20.60	123.74	18.12	138.90	8.73	147.66
	296.50	6225.92	278.01	6269.36	307.18	6238.54
	11.33	–	11.33	–	11.32	–
35 (255)	13.73	94.74	**27.46**	192.05	**27.46**	**133.98**
	28 474	2559.57	971657	4632.66	1476.20	4230.73
	759.04	5660.65	288.93	4349.03	915.91	3962.64
	80.72	84 441	77.37	109 412	79.42	101 533
	56.28	–	*38.76*	–	*44.34*	–

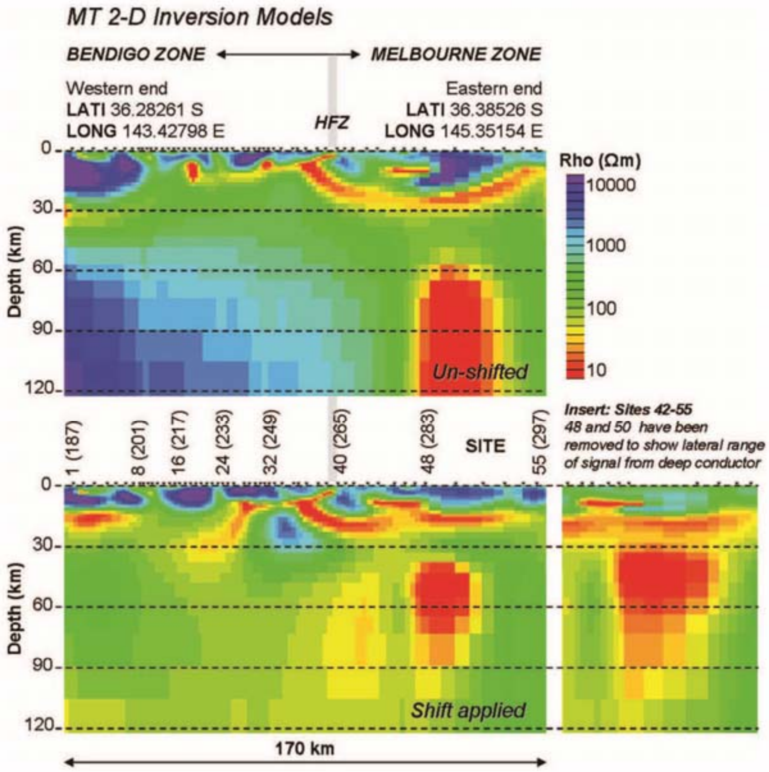


Figure 6 2-D MT inversion models for both shifted and unshifted data. Generated using WinGLink® software, a smooth 2-D inversion was completed for the raw, uncorrected data (top) and then with the shift factors applied to both modes (bottom). Sites 39 and 40 straddle the HFZ, number in brackets is site number as per 2007 MT Report, noting also that sites 49 and 51 only have TEM data and so are not marked on this section.

across the entire line, terminating at around 30 km depth consistent with a transition to upper mantle depths. Finally, moving down into the MOHO, the deep structure presented in these sections is of more uniform character than the faulted crust, with the inclusion of a significant conductive region located beneath the Melbourne Zone in the eastern section of the transect.

Deep Structure

The largest effect of applying static shift corrections is the loss of the original deep resistive region which dominates the western section of the line below ~60 km. The resistivity decreases from a few thousand to a few hundred Ωm, corresponding to an upward shift to the raw curves. In particular, sites 2 and 7 demonstrate values of s much less than unity, with values of 0.1 and 0.2 for site 2, and 0.7 and 0.3 for site 7 determined for the TE and TM modes respectively.

Also for deep sections situated below sites 47–52, a significantly conductive region can be identified. With a resistivity of approximately 10 Ωm in a host region with resistivities closer to 100 Ωm, this body is located at a depth that correlates with seismic anomalies detected in a teleseismic tomography analysis by Graeber *et al.* (2002), covering an adjacent and slightly overlapping (western end of MT07 line) area within Victoria. The new MT models presented here provide a significant extension to the depth resolved in the Gold Undercover reports and are thus a first look at the electrical structure of the upper mantle in this region. Additional geothermal modelling is required to extend and confirm these findings; however this anomalous conductive zone possibly indicates a deep heat source in the region which extends from the mantle to shallow depths. The apparent depth extent of the anomalous seismic zone is reduced by applying a statics correction to the corresponding MT data which also indicates that the upper boundary is around 40 km depth (as opposed to 60 km in the unshifted MT model) yet without a reduction in the magnitude of the contrasting resistivity.

Also shown by the insert to Figure 6, is this particular region remodelled after the removal of sites 48 and 50 (sites 49 and 51 only have TEM data available and so are excluded from the MT inversions). This additional inversion, with the two sites removed, has been completed in order to demonstrate the significance of the anomaly and ensure it is not a localised artefact specific to one or two sites. Situated at a depth that is deeper than the focus of this paper, further investigation of this body is planned; new heat flow data have been obtained for this purpose and will be the subject of a separate report in consultation with GeoScience Victoria and Geogen P/L (Aivazpour & Cull, personal communication).

Crustal Structure

The Gold Undercover initiative is focused on exploration at crustal depths and thus the implications of these new models are of much greater significance for depths less than approximately 30 km. The gross structure, including the resolved conductivities, is not dramatically altered by the application of shift corrections; however the surficial structure has undergone significant distortion. For fault mapping, a major objective of these transects, such alterations which appear subtle on a regional scale, pose significant issues in matching the electrical anomalies with known structures.

Near sites 39 and 40, east of the HFZ, the crustal structure is dominated by an underlying conductive layer extending across a significant portion of the Melbourne Zone. At a nominal depth of ~20 km, much like the deep conductor below it, this conductor moves upwards after statics corrections. This movement is consistent with the presence of a conductive surface layer, with derived shift factors greater than 1, the multiplier (square root of *s*) results in thicker layers derived from the raw data. This is a general trend seen across a large part of the transect, with an effect observed to both conductive and resistive heterogeneities.

Centred near site 25 an additional conductive band has been resolved in the second model. With a resistivity of a few tens Ωm and extending to a depth of ~30 km this section and the immediate surroundings (sites 13–32) contrast with most of the traverse, in that the shifted model does not appear to have been shortened in terms of vertical thickness; rather, a possible expansion can be inferred. The west-dipping conductive band may be interpreted to be an extension of the deep-crustal conductive layer, prominent in the Melbourne Zone. However, due to its dip orientation, there is strong evidence to support that this electrically conductive region may be associated with the location of the known faults in the region, but overlain by resistive outcroppings.

SEISMIC COMPARISON

Reflection seismology has long since been the primary technique used for the resolution of deep crustal structures. The recent advancements in MT, however, have presented a viable cost-effective and logistically easier alternative; with complimentary resolvable depths a joint analysis is also a feasible option (Key *et al.* 2006).

Figure 7 presents the new static corrected MT model alongside the corresponding seismic reflection profile from the 2006 Central Victoria Seismic Survey (ANSIR 2008). Annotation and analysis of the seismic line have previously been completed by GSV, and the faults and form lines identified have been retained here. Also shown in Figure 7 is an unannotated version of the seismic profile, allowing for a comparison unbiased by structure derived from the seismic analysis. A general trend can be seen for the entire section that where a conductive zone has been resolved, an increase in the number of reflections is also evident. This is of particular significance in terms of fault mapping including the determination of dip angles at depth. It is also interesting to note from this comparison that there is no indication in the seismic section of the mantle depth conductive anomaly resolved by the MT.

Comparing the cross-sections from west to east we can see several features which are distinct in both profiles. The Campbelltown and Muckleford Faults reach the surface just west of the outcropping Wedderburn Granite. With an initially high angle dip these flatten with depth and the annotated reflections become more tightly packed. On the unannotated seismic profile this region shows a high density response, which correlates directly with a zone of high conductivity in the MT that follows a similar shaped path to the surface.

Further east, this highly reflective zone shallows. However, a line of strong reflections is maintained, which merges into what has been identified as the deep extension of the HFZ at about site 40. The region between this line and the surface is the quietest in terms of seismic response and has been resolved as a zone that is predominantly electrically resistive. The high resistivity bodies, however, are intersected at several locations by resistivity lulls, which include where the Whitelaw and Lockington Faults approach the surface. The location of the Whitelaw Fault is also indicated by a short line of increased reflectivity, below site 20 and west dipping it aligns very closely with the

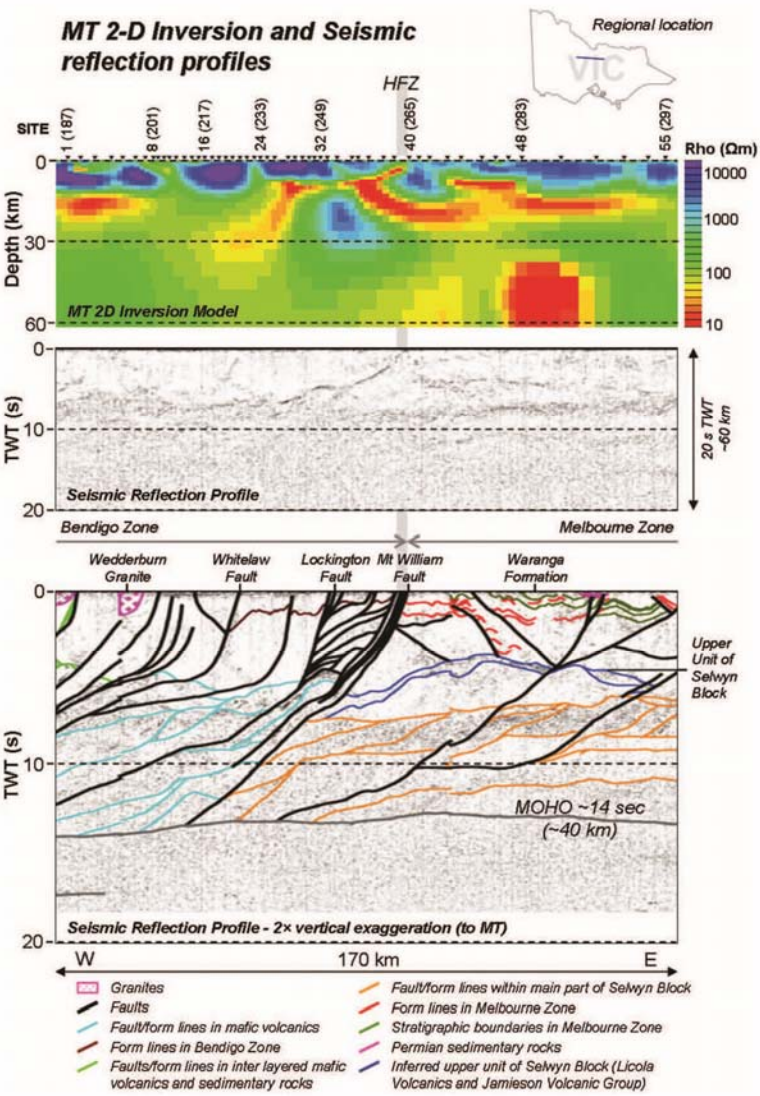


Figure 7 2-D MT inversion model and seismic reflection profile comparison. MT and unannotated seismic sections are presented with the same depth scale, with the vertical axis exaggerated (×2) for clarity on the annotated section. The fault and form lines shown are from the Central Victoria Seismic Survey Report (ANSIR 2006).

eastern edge of the resistive body dominating the near surface between sites 12 and 22.

The HFZ appears in the seismic image as a set of prominent west dipping reflectors. The electrical response is also prominent, with the conductive fault zone showing the same dip and orientation in both profiles. The unannotated seismic section also clearly shows an inflection in the reflection line below site 32; this again follows the conductive zone which can be seen extending to the MOHO.

Both techniques show that the gross structure in the Melbourne Zone is primarily flat lying, in contrast to the west dipping faulted character in the Bendigo Zone. The dominant feature of the subsurface in the Melbourne Zone is the flat lying conductor, which at a depth of ~20 km, corresponds in position and thickness to the upper unit of the Selwyn Block. Shown by a layer of high reflectivity in the seismic data a similar, second horizontal conductive layer is observed between sites 44 and 47. Extending to shallower depths and residing above the main conductive body, this is also indicated by an increase in the reflection density.

SYNTHESIS AND CONCLUSIONS

The application of statics corrections to the MT dataset has resulted in a new resistivity model of the crust, which presents both a significant match with the existing seismic results and enhanced resolution of the surficial structure. With the transect located close to the southerly limit of the Murray Basin, resistive outcrops are still present; therefore assuming a single near surface conductivity level to shift the MT, would induce

The top of the Selwyn Block identified in the seismic section can also be recognised in the MT response as a flat-lying conductor at around 15–20 km below the present ground surface. The conductive nature of this unit is consistent with the graphitic character of the Mt Eastern Shale, the lowest sedimentary unit of the Melbourne Zone, and which is believed to extend across the top of the Selwyn Block (Cayley *et al.* in press).

ACKNOWLEDGEMENTS

MT data were collected as part of a collaborative project between Monash University, GeoScience Victoria (GSV), the Korean Institute of Geoscience and Mineral Resources (KIGAM) and the Geological Survey of Japan (GSJ), including research funding from KIGAM and GSJ. TEM data were collected in whole by the primary author with assistance from Megan Hough and Jonathon Louey-Gung of Monash University. Comments and suggestions for the manuscript were also gratefully received from Dr Toshihiro Uchida at GSJ, with thanks also to the reviewers of the paper, F. E. M. (Ted) Lilley and one other, whose constructive and helpful comments were much appreciated.

REFERENCES

ANSIR. 2008. L178 Central Victoria Seismic Survey Results 2006. Migrated & interpreted migrated sections, and supporting data. National Research Facility for Earth Sounding, GeoScience Australia—*Data CD*.

Bierlein F. P., Arne D. C., Foster D. A. & Reynolds P. 2001. A geochronological framework for orogenic gold mineralisation in central Victoria, Australia. *Mineralium Deposita* **36**, 741–767.

Birch W. D. & VandenBerg A. H. M. 2003. Continuing the advance of Victorian Geology. *In*: Birch W. D. ed. *Geology of Victoria*, 1–13. Geological Society of Australia (Victoria division)—Special Publication, **23**.

Bostick F. X. 1977. A simple almost exact method of MT analysis: Workshop on electrical methods in geothermal exploration—U.S. Geological Survey, contract No. 14080001-8-359, Utah 1976.

Buselli G. 1982. The effect of near-surface superparamagnetic material on electromagnetic measurements. *Geophysics* **47**, 1315–1324.

Cagniard L. 1952. Basic theory of the magneto-telluric method of geophysical prospecting. *Geophysics* **18**, 605–635.

Cayley R. A., Taylor D. H., VandenBerg A. H. M. & Moore D. H. 2002. Proterozoic—Early Paleozoic rocks and the Tyennan Orogeny in central Victoria: the Selwyn Block and its tectonic implications. *Australian Journal of Earth Science* **49**, 225–254.

Cayley R. A., Korsch R. J., Moore D. H., Costelloe R. D., Nakamura A., Willman C. E., Rawling T. J., Morand V. J., Skladzien P. B. & O'Shea P. J. (2011 in press) Crustal architecture of Central Victoria: results from the 2006 deep crustal reflection seismic survey. *Australian Journal of Earth Sciences* **58**.

Chappell B. W., White A. J. R. & Hine R. 1988. Granite provinces and basement terranes in the Lachlan Fold Belt, southeastern Australia. *Australian Journal of Earth Sciences* **35**, 505–521.

Constable S. C., Parker R. L. & Constable C. G. 1987. Occam's inversion: A practical algorithm for generating smooth models from electromagnetic sounding data. *Geophysics* **52**, 289–300.

Cull J., Lee S. K., Lee T. J., Uchida T., Park I. W. & Song Y. 2008. A magnetotelluric survey along part of the 2006 central Victorian seismic transect. *GeoScience Victoria Gold Undercover Report 8*—Department of Primary Industries.

Dennis Z. R., Cull J. P. & Cayley R. A. 2010. Transient electromagnetic profiles across gold mineralisation zones in central Victoria. *Australian Journal of Earth Sciences* **57**, 317–328.

Dennis Z. R., Cull J. P., Lee S. K., Lee T. J., Park I. W., Song Y. & Uchida T. 2009. A magnetotelluric survey to the north of the 2006 central Victorian seismic transect: final report. *GeoScience Victoria Gold Undercover Report 18*—Department of Primary Industries.

Department of Primary Industries, Victoria. 2008. Gold undercover—Underneath Victoria's Gold. Online <http://www.dpi.vic.gov.au/DPI/nrenmp.nsf/fid/EFE1904454BD74C3CA2574B9000C499A>

Glen R. A. 2005. *The Tasmanides of eastern Australia—terrane processes at the Margins of Gondwana*. Geological Society of London Special Publication. **246**, 23–96.

Graeber F. M., Houseman G. A. & Greenhalgh S. A. 2002. Regional teleseismic tomography of the western Lachlan Orogen and the Newer Volcanic Province, southeast Australia. *Geophysical Journal International* **149**, 249–266.

Gray C. M. 1990. A strontium isotope traverse across the granitic rocks of southeastern Australia: petrographic and tectonic implications. *Australian Journal of Earth Sciences* **37**, 331–349.

Gray D. R. 1995. Thrust kinematics and transposition fabrics from a basal detachment zone, eastern Australia. *Journal of Structural Geology* **12**, 1637–1654.

Gray D. R. & Foster D. A. 1998. Character and kinematics of faults within the turbidite-dominated Lachlan Orogen: Implications for tectonic evolution of eastern Australia. *Journal of Structural Geology* **20**, 1691–1720.

Gray D. R. & Mortimer L. 1996. Implications of overprinting deformations and fold interference patterns in the Melbourne Zone, Lachlan Fold Belt. *Australian Journal of Earth Sciences* **43**, 103–114.

Gray D. R. & Willman C. E. 1991. Deformation in the Ballarat Slate Belt, central Victoria, and implications for the crustal structure across southeast Australia. *Australian Journal of Earth Sciences* **38**, 171–201.

Gray D. R., Wilson C. J. L. & Barton T. J. 1991. Intracrustal detachments and implications for crustal evolution within the Lachlan fold belt, southeastern Australia. *Geology* **19**, 574–577.

Hoffmann-Rothe A., Ritter O. & Haak V. 2001. Magnetotelluric and geomagnetic modelling reveals zones of very high electrical conductivity in the upper crust of Central Java. *Physics of the Earth and Planetary Interiors* **124**, 131–151.

Hough M. A., Bierlein F. P. & Wilde A. R. 2007. A review of the metallogeny and tectonics of the Lachlan Orogen. *Mineralium Deposita* **42**, 435–448.

Jones A. G. 1983. Letter to the editor—On the equivalence of the 'Niblett' and 'Bostick' transformations in the magnetotelluric method. *Journal of Geophysics* **53**, 72–73.

Keller G. V. 1970. Natural-field and controlled-source methods in electromagnetic exploration. *Geoexploration* **9**, 99–147.

Key K. W., Constable S. C. & Weiss C. J. 2006. Mapping 3D salt using the 2D marine magnetotelluric method: Case study from Gemini Prospect, Gulf of Mexico. *Geophysics* **71**, B17–B27.

Korsch R. J., Barton T. J., Gray D. R., Owen A. J. & Foster D. A. 2002. Geological interpretation of a deep seismic reflection transect across the boundary between the Delamerian and Lachlan orogens, in the vicinity of The Grampians, Western Victoria. *Australian Journal of Earth Sciences* **49**, 1057–1075.

Korsch R. J., Moore D. H., Cayley R. A., Costelloe R. D., Nakamura A., Willman C. E., Rawling T. J., Morand V. J. & O'Shea P. J. 2008. Crustal architecture of Central Victoria: results from the 2006 deep crustal reflection seismic survey—new generation advances in geoscience. Abstract No. 89 of the 19th Australian Geological Convention, Perth Convention and Exhibition Centre, July 2008, p. 155.

Lee S. K., Kim H. J., Song Y. & Lee C-K. 2009. MT2DInvMatlab—A program in MATLAB and FORTRAN for two-dimensional magnetotelluric inversion. *Computers and Geosciences* **35**, 1722–1734.

Lilley F. E. M. 1976. A magnetometer array study across southern Victoria and the Bass Straight area, Australia. *Geophysical Journal Royal Astronomical Society* **46**, 165–184.

Lisitsin V., Olshina A., Moore D. H. & Willman C. E. 2007. Assessment of undiscovered mesozonal orogenic gold endowment under cover in the northern part of the Bendigo Zone. *Geoscience Victoria Gold Undercover Report 2*—Department of Primary Industries.

MacNae J., Lay L. & Weston L. 1998. Measurement of static shift in MT and CSAMT surveys. *Exploration Geophysics* **29**, 494–498.

Meju M. A. 1996. Joint inversion of TEM and distorted MT soundings: Some effective practical considerations. *Geophysics* **61**, 56–65.

Meju M. A. 1998. A simple method of transient electromagnetic data analysis. *Geophysics* **63**, 405–410.

Moore D. H. & McLean M. A. 2009. The northern Bendigo Zone: Interpretation of the Bendigo-Mitiamo gravity survey, including depths to basement. *Geoscience Victoria Gold Undercover Report 11*—Department of Primary Industries.

Murphy F. C., Cull J., Lee S. K. & Lee Y. 2008. Magnetotelluric soundings and crustal architecture at Century mine, northern Australia. *Australian Journal of Earth Sciences* **55**, 655–667.

Nabighian M. N. & MacNae J. C. 1991. Time domain electromagnetic prospecting methods. In: M. N. Nabighian ed. *Electromagnetic Methods in Applied Geophysics—Vol. 2*, pp. 427–520. Society of Exploration Geophysics, Tulsa, OK.

O'Shea P. J. 2008. Seismic Shift: Victoria's seismic transect update. *Discovery—Victoria's Earth Resources Journal*, April 2008, 18–19.

Pellerin L. & Hohmann G. W. 1990. Transient electromagnetic inversion: A remedy for magnetotelluric static shifts. *Geophysics* **55**, 1242–1250.

Phillips G. N. & Hughes M. J. 1996. The geology and gold deposits of the Victorian gold province. *Ore Geology Reviews* **11**, 255–302.

Powell C. McA., Baillie P. W. & VandenBerg A. H. M. 2003. Silurian to mid-Devonian basin development of the Melbourne Zone, Lachlan Fold Belt, southeastern Australia. *Tectonophysics* **375**, 9–36.

Raiche A. P., Jupp D. L. B., Rutter H. & Vozoff K. 1985. The joint use of coincident loop transient electromagnetic and Schlumberger sounding to resolve layered structures. *Geophysics* **50**, 1618–1627.

Ramsay W. R. H., Bierlein F. P., Arne D. C. & VandenBerg A. H. M. 1998. Turbidite-hosted gold deposits of Central Victoria, Australia: their regional setting, mineralising styles, and some genetic constraints. *Ore Geology Reviews* **13**, 131–151.

Ritter O., Hoffmann-Rothe A., Bedrosian P. A., Weckmann U. & Haak V. 2005. *Electrical conductivity images of active and fossil fault zones: high-strain zones—structure and physical properties. Geological Society, London, Special Publication* **245**, 165–186.

Rodi W. & Mackie R. L. 2001. Nonlinear conjugate gradient algorithm for 2-D magnetotelluric inversion. *Geophysics* **66**, 174–187.

Rossiter A. G. & Gray C. M. 2008. Barium contents of granites: key to understanding crustal architecture in the southern Lachlan Fold Belt. *Australian Journal of Earth Sciences* **55**, 433–448.

Seymour D. B., Green G. R. & Calver C. R. 2007. The geology and mineral deposits of Tasmania: a summary. *Mineral Resources Tasmania, Geological Survey Bulletin* **72**.

Simpson F. & Bahr K. 2005. *Practical magnetotellurics.* Cambridge University Press.

Smith R. J. & Frankcombe K. 2006. Role of geophysical methods applied to mapping mineral systems under the Murray Basin cover. *Australian Journal of Earth Sciences* **53**, 767–781.

Spies B. R. 1989. Depth of investigation in electromagnetic sounding methods. *Geophysics* **54**, 872–888.

Sternberg B. K., Washburne J. C. & Anderson R. G. 1985. Investigation of MT static shift correction methods. *55th Annual International Meeting*, SEG, Expanded Abstracts, 264–267.

Sternberg B. K., Washburne J. C. & Pellerin L. 1988. Correction for the static shift in magnetotellurics using transient electromagnetic soundings. *Geophysics* **53**, 1459–1468.

Tournerie B., Chouteau M. & Marcotte D. 2004. Estimation and removal of the MT static shift effect using geostatistical methods. 17th IAGA WG 1.2 Workshop on Electromagnetic Induction in the Earth, Hyderabad, India, October 2004.

Uchida T. 1993. Smooth 2-D inversion for magnetotelluric data based on statistical criterion ABCI. *Journal of Geomagnetism and Geoelectricity* **45**, 841–898.

VandenBerg A. H. M. 1999. Timing of orogenic events in the Lachlan Orogen. *Australian Journal of Earth Sciences* **46**, 691–701.

VandenBerg A. H. M., Willman C. E., Maher S., Simons B. A., Cayley R. A., Taylor D. H., Morand V. J., Moore D. H. & Radojkovic A. 2000. *The Tasman Fold Belt System in Victoria.* Geological Survey of Victoria, Special Publication.

Vozoff K. 1972. The magnetotelluric method in the exploration of sedimentary basins. *Geophysics* **37**, 98–141.

Vozoff K. 1991. The magnetotelluric method: Electromagnetic Methods in Applied Geophysics—Vol. 2 Applications. *Investigations in Geophysics No. 3.*

Willman C. E., Korsch R. J., Moore D. H., Cayley R. A., Lisitsin V. A., Rawling T. J., Morand V. J. & O'Shea P. J. (in press). Crustal-scale fluid pathways and source rocks in the Victorian gold province, Australia: Insights from deep seismic reflection profiles. *Economic Geology.*

Yi M-J., Kim J-H. & Chung S-H. 2003. Enhancing the resolving power of least-squares inversion with active constraint balancing. *Geophysics* **68**, 931–941.

Received 5 May 2010; accepted 3 August 2010

Appendix 2

Z. R. DENNIS, D. H. MOORE & J. P. CULL. 2011 MAGNETOTELLURIC SURVEY FOR UNDERCOVER STRUCTURAL MAPPING, CENTRAL VICTORIA.

Australian Journal of Earth Sciences, 58: 1, 33 — 47.

Australian Journal of Earth Sciences
Publication details, including instructions for authors and subscription information:
http://www.tandfonline.com/loi/taje20

A geological interpretation of the Echuca magnetotelluric survey, Victoria

Z. R. Dennis [a], D. H. Moore [a] & J. P. Cull [a]
[a] School of Geoscience, Monash University, Vic, 3800, Australia

Available online: 02 Aug 2011

To cite this article: Z. R. Dennis, D. H. Moore & J. P. Cull (2011): A geological interpretation of the Echuca magnetotelluric survey, Victoria, Australian Journal of Earth Sciences, 58:6, 587-597

To link to this article: http://dx.doi.org/10.1080/08120099.2011.577808

Australian Journal of Earth Sciences (2011) **58**, (587–597)

A geological interpretation of the Echuca magnetotelluric survey, Victoria

Z. R. DENNIS*, D. H. MOORE AND J. P. CULL

School of Geoscience, Monash University, Vic 3800, Australia

The northern part of the auriferous Bendigo Zone is obscured by thick Cenozoic sediments of the Murray Basin, and as such remains poorly explored. Consequently, in addition to the 2006 deep seismic line obtained in Central Victoria, a magnetotelluric (MT) survey was completed to provide a signature for the major structures previously defined in the Bendigo area. Based on these correlations a second MT line, which we present here, was completed some 50 km to the north of the original line in an attempt to trace the deep structural trends extending north to the Victorian/NSW border. Extending some 155 km across the central north of the state, data were collected at 52 sites along an east–west profile. The new electrical conductivity model generated correlates well with the results from the southern transect; it confirms previously identified structural trends to the north and identifies additional unknown deep structures, thus adding to the understanding of the geology of the covered region and to its gold-bearing potential.

KEY WORDS: magnetotellurics, resistivity, conductivity, gold undercover, Lachlan Fold Belt, Murray Basin cover.

INTRODUCTION

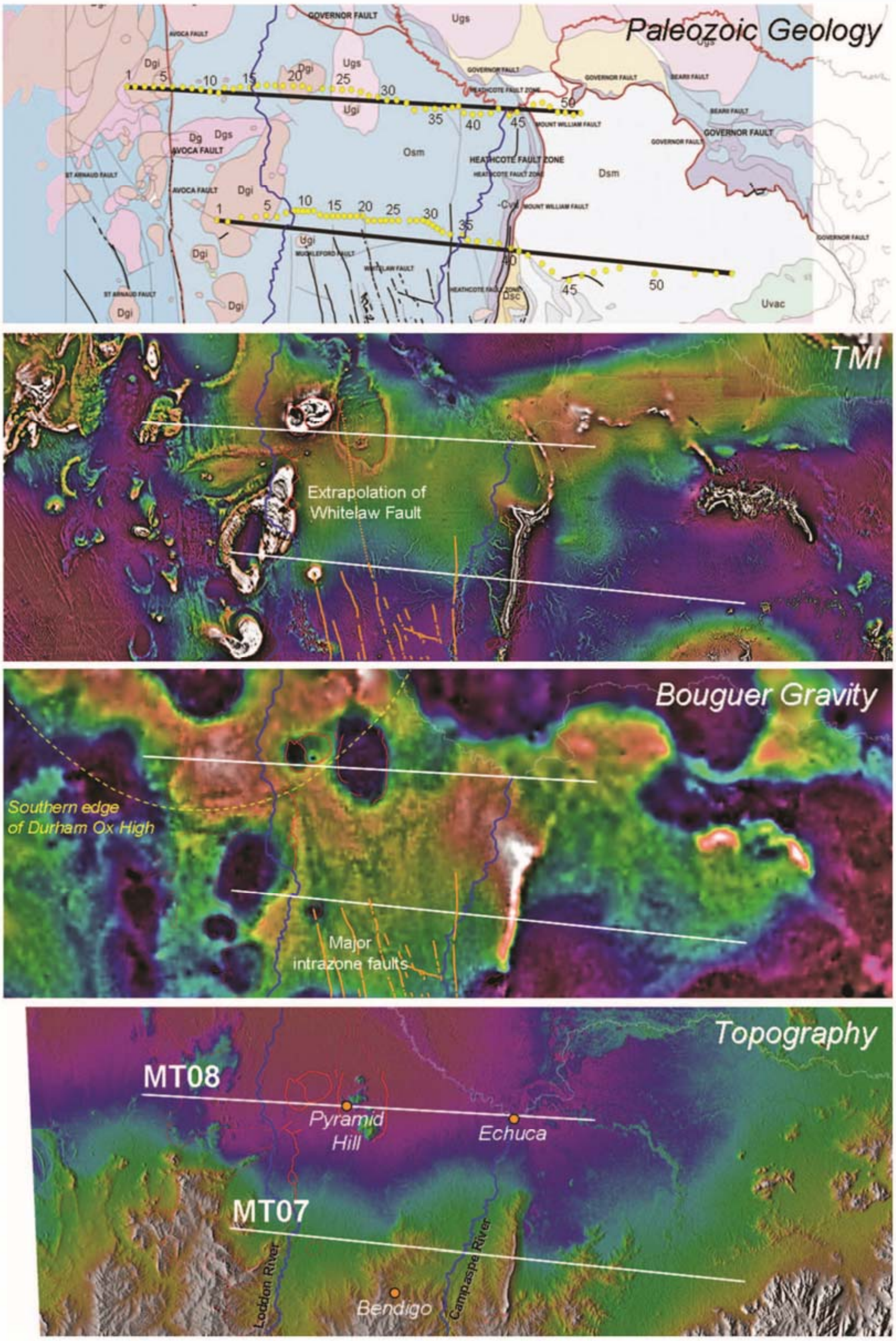
Magnetotelluric (MT) surveys (MT08) have been conducted in an area of northern Victoria (Figure 1), dominated by the Cenozoic Murray Basin cover sequence, which obscures a potentially prospective basement (Moore 2005). The Paleozoic geology of the northern region has been interpreted by Moore (2004, 2005, 2006) from airborne magnetic and gravity data, constrained by several hundred boreholes and geological mapping. In contrast, a previous MT survey on the Bendigo seismic line (MT07; Dennis *et al.* 2011), was located approximately 50 km to the south where the basement is much closer to the surface, and several exposed areas of the western Lachlan Orogen can still be observed (VandenBerg *et al.* 2000). The available correlations indicate that geophysical signatures of the obscured geology in the Echuca survey region are very similar to those observed on the southern margins of the Murray Basin, and suggest a northern extrapolation of the Bendigo Zone as a whole (Lisitsin *et al.* 2007). The new MT data will provide further detail relating to these proposed trends.

The depth of cover in northern Victoria, has led to most of the previous geological exploration in the area to be concentrated on the Murray Basin sequence, as opposed to the Paleozoic basement beneath it. This examination has led to the discovery of several heavy mineral sand deposits; however there has been only limited gold exploration this far north. To date, only uncommercial discoveries have been made at Lockington and Tandarra and north of Stawell, with a small amount of gold exploration reported around Cohuna and Kerang (Moore 2005; Smith & Frankcombe 2006). From the early 1990s, new data and understanding have led to the gold deposits of Victoria to be viewed as a coherent province, rather than a series of individual mines (Phillips *et al.* 2003). This concept, coupled with the currently available geological and geophysical datasets, indicates that a similar metallogeny may also exist to the north (Lisitsin *et al.* 2010).

The current understanding of metallogenic zoning in the western Lachlan Orogen also highlights the likelihood of deposits being present in some of the obscured northern parts of the Bendigo and neighbouring zones. The lack of any currently known major gold deposits in the area is hence more likely to be the result of limited effective exploration as opposed to the absence of economic mineralisation (Lisitsin *et al.* 2007). Models suggesting that the geometry and distribution of the groups of goldfields are the surface expressions of fluid flows related to the major faults (Cox *et al.* 1991; Willman 2007; Willman *et al.* 2010) may encourage the use of MT methods as a tool for further exploration. Previous comparisons with deep seismic data (Dennis *et al.* 2011) also demonstrate the merits of the MT method for the mapping of major faults.

*Corresponding author: zara.dennis@monash.edu

ISSN 0812-0099 print/ISSN 1440-0952 online © 2011 Geological Society of Australia
DOI: 10.1080/08120099.2011.577808


Paleozoic Geology
GOVERNOR FAULT
AVOCA FAULT
ST ARNAUD FAULT
HEATHCOTE FAULT ZONE
MOUNT WILLIAM FAULT
MUCKLEFORD FAULT
WHITELAW FAULT
BEARII FAULT
Dgi
Dg
Dgs
Ugs
Ugi
Osm
Dsm
Dsc
Uvac
TMI
Extrapolation of Whitelaw Fault
Bouguer Gravity
Southern edge of Durham Ox High
Major intrazone faults
Topography
MT08
MT07
Pyramid Hill
Echuca
Bendigo
Loddon River
Campaspe River

MAGNETOTELLURIC DATA

The MT method used for exploration and interpretation in central Victoria has been described in detail by Cull *et al.* (2008) and Dennis *et al.* (2009). The new MT08 profile covers almost 20 km of the Stawell Zone to the west, extends through the Bendigo Zone and terminates around 30 km east of Echuca, in the Melbourne Zone. The zone boundary faults, Avoca and Mount William, are both intersected (approximately) at right-angles by the line, which also traverses several buried granites (both previously identified and suspected) and large regions of flat low-lying land, where the conductive near-surface sediments obscure the deeper structural architecture.

Five component MT data were collected at 52 sites with a nominal spacing of 3 km. Data were obtained using the MTU-5A system of Phoenix Geophysics Ltd set to record for an average of 15 h each day over a frequency range of ~317 to 0.00034 Hz. In addition, a GPS synchronised, remote reference site was operated continuously throughout the survey. Situated ~70 km north of the nearest sounding site (site no.1 on the line) and away from any major cultural noise sources, synchronous recordings were taken at the remote reference in order to overcome any local coherent noise contamination.

After the successful deployment of the MT/TEM methods along the MT07 line, in-loop TEM data were collected (using the terraTEM system manufactured by Monex Geoscope P/L) at all sites in order to assess the extent of, and subsequently correct, any static offsets that were present in the MT08 dataset. A 1-D Earth has been resolved at high-frequencies with reasonable consistency along most of the northern line—situated at the foot of the Pyramid Hill granite outcrop, site 25 was the only site observed to depart significantly from this, showing splitting of the TM and TE apparent resistivity curves at all frequencies. Sites 7, 24, 26, 38 and 42 also demonstrate a difference in shift factors (*s*); however, the disparity between the orthogonal modes was much smaller at ~0.4 *s* (compared with 1.9 *s* for site 25 near-surface). All 46 remaining sites are predominantly 1-D near the surface. Figure 2 shows apparent resistivity and phase curves for six sounding locations, selected as being representative of the major trends identified along the MT08 transect and their adjacent sites. Displayed in numerical order, site 1 is located at the western end of the line in the Stawell Zone, with site 50, in the Melbourne Zone and about 6 km from the eastern limit of the line. With the exception of site 25, a 1-D Earth is shown by the high-frequency portion of the curves; the TE and TM modes then typically diverge for frequencies greater than ~1 Hz. Also shown are some specific examples (sites 21 and 28) where the structure appears predominantly 1-D at all frequencies.

Pseudo-sections comparing the observed and modelled data are presented by Figure 3, demonstrating the data fit achieved by the inversion. Overall, a good fit has been achieved, with only two significant mismatches present where the 2-D inversion has not been able to model the complex response. These are: site 25 (Pyramid Hill), where the TM observed apparent resistivity shows an isolated strip of particularly high magnitude, and site 37, showing an out-of-quadrant phase response approximately 110 km from origin in the TE observed section. Unusual high phase responses (>90°) such as this are an indication of a possibly complex response from chargeable ground; Heise & Pous (2003) proposed that such a response comes as a consequence of strong macroanisotropic structure. This supports the inversion results as we would not expect a 2-D inversion routine to be able to resolve complex, multidimensional measurements such as these. As such, sites displaying a clear multidimensional response (25 and 37 as discussed and additionally site 16 where a corrupted file was suspected) were consequently excluded from the final inversion.

INDUCTION ARROW RESPONSE

Induction arrows show regional variations in the lateral resistivity distribution free of any complications related to static shifts (Thiel & Heinson 2010). Figure 4, following the Parkinson convention, shows arrows at various frequencies that have been derived for the Echuca transect. The line crosses the Avoca and Mount William Faults at sites 6 and 46 respectively, at approximately 730 and 845 km E in Figure 4. Separating the Bendigo and Melbourne Zones, the Mount William Fault (E boundary of the HFZ) appeared in the MT07 transect as a central pivot, about which the conductivity gradient shifts from north to south with a focus towards the fault (E–W alignment) at ~0.022 Hz (see Figure 4; Dennis *et al.* 2011) consistent with an off-line conductive body to the north. This trend can clearly be seen to repeat in Figure 4 for the northern transect but with a focus occurring at ~0.0092 Hz, below which the coast effect (Lilley 1976) becomes progressively more dominant.

As stated by Spies (1989), the depth at which the MT method is most sensitive for a specified frequency is

Figure 1 Survey lines indicated on geological, topographic, gravity and magnetic (TMI) maps of north-central Victoria. Both the survey lines and the individual MT sites (plotted by KIGAM) have been indicated on the geology map, with lines only shown on all others. Whites show regions of gravity/magnetic/topographic highs, whereas blues and violets show lows. Rock types on geological underlay are: Cvu, Cambrian volcanics (HFZ); Dg, Devonian granites; Dgi, Devonian I-type granites; Dgs, Devonian S-type granites; Dsc, Devonian sediments (continental sandstone); Dsm, Devonian sediments (marine); Osm, Ordovician sediments (marine); Ug, Upper Devonian granites; Ugi, Upper Devonian I-type granites; Ugs, Upper Devonian S-type granites; Uvac, Upper Devonian volcanics. Major granites have also been outlined in red. Other features indicated are (in orange) the locations of the major faults to the south of the region including the northern extrapolation of the Whitelaw Fault, major rivers (in blue) running N–S which are crossed by the survey lines and (in yellow) the southern limit of the Durham Ox High.

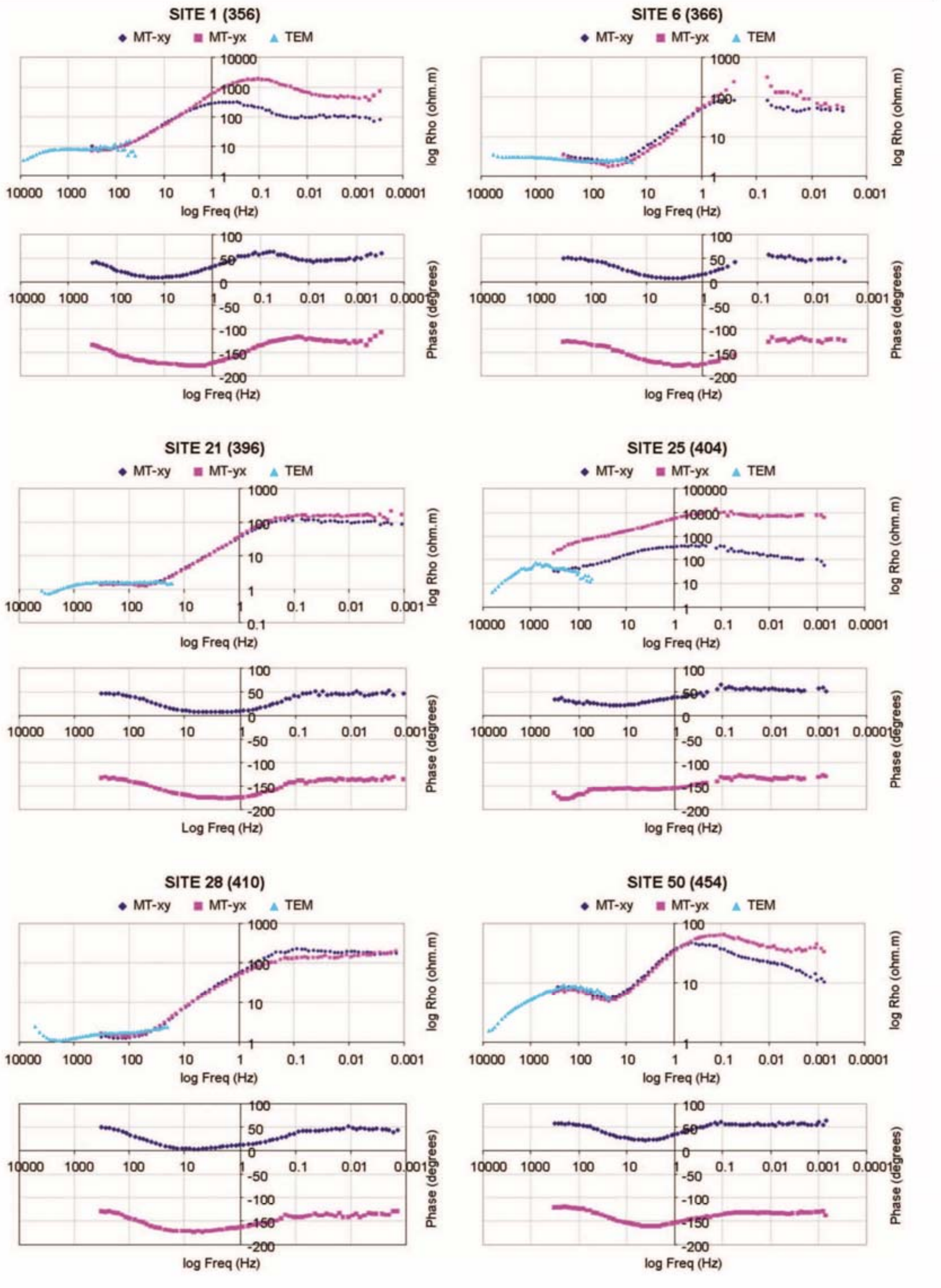


Figure 2 Apparent resistivity and phase curves for representative sites along the MT08 line. The selected sites show the 1-D near surface character that is common across the majority of the transect with the primary exception of site 25 showing splitting of the orthogonal modes at all periods. Other sites presented show examples of approximate 1-D structure across the entire frequency range (21 and 28), an edited site where poor-quality, outlying data points have been removed (6, Avoca Fault site), and sites from either end of the line residing in both the Stawell and Melbourne zones; sites 1 and 50 respectively.

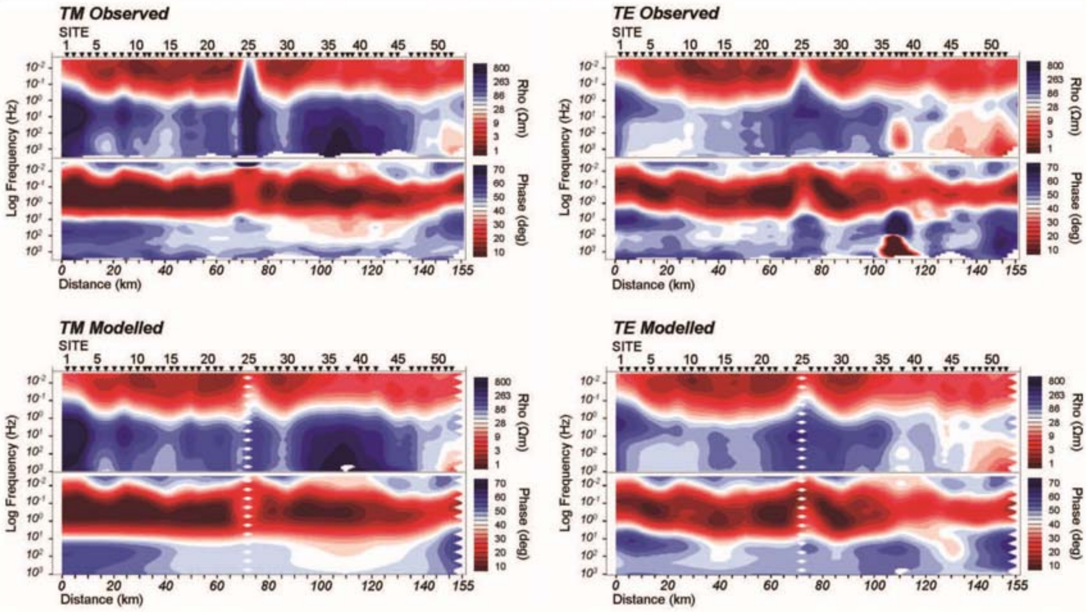


Figure 3 Comparison of the observed and modelled apparent resistivity and phase pseudo-sections for TM (yx) and TE (xy) mode data. Overall, a good data fit is seen. The origin of the horizontal distance axis is situated ~2 km west of site 1 with all sections scaled equally for ease of comparison. Note that high phase angles typically were not recorded, and so the range >70° has not been shown.

determined by calculation of the skin-depth $\sim 0.5\sqrt{(\rho T)}$; defined as the depth at which the signal has fallen to e^{-1} of its original amplitude, where ρ is the half-space resistivity in Ωm, and T is the signal period in seconds. In order to give some appreciation of the depths at which this east–west alignment occurs, therefore, equivalent skin-depths for these frequencies have been calculated (Table 1). A nominal resistivity of ~100 Ωm has been estimated for the entire crust. This contrasts with responses from the Shepparton Formation, a Late Neogene nonmarine clay-rich unit in the Murray Basin which demonstrates values typically <10 Ωm, illustrated by the high-frequency responses shown in Figure 2.

Subsequently, in order to highlight the governance of resistivity on sounding depth, half-space resistivities of 1, 10, 100 and 1000 Ωm were compared. The observed variations in apparent resistivity are consistent with a northerly dip in the transition point of the induction arrows from north to south. With a crustal thickness of approximately 36 km reported under central Victoria (Rawlinson & Kennett 2008, figure 12) and with reference to the values calculated for ~100 Ωm net response, this focusing of the induction arrows persists to upper mantle depths. It thus provides an indication of the deep zone boundary separating the Bendigo and Melbourne Zones and so demonstrates a linkage between crustal and mantle compositions.

Above 1.72 Hz the high frequency response shows no clear focusing of the arrows and magnitudes (length of the arrows) are small. This is quite different to the Bendigo response at these frequencies, where large magnitudes and erratic orientations were resolved for these surficial depths. For the frequencies of 9.40 and 4.10 Hz, therefore, the low magnitude and unfocused response of the induction arrows indicates a predominantly homogeneous layer, consistent with the 1-D response of the known near surface geology, where thick early Paleozoic oceanic sedimentary rocks are in abundance.

INVERSION MODEL

The preliminary analysis of this dataset (presented in Dennis *et al.* 2009) was conducted using two independent algorithms, generating two comparative models and thus providing increased confidence in the modelled results. With these as an initial guide, the inversion model presented here (Figure 5) was produced using the WinGLink® software package for MT data (distributed by Geosystem SRL). With a starting half-space resistivity of 100 Ωm, including a surficial layer (~300 m thick) of 10 Ωm to represent the Murray Basin overburden, a smooth inversion routine was implemented. The algorithm finds regularised solutions (Tikhonov Regularisation) to the 2-D inverse problem for MT data using the method of non-linear conjugate gradients (Geosystem SRL 2008), comprehensive details of which are available in Rodi & Mackie (2001). Topography has not been included in the inversion, as the area can be considered flat with respect to the length of the survey line (see Figure 1).

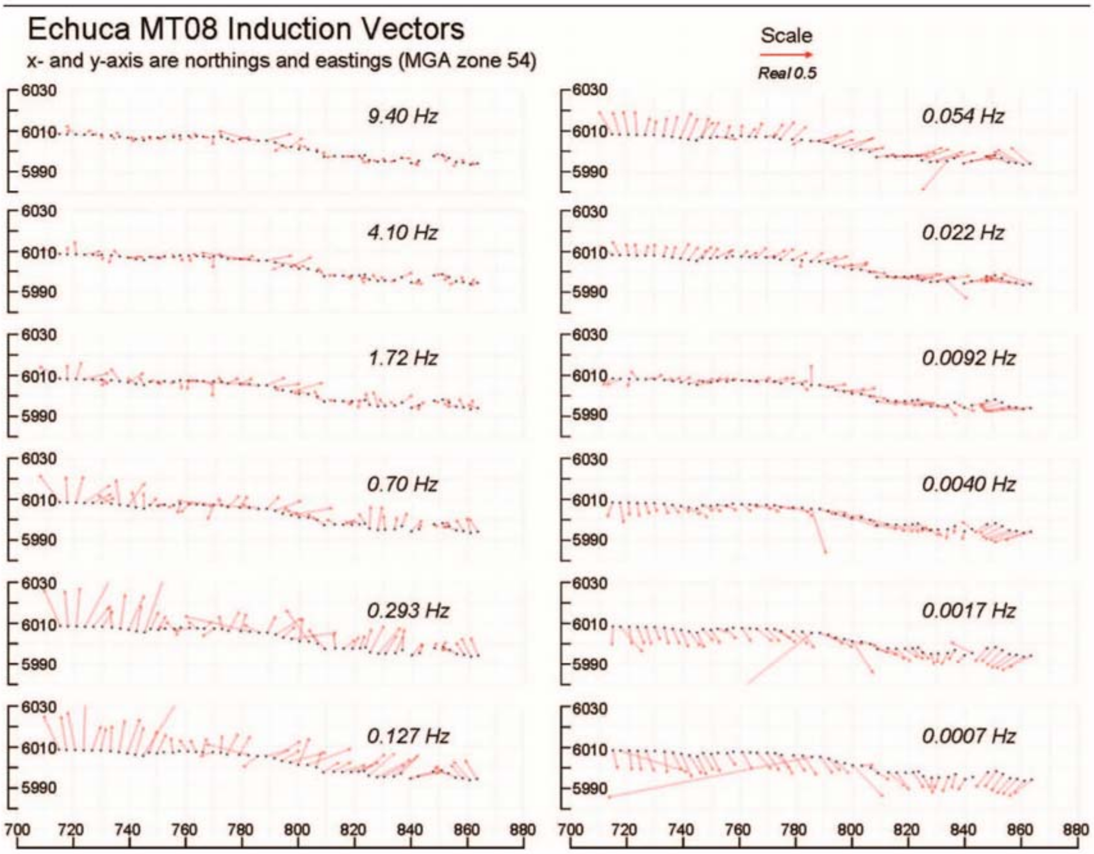


Figure 4 Induction arrows for the MT08 survey line at various frequencies. Following the Parkinson convention, the induction arrows point towards regions of high conductivity. Longer periods show character at greater depth. The major faults defining the zone boundaries are located at approximately 730 and 845 km E, for the Avoca and Mount William Faults respectively.

Table 1 Skin-depth approximations for various half-space resistivities, calculated for the frequencies at which the east–west focus of the induction arrow response occurs.

Frequency (Hz)	Period (s)	Skin depth (km)			
		1 Ωm	10 Ωm	100 Ωm	1000 Ωm
0.022	45.45	3.37	10.66	33.7	106.6
0.0092	108.7	5.21	16.48	52.13	164.85

Skin-depth $\sim 0.5\sqrt{(\rho T)}$, where half-space resistivities of 1, 10, 100 and 1000 Ωm have been used in order to highlight the influence of resistivity on the sounding depth.

Annotated 2-D section

Figure 5 presents the final 2-D inversion model of the Echuca line, along with annotations indicating the geological structures that can be identified from the electrical response. The features identified in the Bendigo profile have been used as a guide, correlating similar responses with known features previously identified in the seismic interpretation (ANSIR 2008). Probable extensions to the faulted regions (in particular the Whitelaw Fault and the deep section of the northern most part of the Mount William Fault) are indicated by dashed lines.

To summarise the electric response in this region, the section can be broadly divided into a series of layers. The thin (~0.5 km) uppermost surficial layer is one of high conductivity, with typical resistivities of less than 10 Ωm resolved by the inversion model correlates with a nonmarine clay-rich unit in the Murray Basin. Below this, extending to around 15 km in depth (approximate boundary indicated in Figure 5), large resistive units of >1000 Ωm dominate the section, predominantly appearing as rounded isolated bodies; the host resistivity is lower, nominally in the range 100–200 Ωm. The lower crustal structure is a much more conductive region, approximately 10 Ωm in resistivity. Slab-like units descend into the mantle, where a much more homogeneous character (~100 Ωm) can be seen.

In order to test its robustness, Figure 6 shows the sensitivity map for the inversion model; obtained from the diagonal part of $A^T R^{-1}_{dd} A$, where A is the sensitivity matrix, R^{-1}_{dd} is the inverse of the data covariance matrix, and T denotes the transpose matrix (Geosystem SRL 2008; Thiel & Heinson 2010). This

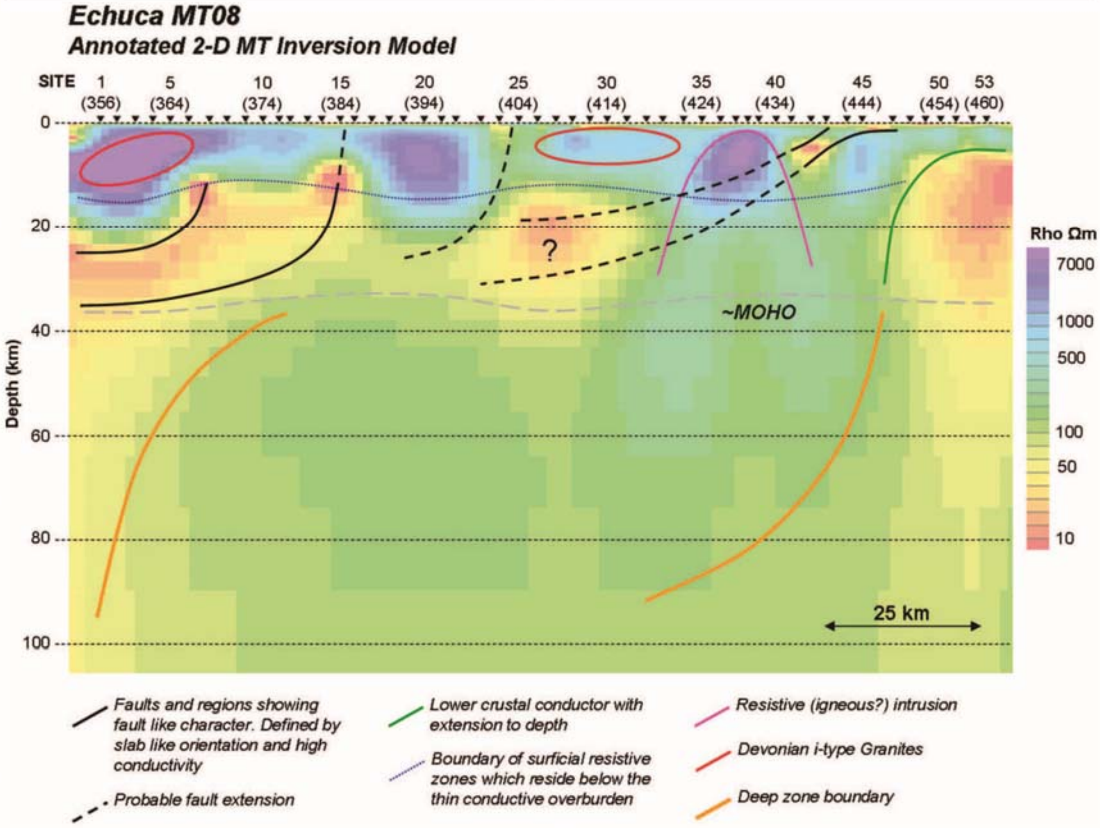


Figure 5 Annotated 2-D MT inversion model for the Echuca line. MT model (statics corrections applied) with structural features derived from the electrical character also marked. Thick dashed lines show the likely extension of the faults and/or their location prior to the resistive intrusions which intersect the original conductive wedge. The Avoca and Mount William Faults are located at sites 6 and 46 respectively with site 24 situated at the foot (~500 m W) of the Pyramid Hill Outcrop—the proposed extension of the Whitelaw Fault corresponding with this location also shown. The Loddon River crosses the line at approximately site 15. The boundaries shown near the western edge of the section are beyond the limits of the profile and so have limited validity, with the question mark also indicating a degree of uncertainty in the resolved conductor. The site number given in brackets is the site number as used in the preliminary KIGAM inversion and as such corresponds with data presented in Dennis *et al.* (2009).

information is used to generate an image demonstrating how strongly changes in the resistivity of each cell will affect the overall misfit of the data. Low sensitivities indicate regions where a change in the cell value will have a minimal effect on the data fit, whereas the zones of greatest sensitivity cannot be perturbed without causing a substantial change in the misfit. A comparison of Figures 5 and 6 therefore shows that zones of high conductivity are also high in sensitivity (resistive zones correspondingly are low in sensitivity).

STRUCTURAL RESPONSES

Fault zones

Deep crustal faults are considered to provide the pathways for the significant mineralisation and magmatic processes in the region (Willman *et al.* 2010), making major fault identification an important outcome from these new MT results. Aside from the regional-scale faults which define the zone boundaries, the surface locations of second-order faults that were identified to the south become less apparent to the north. As observed from the 2-D cross-section (Figure 5), the Avoca Fault does not show the expected highly conductive response near surface, due to the overprinting of the Loddon Vale Pluton (Figure 1; Dgi coincident with sites 1 to 6) where the fault and survey line intersect. However, with a minimum depth of ~10 km and located beneath site 6 (identified as the surface location of the Avoca Fault from regional maps, e.g. Paleozoic Geology, Figure 1) we can see a distinct west dipping conductor, which has been cut by the resistive Devonian granites. With an adjoining wedge peaking below site 15 (~10 Ωm at 10 km), a projection of this conductive zone meets the surface where the Loddon River crosses the traverse (Figure 1). This is similar to the way in which the forward modelling to compare the new RMS misfits. An initial RMS of 4.79 was determined for the entire section, with specific sites where complex responses have been inferred increasing the net section response. Most sites ranged from 1.5 to 2.0, but a few were outside this. In particular, site 25 gave an RMS of 25.25 and site 36 an RMS of 16.9; removal of these two sites alone achieved a significantly improved section RMS of 4.15. Sites 11, 26 and 53 were then scrutinised further, selected specifically due to their central location above the three main conductive units. With initial RMS misfits of 1.41, 1.70 and 1.24 respectively, replacing the major conductive zone with a block of 100 Ωm (representative of the host rocks) increased the misfits to 2.45, 1.78 and an unacceptable 7.91. The conductive units subsequently reappeared with RMS values similar to the initial misfits after inverting the forward model. Consequently, these units are in general considered to be well resolved. However, the small variation in misfit obtained for the conductor below site 26, along with the lower sensitivities observed in Figure 6, suggests that this mid to lower crustal conductor (indicated by the question mark in Figure 5), is not as well defined as the two other major units, as it allows for a wider range of resistivities to be resolved by the inversion.

OTHER OBSERVATIONS

The Durham Ox High (VandenBerg *et al.* 2000, p. 209; Moore 2004, figure 12) is a region of high gravity and magnetic responses, near the western end of the line, the south-eastern extent of which has been marked on the gravity map in Figure 1. It cuts across the surface geology in such a way that indicates its source is a deep feature, with Moore (2004) reporting a probable depth of ~15 km and the southern edge being the locus of many shallow earthquakes in the overlying granites. VandenBerg *et al.* (2000) proposed two possible explanations for the area of higher magnetic and gravity responses; either a large mafic intrusion or a microplate (Moore 2004). Moore (2004) also suggested a third possibility, that the high is the result of a region of duplexed dense Cambrian ocean crust, a suggestion fitting with many empirical observations.

With the high covering the western section of the line, up to and including the Pyramid Hill granites, much of the MT line could resolve the structure relating to the anomaly. In the MT cross-section both a strong resistive layer and a conductive one beneath this are present in the crust. Mafic rocks are generally resistive; we could assume therefore that the overlaying resistive zone of >1000 Ωm thus reflects the structure. However, analysis of the Heathcote stratigraphy by Edwards *et al.* (1998) has shown mafic rock units to be interlayered with shales and cherts that contain highly conductive minerals such as pyrite and graphite (Edwards *et al.* 1998, figure 9; VandenBerg *et al.* 2000, figure 2.2). These would result in the conductivity structure as a whole being significantly elevated. Therefore, with a similar electrical response to that of the Heathcote Fault Zone, the dipping wedge-like conductors we see below the resistive layer (peaking at sites 6 and 15) may reflect the Durham Ox High; a region of high-density mafic volcanic rocks, with interlayers of deep oceanic sediments and their associated conductive minerals. Figure 1 shows a high gravity response at the corresponding location.

Several large granite intrusions are located around the southern edge of the Durham Ox High (outlined in red on Figure 1). Characteristically they have low gravity responses and high resistivities. The shallow seismic activity in this area may however have resulted in additional fracturing of the brittle granites, increasing their permeabilities and consequently resulting in lower-than-average resistivity signatures. This fracturing may be the reason for the Devonian granite centred at site 30 not appearing as high in resistivity as we might otherwise expect from similar rock units.

The Yando Swamp, its surrounds (sites 13–23) and similar low lying wet regions between sites 33 and 40 are all shown by the MT results to be situated above high resistivity and consequently low porosity underlying bedrocks. The TEM response (unpublished data collected by the primary author) across these zones also shows significantly elevated conductivity in the top 10–20 m, a response that indicates surficial water saturation. Radiometric data in such water-logged areas generate a low response as the signal is inhibited by the saturation; a response such as this is seen above the Wedderburn Granite (western end of the Bendigo profile), where radiometric lows away from the outcrops correlate with swampy regions (VandenBerg *et al.* 2000; figure 4.10, page 341). This association is consistent with the buried granite between sites 33 and 41 proposed by Moore (2006).

Alternative interpretation of 2-D cross-section

The lower crustal conductor underneath site 26 has been interpreted in Figure 5 as a possible west-dipping extension of the faults in the Heathcote Fault Zone. An alternative interpretation could connect this conductive body with the northern extension of the Whitelaw Fault. This would imply a steeply east-dipping orientation for the fault, which would possibly cross the MOHO and displace the upper mantle. This interpretation seems less likely, since it does not correspond with the results of the 2006 seismic line (Willman *et al.* 2010; Cayley *et al.* 2011), which showed the Whitelaw Fault as steeply west-dipping in the upper crust and flattening to about 15° below about 12 km. Indeed, where the seismic line crosses the eastern Stawell Zone, the Melbourne Zone and the western Melbourne Zone, the seismic data interpretation is dominated by east-dipping structures. The only west-dipping structures are interpreted to be back-thrusts.

DISCUSSION

The current results confirm that MT methods can be used as a tool to identify faults and other significant structures at depth. The results agree with interpretations based on the reflection seismic method, and consequently MT sections can be reliably extrapolated into areas along strike from the seismic data. Apart from the economic advantages it is clear that parallel MT surveys off-line from previously obtained seismic traverse may enable an interpretation to be more confidently extrapolated into other less well-controlled areas.

In the present central Victorian example, where gold exploration has been hampered by the extensive Murray Basin cover, MT offers a comparatively cheap, reliable way of determining the locations and dips of the major fault zones. Since it is likely that the most highly mineralised zones occur at the brittle–ductile transition in the hangingwalls of the major intrazone faults (e.g. Willman 2007; Willman *et al.* 2010), a method to identify the major faults and any associated brittle–ductile transitions is a key step in determining new targets for gold exploration. Willman *et al.* (2010) also suggested that the points of inflexion of major faults may be important in determining the locations of gold deposits, since they determine the places where the faults become impermeable to fluids that were rising along them. If so, MT surveys may be able to better locate these positions, assisting with the selection of more specific exploration targets and potential regions for mineralisation. Although the 3 km station spacing of the MT08 survey is inadequate for target selection, it is possible to identify some areas where the faults steepen and become less permeable for the gold-rich fluids. Closer station spacing would improve this capacity further.

Since the maximum permeability zone of the gold-bearing fluid system was a three-dimensional system, defining the most favourable locations for exploration is ultimately also a three-dimensional problem. Since acquiring MT data is significantly less expensive than seismic data, the MT method offers a correspondingly greater capacity to define the most favourable zones in three dimensions. Closely spaced MT lines may not only highlight fault systems of particular interest but also potentially highlight dilational offsets within them, further narrowing a target area (Cox 2005).

Bierlein *et al.* (1998) showed that, at the Lake Cooper Quarry in the Heathcote Fault Zone, Au averaged 0.07 g/t in the sulfide-rich (locally $>60\%$ pyrite) fine grained sedimentary rocks between the tholeiitic basalt layers. Seven sulfide-rich layers, each about 0.5 m thick, were logged in a stratigraphic thickness of 280 m. Since the adjacent basalts and turbidites were sulfide-poor, it is possible that sedimentary rocks such as these might be capable of giving an MT response at a considerable depth below the present ground surface.

Lower resistivities obtained near the Durham Ox High may also be important for gold exploration. Willman *et al.* (2010) considered that the duplexed mafic crust and the interlayered metasedimentary rocks at depth may have been the source of the gold in the Bendigo and Stawell Zones. If the Durham Ox High reflects a thicker section of duplexed mafic and metasedimentary rocks, then it would provide a larger body of potential source rocks with which to form a gold deposit. The lower resistivity, perhaps implying a higher pyrite content, adds encouragement to this possibility.

The calculated depth to the top of the body is about 15 km (Edwards *et al.* 2001). Assuming similar fault shapes to that shown in Willman *et al.* (2010) would place the top close to the region where any deep flat faults start to emerge at greater angles as they shallow. This suggests that any potentially more favourable zones would be either above or a few kilometres east of the edge of the Durham Ox High. This distance would vary depending on the actual position of the change of dip and the location of the brittle–ductile transition in the hanging-wall of the fault.

CONCLUSIONS

The new MT08 (Echuca) dataset presented here supports the structural trends found along the southern MT07 (Bendigo) line (Dennis *et al.* 2011) and shows that the MT07 results previously obtained can be extrapolated north. Following the application of high-frequency statics corrections, a 2-D inversion of the data was completed. The resistivity models show a highly electrically conductive surficial layer of the Murray Basin. The lower part of the Murray Basin and the underlying Paleozoic crust have a net resistivity of $\sim 100\ \Omega$m; the Paleozoic rocks are cut by west-dipping conductive faults of north–south strike and have been intruded by highly resistive granitic bodies. By comparison the conductivity of the upper mantle is more uniform, but the deep zone boundary can still be inferred from the weak contrast between the Bendigo and Melbourne Zones.

The current MT results have shown that the major faults in the Bendigo Zone are readily identifiable by their higher conductivities and can be confidently projected to provide correlations between the parallel transects. Consequently, several high-priority areas can be identified for future gold exploration. In particular, the extension of the Whitelaw Fault appears highly significant as originally proposed by Moore (2006). Future surveys with higher-density sounding sites giving a better three-dimensional picture may also have the capacity to highlight not just favourable linear zones but higher-priority areas within these zones. In addition, the Durham Ox High is identified as an area with a thicker sequence of source rocks potentially related to gold deposits. In these circumstances, MT methods may provide critical structural constraints as well as a direct target detection method for gold deposits under cover.

ACKNOWLEDGEMENTS

MT data were collected as part of a collaborative project between Monash University, GeoScience Victoria (GSV) as part of their Gold Undercover Program, the Korean Institute of Geoscience and Mineral Resources (KIGAM) and the Geological Survey of Japan (GSJ), including research funding from KIGAM and GSJ. Dr Stephan Thiel at the University of Adelaide provided assistance with the modelling software. Comments and suggestions received from the reviewers, F. E. M. (Ted) Lilley and one other have been of considerable benefit and have greatly improved the manuscript.

REFERENCES

ANSIR 2008. L178 Central Victoria Seismic Survey Results, 2006. Migrated & interpreted migrated sections, and supporting data: National Research Facility for Earth Sounding, GeoScience Australia—*Data CD*.

BIERLEIN F. P., ARNE D. C., BROOME J. M. N. & RAMSAY W. R. H. 1998. Metatholeiites and interflow sediments from the Cambrian Heathcote greenstone belt, Australia; sources for gold mineralisation in Victoria? *Economic Geology* **93**, 84–101.

CAYLEY R. A., KORSCH R. J., MOORE D. H., COSTELLOE R. D., NAKAMURA A., WILLMAN C. E., RAWLING T. J., MORAND V. J., SKLADZIEN P. B. & O'SHEA P. J. 2011. Crustal architecture of Central Victoria: results from the 2006 deep crustal reflection seismic survey. *Australian Journal of Earth Sciences* **58**, 113–156.

COX S. F., ETHERIDGE M. A., CAS R. A. F. & CLIFFORD B. A. 1991. Deformational style of the Castlemaine area, Bendigo-Ballarat Zone: Implications for evolution of crustal structure in central Victoria. *Australian Journal of Earth Sciences* **38**, 151–170.

COX S. F. 2005. Coupling between deformation, fluid pressures, and fluid flow in ore-producing hydrothermal systems at depth in the crust. *In:* Hedenquist J. W., Thompson J. F. H., Goldfarb R. J. & Richards J. R. (eds) *Economic Geology 100th Anniversary Volume*, 30–75.

CULL J., LEE S. K., LEE T. J., UCHIDA T., PARK I. W. & SONG Y. 2008. A magnetotelluric survey along part of the 2006 central Victorian seismic transect. *GeoScience Victoria Gold Undercover Report 8*—Department of Primary Industries.

DENNIS Z. R., CULL J.P., LEE S. K., LEE T. J., PARK I. W., SONG Y. & UCHIDA T. 2009. A magnetotelluric survey to the north of the 2006 central Victorian seismic transect. Final report: *GeoScience Victoria Gold Undercover Report 18*—Department of Primary Industries.

DENNIS Z. R., MOORE D. H. & CULL J. P. 2011. Magnetotelluric survey for structural mapping in regions affected by static shift, central Victoria. *Australian Journal of Earth Sciences* **58**, 1–15.

EDWARDS J., MOORE D. H., BIBBY L. M. & LYNN S. F. 2001. Wedderburn 1:100 000 map area geological report. *Geological Survey of Victoria Report* **122**.

EDWARDS J., WOHLT K. E., SLATER K. R., OLSHINA A. & HUTCHINSON D. F. 1998. Heathcote and parts of Woodend and Echuca 1:100 000 map area geological report. *Geological Survey of Victoria Report* **108**.

GEOSYSTEM SRL 2008. *Chapter 18—MT 2D inversions: WinGLink® user's guide.*

HEISE W. & POUS J. 2003. Anomalous phases exceeding 90° in magnetotellurics: anisotropic model studies and a field example. *Geophysics Journal International* **155**, 308–318.

LILLEY F. E. M. 1976. A magnetometer array study across southern Victoria and the Bass Straight area, Australia. *Geophysical Journal Royal Astronomical Society* **46**, 165–184.

LISITSIN V. A., MOORE D. H., OLSHINA A. & WILLMAN C. E. 2010. Undiscovered orogenic gold endowment in Northern Victoria, Australia. *Ore Geology Reviews* **38**, 251–269.

LISITSIN V., OLSHINA A., MOORE D. H. & WILLMAN C. E. 2007. Assessment of undiscovered mesozonal orogenic gold endowment under cover in the northern part of the Bendigo Zone. *Geoscience Victoria Gold Undercover Report 2—Department of Primary Industries.*

MOORE D. H. 2004. St Arnaud 1:250 000 area: A geological interpretation of the geophysical data. *Victorian Initiative for Minerals and Petroleum Report* **82**. Department of Primary Industries, Victoria.

MOORE D. H. 2005. Swan Hill 1:250 000 and parts of Balranald and Deniliquin 1:250 000 map areas: A geological interpretation of the geophysical data. *Victorian Initiative for Minerals and Petroleum Report* **84**. Department of Primary Industries, Victoria.

MOORE D. H. 2006. Bendigo 1:250 000 and part of Deniliquin 1:250 00 map areas: A geological interpretation of the geophysical data. *Victorian Initiative for Minerals and Petroleum Report.* 86. Department of Primary Industries, Victoria.

PHILLIPS G. N., HUGHES M. J., ARNE D. C., BIERLEIN F. P., CAREY S. P., JACKSON T. & WILLMAN C. E. 2003. Gold—Historical wealth, future potential. *In:* Birch W. D. ed. *Geology of Victoria.* Geological Society of Australia Special Publication 23. Geological Society of Australia (Victoria Division), 377–433.

RAWLINSON N. & KENNETT B. L. N. 2008. Teleseismic tomography of the upper mantle beneath the southern Lachlan orogen, Australia. *Physics of the Earth and Planetary Interiors* **167**, 84–97.

RODI W. & MACKIE R. L. 2001. Nonlinear conjugate gradient algorithm for 2-D magnetotelluric inversion. *Geophysics* **66**, 174–187.

SMITH R. J. & FRANKCOMBE K. 2006. Role of geophysical methods applied to mapping mineral systems under the Murray Basin cover. *Australian Journal of Earth Sciences* **53**, 767–781.

SPIES B. R. 1989. Depth of investigation in electromagnetic sounding methods. *Geophysics* **54**, 872–888.

THIEL S. & HEINSON G. 2010. Crustal imaging of a mobile belt using magnetotellurics: An example of the Fowler Domain in South Australia. *Journal of Geophysical Research* **115**, B06102.

VANDENBERG A. H. M., WILLMAN C. E., MAHER S., SIMONS B. A., CAYLEY R. A., TAYLOR D. H., MORAND V. J., MOORE D. H. & RADOJKOVIC A. 2000. The Tasman Fold Belt System in Victoria. *Geological Survey of Victoria Special Publication.*

WILLMAN C. E. 2007. Regional structural controls of gold mineralisation. Bendigo and Castlemaine goldfields, Central Victoria, Australia. *Mineralium Deposita* **42**, 449–463.

WILLMAN C. E., KORSCH R. J., MOORE D. H., CAYLEY R. A., LISITSIN V. A., RAWLING T. J., MORAND V. J. & O'SHEA P. J. 2010. Crustal-scale fluid pathways and source rocks in the Victorian gold province, Australia: Insights from deep seismic reflection profiles. *Economic Geology* **105**, 895–915.

Received 12 June 2010; accepted 29 March 2011

Appendix 3

D. H. MOORE 2016. COMMENT ON “EARLY OPENING OF AUSTRALIA AND ANTARCTICA: NEW INFERENCES AND REGIONAL CONSEQUENCES” BY JENSEN JACOB AND JÉRÔME DYMENT

Tectonophysics, 668 — 669, 105 — 107

Tectonophysics 668–669 (2016) 105–107

Contents lists available at ScienceDirect

Tectonophysics

journal homepage: www.elsevier.com/locate/tecto

Comment on "Early Opening of Australia and Antarctica: New inferences and regional consequences" by Jensen Jacob & Jérôme Dyment

David H. Moore*

School of Earth, Atmosphere & Environment, PO Box 28E, Monash University, Clayton, Vic 3800, Australia

ARTICLE INFO

Article history:
Received 12 November 2014
Received in revised form 22 January 2015
Accepted 26 January 2015
Available online 26 November 2015

Keywords:
Australia–Antarctica
Tasmania
Selwyn Block
Gondwana
Plate reconstruction

The article by Jacob and Dyment (2014) offers a new reconstruction of the Australian–Antarctic section of Gondwana based on matching the satellite-derived free air gravity responses of the two continental margins. They note the close match between Australia and Antarctica and propose that this match can be improved if Tasmania was located approximately 300 km northwest of its present position at 130 Ma, with the subsequent movement accommodated by rotation of their 'Tasmanian Block' of 7.936° around a pole at 54.9°S 110.0°E. They suggest that Tasmania arrived at its present position with respect to mainland Australia at 83 Ma.

I draw the authors' attention to the papers by Cayley et al. (2002), Cayley et al. (2011) and Moore et al. (2013), which evaluated the various hypotheses that Tasmania had moved with respect to mainland Australia. All concluded that the pre-Ordovician rocks of western Tasmania extend north into central Victoria and underlie the Ordovician to Middle Devonian Melbourne Zone, forming the Selwyn Block (Fig. 1). The Selwyn Block came to its present position with respect to the rest of the Western Lachlan Orogen in the Middle Devonian and extends almost 300 km north from the Victorian coast. This hypothesis has been successfully tested against the outcropping geology, and potential field and deep seismic interpretations and is further supported by a recent ambient seismic noise tomographic study (Pilia et al., 2015).

[redacted]

For example, at A in Fig. 1, on the east coast of King Island, Meffre et al. (2004) described rift tholeiites with a Nd–Sm isochron age of 579 ± 16 Ma, coeval with a U–Pb zircon age of 575 ± 3 Ma by Calver et al. (2004) on the same suite of rocks. These rocks have magnetic susceptibilities of up to 69.5×10^{-3} SI. Following the magnetic high north, Henry and Birch (1992) found geochemically similar rocks at present at B on the southern Victorian coast; these have susceptibilities of up to 91×10^{-3} SI and the magnetic continuity between the two supports this correlation. The high continues north, until approximately 30 km north of the coast it comes from rocks that lie under nonmagnetic Pridoli sedimentary rocks of the Melbourne Zone. To the south of A, the highly magnetic responses continue just offshore along the entire west coast of Tasmania (Moore et al., 2014).

Thus the model shown in Fig. 3 of Jacob and Dyment (2014) would require that not only Tasmania to have moved approximately 300 km but that the whole Melbourne Zone and Selwyn Block must also have moved a similar distance. Their model contradicts what is known about the geology of western and central Victoria, it would require that the Paleozoic and Mesozoic geology of southeastern Australia be rewritten, and it would cause major issues in interpreting other seismic and potential field data sets. In fact, what little movement there has been between Victoria and Tasmania across Bass Strait has largely been accommodated by localised Late Cretaceous and Cenozoic rifting (e.g. Briguglio et al., 2013). Were the authors unaware of the Selwyn Block model or did they choose to ignore it?

A similar comment could be made about potential correlations between Australian and Antarctic geology. There are abundant alternatives in the literature, including those by Stump et al. (1986), Münker and Crawford (2000), Bradshaw (2007), Whittaker et al. (2007), Payne et al. (2009), Goodge and Fanning (2010), Boger (2011) and the subsequent comment by Direen (2012), Veevers (2012), Williams et al. (2012), White et al. (2013) and Aitken et al. (2014). Each of these examines not only the shapes of the continent outlines, but also attempts to match aspects of the geology of the two continents. A coherent geological match is a fundamental prerequisite for any valid fit and yet Jacob and Dyment (2014) offer no such a match. I invite them to provide an internally and externally consistent basement geological map that shows their pre-breakup correlations between Antarctica and Australia, including Tasmania and the Selwyn Block.

As well, the geology as known from dredge samples from the South Tasman Rise strongly suggests that the pre-Gondwana breakup position of the West South Tasman Rise was close to the rocks of south-easternmost South Australia and western Victoria (Berry et al., 2007).

http://dx.doi.org/10.1016/j.tecto.2015.01.026

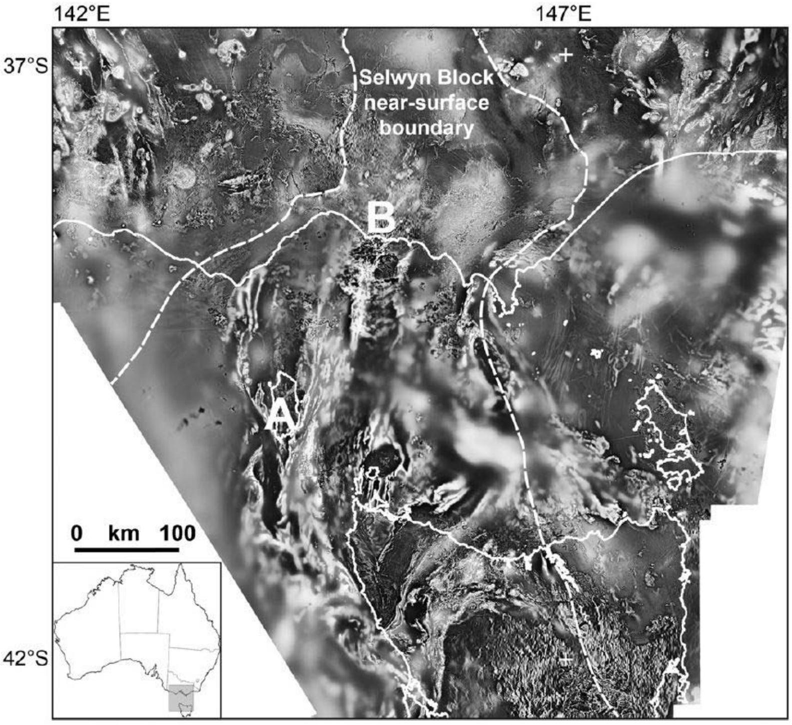


Fig. 1. Bass Strait and environs, total magnetic intensity reduced to the pole combined with 50% first vertical derivative, total magnetic intensity. White, magnetic high, black, magnetic low. A marks the location of 579 Ma mafic volcanic rocks on King Island and B the location of similar rocks on the south coast of Victoria. Inset shows the location with respect to the rest of Australia.

The East South Tasman Rise was most likely in its present position and the East Tasman Plateau was separated from the East South Tasman Rise during the Tasman Sea opening (Berry et al., 2008; Moore et al., 2014). The boundary between the West and East South Tasman Rises trends at approximately 350° (Exon et al., 1997; Fig. 18 Moore et al., 2014). In their Fig. 5, the boundary between the East and West South Tasman Rises trends at approximately 315°. How do Jacob and Dyment (2014) reconcile the lithologies, age data and trend differences in these papers with their placement of the various parts of the South Tasman Rise? How do they explain any overlap that their reconstruction gives rise to?

Other problems may be present. Can they clearly demonstrate that their pseudo-isochrons determined from the gravity data do not result from peridotites, which are known to be present along the Antarctic margin and which have previously been interpreted as isochrons (Colwell et al., 2006; Direen et al., 2012)? Such 'pseudo-isochrons' would be meaningless. Using seismic, magnetic and gravity data, Colwell et al. (2006) demonstrated that in at least one area, a previously interpreted 'chron 34' was inboard of the Antarctic continent–ocean transition and was more likely to result from serpentinised mantle and basalt. Although the Whittaker et al. (2007) reference quoted by Jacob and Dyment (2014) used what they considered to be chrons older than chron 34 (83 Ma), Tikku and Direen (2008) suggested that spreading features older than chron 32 (71 Ma) were "speculative". It may be no coincidence that a subsequent paper co-authored by Whittaker (Williams et al., 2012) considered that the oldest anomalies off the Bight Basin were at chron 34 (83 Ma). How do Jacob and Dyment (2014) reconcile this 83 Ma Bight Basin separation age with their conclusion that spreading started at 130 Ma and was completed by 84 Ma?

This paper shows the perils of largely relying on a single geophysical data set and ignoring the geology and other available data sets. I look forward to a reinterpretation of their data that takes all of the relevant material, but particularly the geology, into account.

Peter Betts and Lloyd White are thanked for reviewing drafts of this manuscript and suggesting valuable improvements. Their help is gratefully acknowledged. The recent papers by the author are part of a current PhD project at Monash University.

References

Aitken, A.R.A., Young, D.A., Ferraccioli, F., Betts, P.G., Greenbaum, J.S., Richter, T.G., Roberts, J.L., Blankenship, D.D., Siegert, M.J., 2014. The subglacial geology of Wilkes Land, East Antarctica. Geophys. Res. Lett. 41, 2390–2400. http://dx.doi.org/10.1002/2014GL059405.

Berry, R.F., Chmielowski, R.M., Steele, D.A., Meffre, S., 2007. Chemical U–Th–Pb monazite dating of the Cambrian Tyennan Orogeny, Tasmania. Aust. J. Earth Sci. 54, 757–771. http://dx.doi.org/10.1080/08120090701305269.

Berry, R.F., Steele, D.A., Meffre, S., 2008. Proterozoic metamorphism in Tasmania: implications for tectonic reconstructions. Precambrian Res. 166, 387–396. http://dx.doi.org/10.1016/j.precamres.2007.05.004.

Boger, S.D., 2011. Antarctica — before and after Gondwana. Gondwana Res. 19, 335–371. http://dx.doi.org/10.1016/j.gr.2010.09.003.

Bradshaw, J.D., 2007. The Ross Orogen and Lachlan Fold Belt in Marie Byrd Land, Northern Victoria Land and New Zealand: implication for the tectonic setting of the Lachlan Fold Belt in Antarctica. In: Cooper, A.K., Raymond, C.R. (Eds.), Antarctica: a keystone in a changing world—online proceedings of the 10th International Symposium on Antarctic Earth Sciences. United States Geological Survey Open File Report 2007-1047, Short Research Paper 059, p. 4.

Briguglio, D., Kowalczyk, J., Stilwell, J.D., Hall, M., Coffa, A., 2013. Detailed paleogeographic evolution of the Bass Basin: Late Cretaceous to present. Aust. J. Earth Sci. 1–16 http://dx.doi.org/10.1080/08120099.2013.826282.

Calver, C.R., Black, L.P., Everard, J.L., Seymour, D.B., 2004. U–Pb zircon age constraints on late Neoproterozoic glaciation in Tasmania. Geology 32, 893–896. http://dx.doi.org/10.1130/G20713.1.

Cayley, R.A., Taylor, D.H., VandenBerg, A.H.M., Moore, D.H., 2002. Proterozoic – Early Palaeozoic rocks and the Tyennan Orogeny in central Victoria: the Selwyn Block and its tectonic implications. Aust. J. Earth Sci. 49, 225–254. http://dx.doi.org/10.1046/j.1440-0952.2002.00921.x.

Cayley, R.A., Korsch, R.J., Moore, D.H., Costelloe, R.D., Nakamura, A., Willman, C.E., Rawling, T.J., Morand, V.J., Skladzien, P.B., O'Shea, P.J., 2011. Crustal architecture of Central Victoria: results from the 2006 deep crustal reflection seismic survey. Aust. J. Earth Sci. 58, 113–156. http://dx.doi.org/10.1080/08120099.2011.543151.

Colwell, J.B., Stagg, H.M.J., Direen, N.G., Bernardel, G., Borissova, I., 2006. The Structure of the Continental Margin off Wilkes Land and Terre Adélie Coast, East Antarctica. In: Fütterer, D.K., Damaske, D., Kleinschmidt, G., Miller, H., Tessensohn, F. (Eds.), Antarctica: Contributions to Global Earth Sciences. Springer-Verlag, Berlin, Heidelberg, New York, pp. 327–340.

Direen, N.G., 2012. Comment on "Antarctica — Before and after Gondwana" by S.D. Boger Gondwana Research, Volume 19, Issue 2, March 2011, Pages 335–371. Gondwana Res. 21, 302–304. http://dx.doi.org/10.1016/j.gr.2011.08.008.

Direen, N.G., Stagg, H.M.J., Symonds, P.A., Norton, I.O., 2012. In: Mohriak, W.U., Danforth, A., Post, P.J., Brown, D.E., Tari, G.C., Nemčok, M., Sinha, S.T. (Eds.), Variations in rift symmetry: cautionary examples from the Southern Rift System (Australia–Antarctica). Geological Society, London, Special Publication 369.

Exon, N.F., Moore, A.M.G., Hill, P.J., 1997. Geological framework of the South Tasman Rise, south of Tasmania, and its sedimentary basins. Aust. J. Earth Sci. 44, 561–577. http://dx.doi.org/10.1080/08120099708728337.

Goodge, J.W., Fanning, C.M., 2010. Composition and age of the East Antarctic Shield in eastern Wilkes Land determined by proxy from Oligocene–Pleistocene glaciomarine sediment and Beacon Supergroup sandstones, Antarctica. Geol. Soc. Am. Bull. 122, 1135–1159. http://dx.doi.org/10.1130/B30079.1.

Henry, D.A., Birch, W.D., 1992. Cambrian greenstone on Phillip Island, Victoria. Aust. J. Earth Sci. 39, 567–575. http://dx.doi.org/10.1080/08120099208728050.

Jacob, J., Dyment, J., 2014. Early opening of Australia and Antarctica: new inferences and regional consequences. Tectonophysics 636, 244–256. http://dx.doi.org/10.1016/j.tecto.2014.08.020.

Meffre, S., Direen, N.G., Crawford, A.J., Kamenetsky, V., 2004. Mafic volcanic rocks on King Island, Tasmania: evidence for 579 Ma break-up in east Gondwana. Precambrian Res. 135, 177–191. http://dx.doi.org/10.1016/j.precamres.2004.08.004.

Moore, D.H., Betts, P.G., Hall, W.D.M., 2013. Towards understanding the early Gondwanan margin in southeastern Australia. Gondwana Res. 23, 1581–1598. http://dx.doi.org/10.1016/j.gr.2012.08.006.

Moore, D.H., Betts, P.G., Hall, M., 2014. Fragmented Tasmania: the transition from Rodinia to Gondwana. Aust. J. Earth Sci. http://dx.doi.org/10.1080/08120099.2014.966757.

Münker, C., Crawford, A.J., 2000. Cambrian arc evolution along the SE Gondwana active margin: a synthesis from Tasmania–New Zealand–Australia–Antarctica correlations. Tectonics 19, 415–432 (doi:http://dx.doi.org/0278-7407/00/2000TC900002512).

Payne, J.L., Hand, M., Barovich, K.M., Reid, A., Evans, D.A.D., 2009. Correlations and reconstruction models for the 2500–1500 Ma evolution of the Mawson Continent. In: Reddy, S.M., Mazumder, R., Evans, D.A.D., Collins, A.S. (Eds.), Palaeoproterozoic supercontinents and global evolutionSpecial Publication 323. Geological Society of London, London, pp. 319–355.

Pilia, S., Rawlinson, N., Direen, N.G., Reading, A.M., Cayley, R., Pryer, L., Arroucau, P., Duffett, M., 2015. Linking mainland Australia and Tasmania using ambient seismic noise tomography: Implications for the tectonic evolution of the east Gondwana margin. Gondwana Res. 28, 1212–1227. http://dx.doi.org/10.1016/j.gr.2014.09.014.

Stump, E., White, A.J.R., Borg, S.G., 1986. Reconstruction of Australia and Antarctica: evidence from granites and recent mapping. Earth Planet. Sci. Lett. 79, 348–360.

Tikku, A.A., Direen, N.G., 2008. Comment on "Major Australian–Antarctic plate reorganization at Hawaiian-Emperor bend time". Science 321, 490. http://dx.doi.org/10.1126/science.1157163.

Veevers, J.J., 2012. Reconstructions before rifting and drifting reveal the geological connections between Antarctica and its conjugates in Gondwanaland. Earth-Sci. Rev. 111, 249–318. http://dx.doi.org/10.1016/j.earscirev.2011.11.009.

White, L.T., Gibson, G.M., Lister, G.S., 2013. A reassessment of paleogeographic reconstructions of eastern Gondwana: bringing geology back into the equation. Gondwana Res. 24, 984–998. http://dx.doi.org/10.1016/j.gr.2013.06.009.

Whittaker, J.M., Müller, R.D., Leitchenkov, G., Stagg, H., Sdrolias, M., Gaina, C., Goncharov, A., 2007. Major Australian–Antarctic plate reorganization at Hawaiian-Emperor bend time. Science 318, 83–86. http://dx.doi.org/10.1126/science.1143769.

Williams, S.E., Whittaker, J., Muller, R.D., 2012. Full-fit reconstructions of the southern Australian margin and Antarctica—implications for correlating geology between Australia and Antarctica. Proceedings of the Eastern Australasian Basins Symposium IV, Brisbane Qld, pp. 1–8.

Appendix 4

FIELD WORK

FIELDWORK

During the course of the research described in this thesis, approximately three months were spent on field work in Tasmania, King Island and southern Victoria. This had two aims

- To become familiar with the western Tasmanian rocks and to check the publicly available mapping sourced from Mineral Resources Tasmania. I had only had cursory visits to western Tasmania, and I had last worked on King Island in 1973 on engineering geology projects associated with the mines there. I found the Mineral Resources Tasmania mapping to be of a uniformly high standard and an invaluable guide to my own studies. Even so, some critical areas, like the coastal outcrops of the Penguin Fault, had not been mapped, and their significance only became apparent well into the interpretation phase. Furthermore, almost all of the reasonably accessible and useful outcrops were along coastal rock platforms that are only exposed at low tide as they have tidal ranges in excess of 3 m; and

- To take susceptibility measurements that could be used in any subsequent forward modelling studies. Overall, in excess of 4,000 measurements were taken, and these are included in digital files associated with this thesis.

Less than one week was spent on field work in Victoria. This was partly due to the lack of outcrop and partly to being more familiar with Victorian lithologies. Even so, over 700 susceptibility measurements were taken.

The benefit of this field work may not always be obvious in the text of this thesis. Even so, without it my understanding of the geology of the region would have been much the poorer.

Appendix 5

SENSITIVITY ANALYSIS

Sensitivity Analysis

For a given gravity or magnetic profile, an infinite number of possible solutions could be used to satisfy the extracted profiles (e.g. Whiting, 1986). The forward modeller can use the known geology to limit the possibilities somewhat, but even so, the solutions are almost always non-unique (Nabighian et al., 2005a; Nabighian et al., 2005b). Thus, the forward magnetic and gravity models presented in Chapter 4 have ambiguity. This discussion examines Section 4, the King Island North section and Section 5, through Wilsons Promontory, to examine how varying particular geometric properties of the units might influence variations in the calculated fit when compared to the extracted profile. These sections have been chosen as the magnetic and gravity responses most closely represent the Paleozoic and Proterozoic geology. Others are more strongly influenced by the magnetic Tasmanian Dolerite and Cretaceous and Cenozoic volcanic rocks.

The sections have been slightly simplified, but most characteristics are preserved. In both selected sections, the mantle, the Smithton Basin and the Bass Basin have been left constant as far as possible, with only the steep dips in the basement investigated. The original dips assigned are compared against dips of 45°E, 20°E, vertical and 45°W. Where older faults are inferred to control younger units, like in the Bass Basin, the younger unit has been modified as necessary. The magnetic susceptibilities and densities assigned to the original blocks are retained.

A5.1 SECTION 4, KING ISLAND NORTH

Because much of the contrast in the physical properties of the models was least in the eastern parts of the section, significant variation in the dips there makes relatively little difference to misfits in the models. The main variations in the magnetic and gravity responses are attributed to susceptibility, density and thickness variations in the granite bodies that are interpreted to be widely present and that crop out on many islands in eastern Bass Strait; these properties were not varied. Only modest variations to the properties of the turbidite rocks are modelled in the original sections, and so variations in their attitudes cause little variation in the model. Only minor amounts of MORB tholeiite are modelled close to the surface, and so varying the attitudes of these causes only small changes in the misfits. Even so, examination of the misfits (Figure A5-4) suggests that the steeply east-dipping boundaries and faults are a good fit, notably at the Burnie Zone-Tabberabbera Zone boundary. This accords with the near-surface geological cross sections in Tasmania by workers in Tasmania (e.g. Powell and Baillie, 1992; Patison et al., 2001; Reed and Vicary, 2005) and Victoria (e.g. VandenBerg et al., 2000; Spaggiari et al., 2004).

Forward Model	Magnetic Misfit RMS	Gravity Misfit RMS
Original	18. 468	2.140
Vertical	37.482	5.045
45°E	36.925	5.545
20°E	64.582	7.655
45°W	56.918	5.384

Table A5-1. Comparison of misfits from the final forward model in Section 4, King Island North, in Chapter 4 with models with different fixed dips. The misfits for original forward model here are larger than those from the one presented in Chapter 4 as the model here has been simplified.

By contrast, varying the dips of the basement rocks in the eastern part of the section causes significant changes in the misfits and demonstrates that the model presented is robust. Both the near-surface effects, as indicated in the magnetic solution, and the deeper effects, as indicated in the gravity solution, support the offered solution. The steep west dips of the body at approximately 50 km on the section line are suggested by the comparatively close fits of the west-dipping solution (Figure A5-4) when compared to other possibilities. The dense mafic rocks in the eastern King Island Zone are best represented with steep west dips of the final proffered model, with other solutions providing gravity lows at the edge of the Bass Basin.

The magnetic response of the eastern boundary of the King Island Zone is partly obscured by a local lens of ?Mesozoic basalt. As well, the Bass Basin is interpreted to be 0.6 km thicker in the westernmost Rocky Cape Zone, and so the gravity response varies significantly with the change in dip. Even so, steepening or flattening the mafic volcanic rocks at the eastern margin of the King Island Zone causes major misfits in the region as sedimentary rocks are substitute for the denser mafic rocks.

Overall, the final model presented offers a more robust solution to the magnetic and gravity profiles extracted than other possibilities presented here.

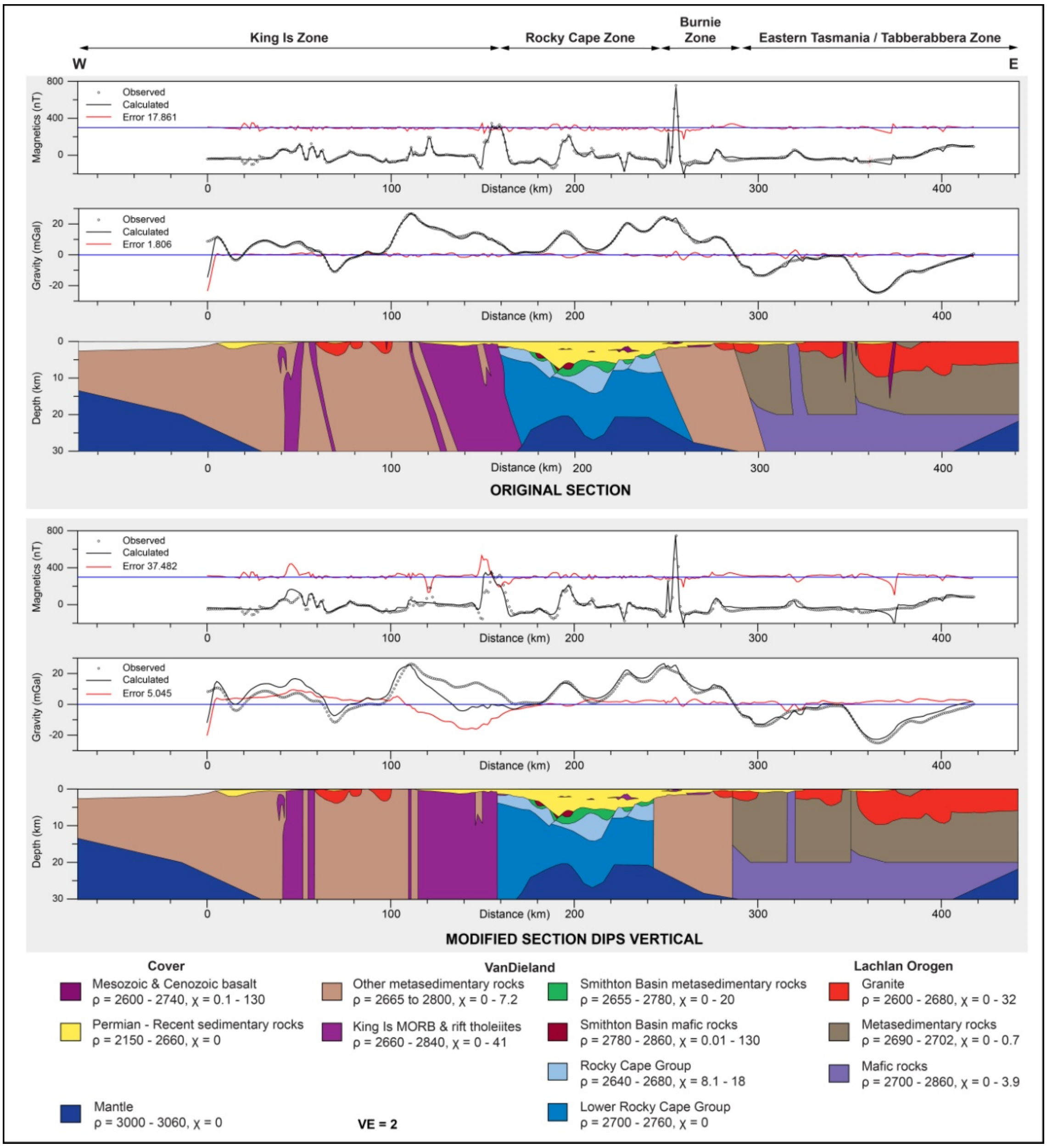


Figure A5-1. Original Section 4, King Island North, and after subsequently modifying the dips of the basement to vertical. Section location indicated on Figure 4-2.

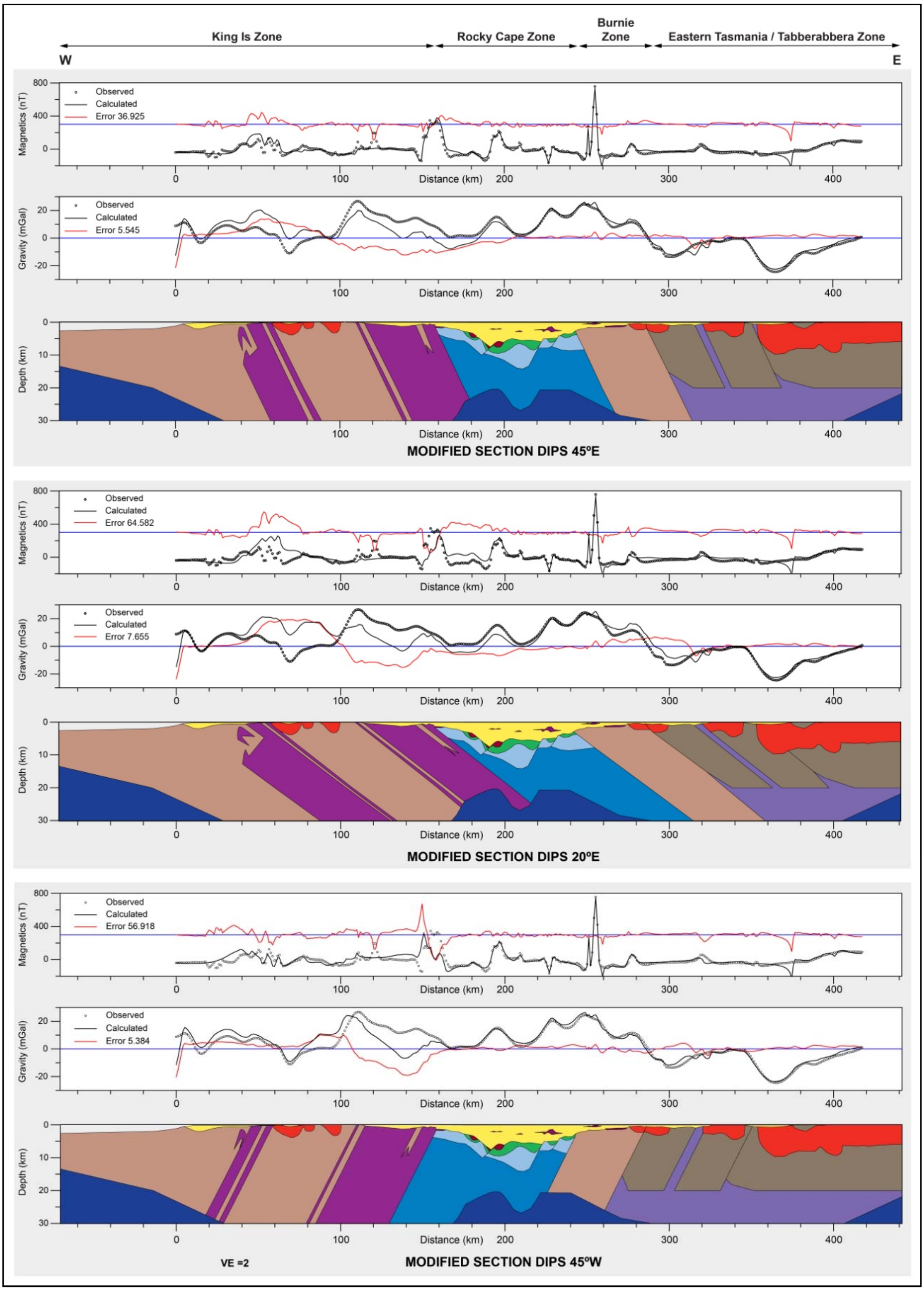


Figure A5-2. Section 4, King Island North, after modifying the dips of the basement to 45°E (top) 20°E (middle) and 45°W. Section location indicated on Figure 4-2. Same colour legend as for Figure A5-1.

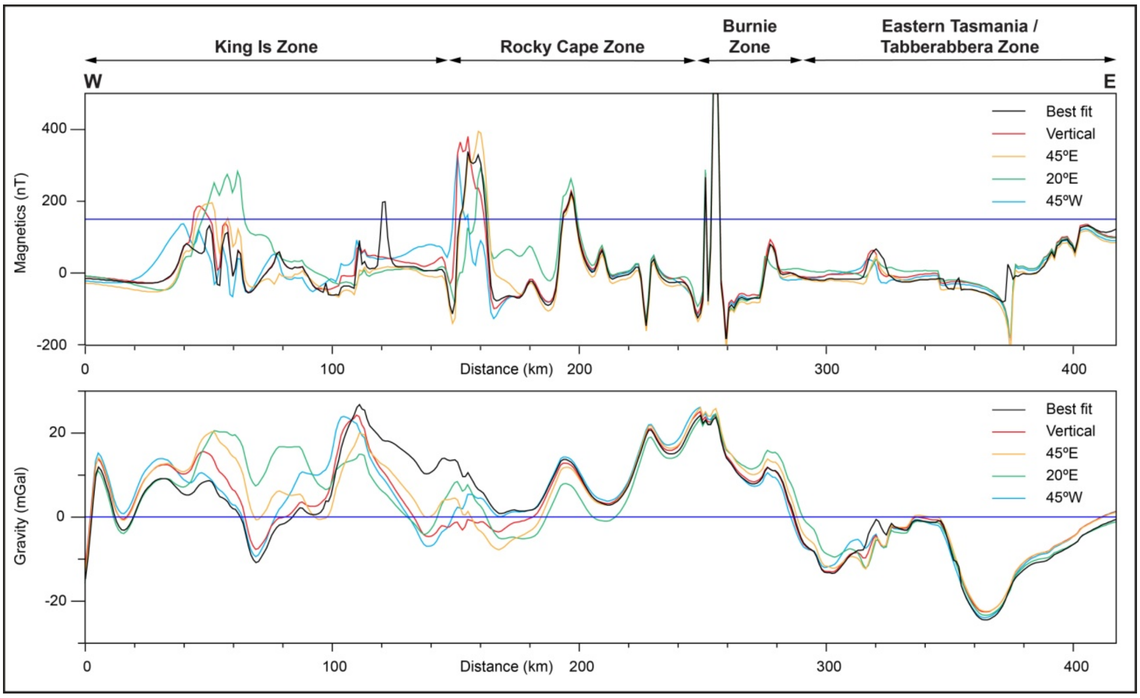


Figure A5-3. King Island North, magnetic and gravity profiles from Figures A5-1 and A5-2 compared.

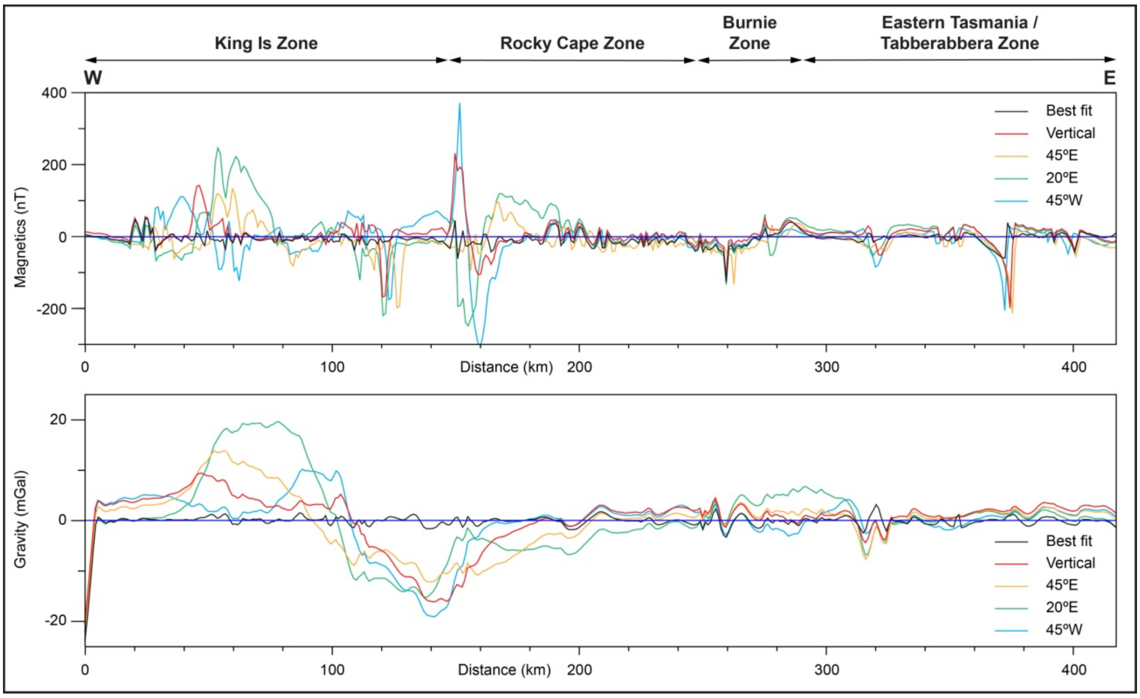


Figure A5-4. King Island North, misfits from Figures A5-1 and A5-2 compared.

A 5.2 SECTION 5, WILSONS PROMONTORY

Similar to Section 4, King Island North, the eastern part of the section showed relatively minor variations in the misfits with changes in the modelled dips (Figures A5-5 to A5-8). The final submitted model and that with vertical dips were particularly close, suggesting that the dips in the original section could have been steepened with little increase in misfit. Even so, there is a consistent pattern, with the section with vertical boundaries closest to the submitted model, with dips of 45°E and 45°W the next closest, with the poorest fit from the model with dips of 20°E. The boundary between the Tabberabbera Zone and the Burnie Zone is modelled with a dip of 40°E, but a vertical dip fits the profile almost as well (Figure A5-8). The large misfit in the gravity data at approximately 380 km is where the profile crosses Wilsons Promontory, and is assumed to be due to poor stitching of the onshore and offshore data.

Forward Model	Magnetic Misfit RMS	Gravity Misfit RMS
Original	25.331	1.784
Vertical	27.812	1.918
45°E	42.365	3.879
20°E	57.108	8.325
45°W	49.588	4.077

Table A5-2. Comparison of misfits from the final forward model in Section 5, Wilsons Promontory, in Chapter 4 with models with different fixed dips. The misfits for original forward model here are larger than those from the one presented in Chapter 4 as the model here has been simplified.

Much of the high frequency misfits at the western end of the section may be due to variations in water depth, which are modelled at more than 1500 m, in an area where submarine canyons can be seen in images of digital data (Webster and Petkovic, 2005). The gravity data implies that the boundary between the Delamerian Orogen and the King Island Zone, at approximately 60 km, dips to the west. Here the Delamerian Orogen is interpreted to be comprised of magnetic, dense rocks. A model with a vertical boundary has the gravity and magnetic responses as too high to the east and the gravity response is too low to the west. Other solutions look even less likely as they create even larger misfits in the model.

The mafic volcanic rocks on the eastern margin of the King Island Zone (from 190 to 220 km on the section) are generally modelled as having internal dips of approximately 70° to 80°E, with the eastern boundary dipping vertically. Steepening or shallowing of the dips causes misfits at the western edge of the gravity models, and the models with

45°E, 20°E and 45°W dips also showing magnetic and gravity misfits at the eastern boundary with the adjacent Rocky Cape Zone.

Overall, the final model presented offers a more robust solution to the magnetic and gravity profiles extracted than other possibilities presented here.

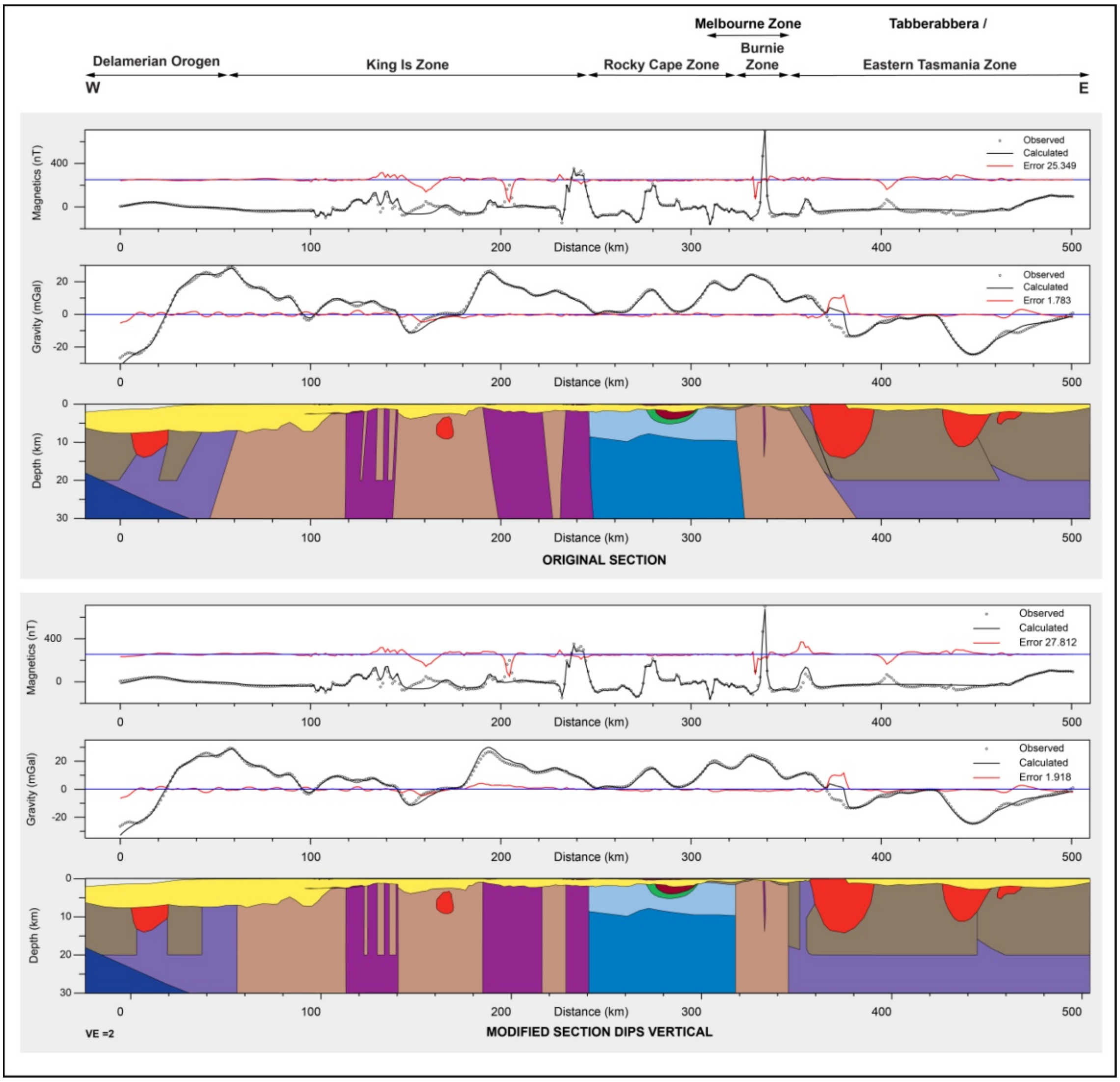


Figure A5-5. Original Section 5, Wilsons Promontory, and after subsequently modifying the dips of the basement to vertical. Section location indicated on Figure 4-2. Same colour legend as for Figure A5-1.

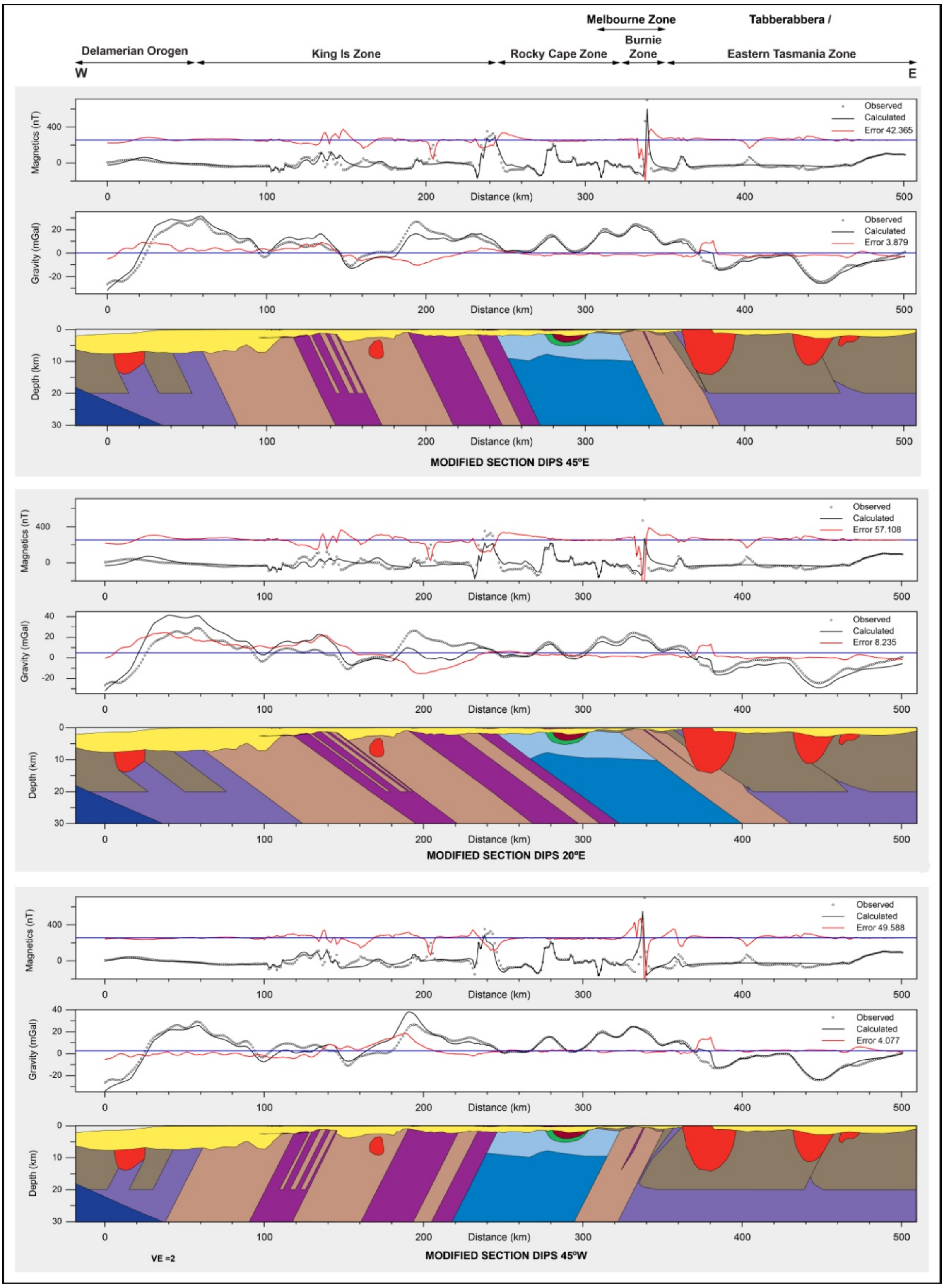


Figure A5-6. Section 5, Wilsons Promontory, after modifying the dips of the basement to 45°E (top) 20°E (middle) and 45°W. Section location indicated on Figure 4-2. Same colour legend as for Figure A5-1.

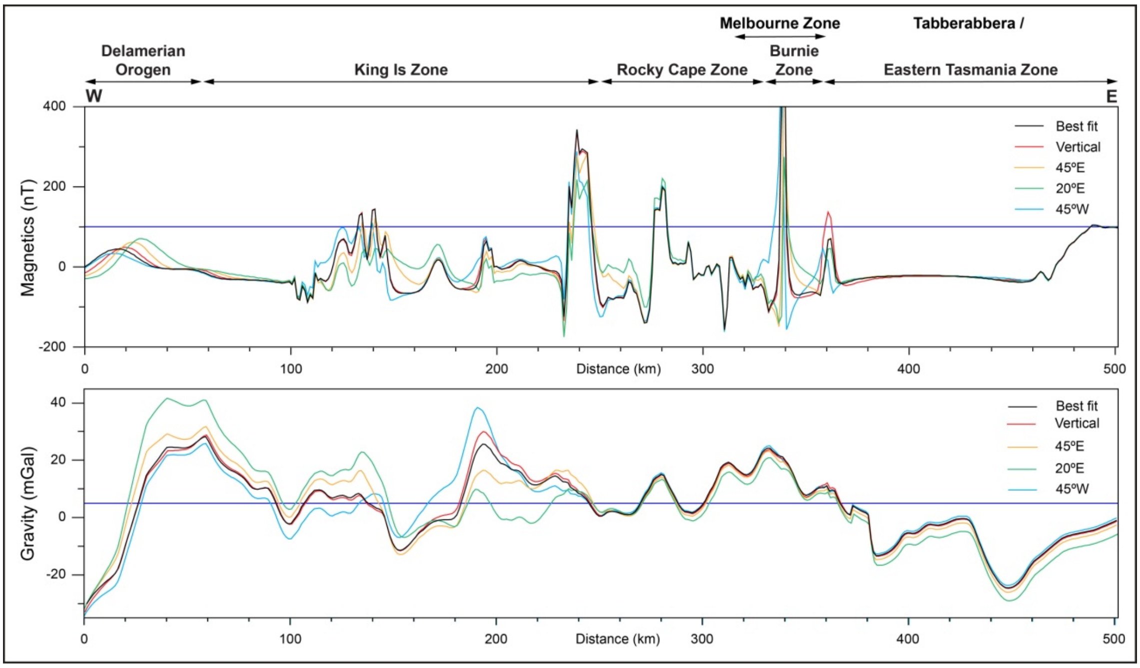


Figure A5-7. Section 5, Wilsons Promontory, magnetic and gravity profiles from Figures A5-1 and A5-2 compared.

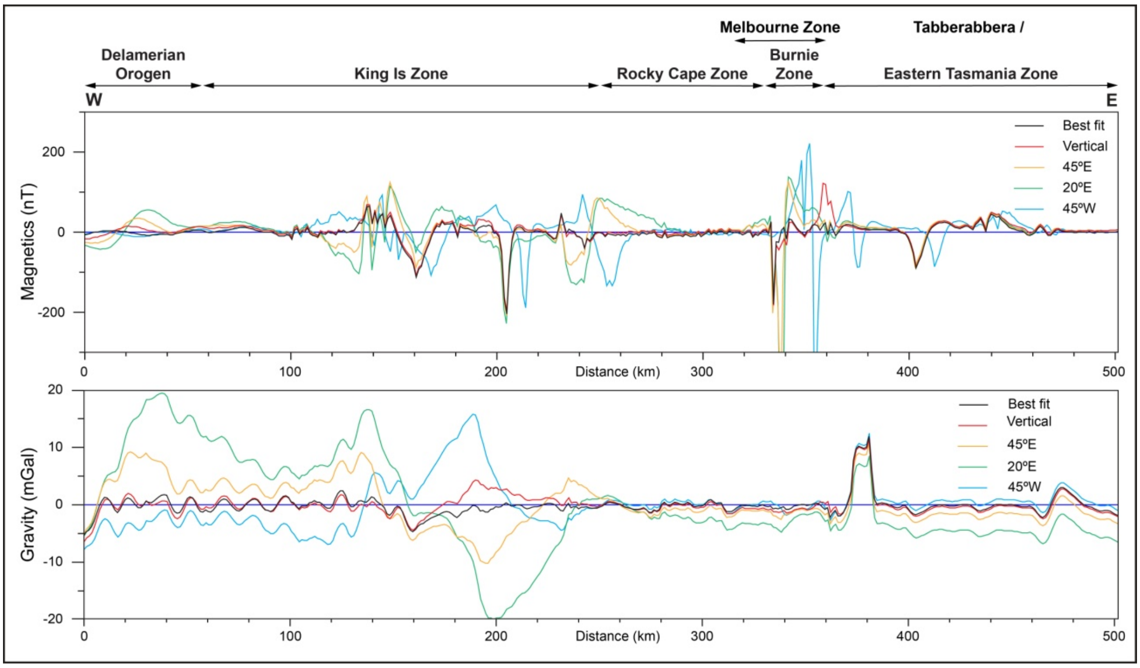


Figure A5-7. Section 5, Wilsons Promontory, misfits from Figures A5-5 and A5-6 compared.

REFERENCES

Nabighian, M.N., Ander, M.E., Grauch, V.J.S., Hansen, R.O., LaFehr, T.R., Li, Y., Pearson, W.C., Peirce, J.W., Phillips, J.D., Ruder, M.E., 2005a. Historical development of the gravity method in exploration. Geophysics 70, 63ND-89ND. doi:10.1190/1.2133785.

Nabighian, M.N., Grauch, V.J.S., Hansen, R.O., LaFehr, T.R., Li, Y., Peirce, J.W., Phillips, J.D., Ruder, M.E., 2005b. The historical development of the magnetic method in exploration. Geophysics 70, 33ND-61ND. doi:10.1190/1.2133784.

Patison, N.L., Berry, R.F., Davidson, G.J., Taylor, B.P., Bottrill, R.S., Manzi, B., Ryba, J., R.E.Shepherd, 2001. Regional metamorphism of the Mathinna Group, northeast Tasmania. Australian Journal of Earth Sciences 48, 281-292. doi:10.1046/j.1440-0952.2001.00862.x.

Powell, C.M., Baillie, P.W., 1992. Tectonic affinity of the Mathinna Group in the Lachlan Fold Belt. Tectonophysics 214, 193-209. doi:10.1016/0040-1951(92)90197-E.

Reed, A., Vicary, M.J., 2005. Sheet 4843 Beaconsfield. Digital Geological Atlas 1:25 000 series. Mineral Resources Tasmania.

Spaggiari, C.V., Gray, D.R., Foster, D.A., 2004. Ophiolite accretion in the Lachlan Orogen, southeastern Australia. Journal of Structural Geology 26, 87-112. doi:http://dx.doi.org/10.1016/S0191-8141(03)00084-1.

VandenBerg, A.H.M., Willman, C.E., Maher, S., Simons, B.A., Cayley, R.A., Taylor, D.H., Morand, V.J., Moore, D.H., Radojkovic, A., 2000. The Tasman Fold Belt System in Victoria, Special Publication. Geological Survey of Victoria, p. 463.

Webster, M., Petkovic, P., 2005. Australian bathymetry and topography grid, June 2005. Geoscience Australia Record 12, 30.

Whiting, T.H., 1986. Aeromagnetics as an aid to geological mapping; a case history from the Arunta Inlier, Northern Territory. Australian Journal of Earth Sciences 33, 271-286.

Appendix 6

TALKS AND PRESENTATIONS

Digital file

Appendix 7

GEOPHYSICAL MODELS

Digital file

Appendix 8

GIS INTERPRETATIONS

Digital file

Appendix9

MEASURED SUSCEPTIBILITIES

Digital file