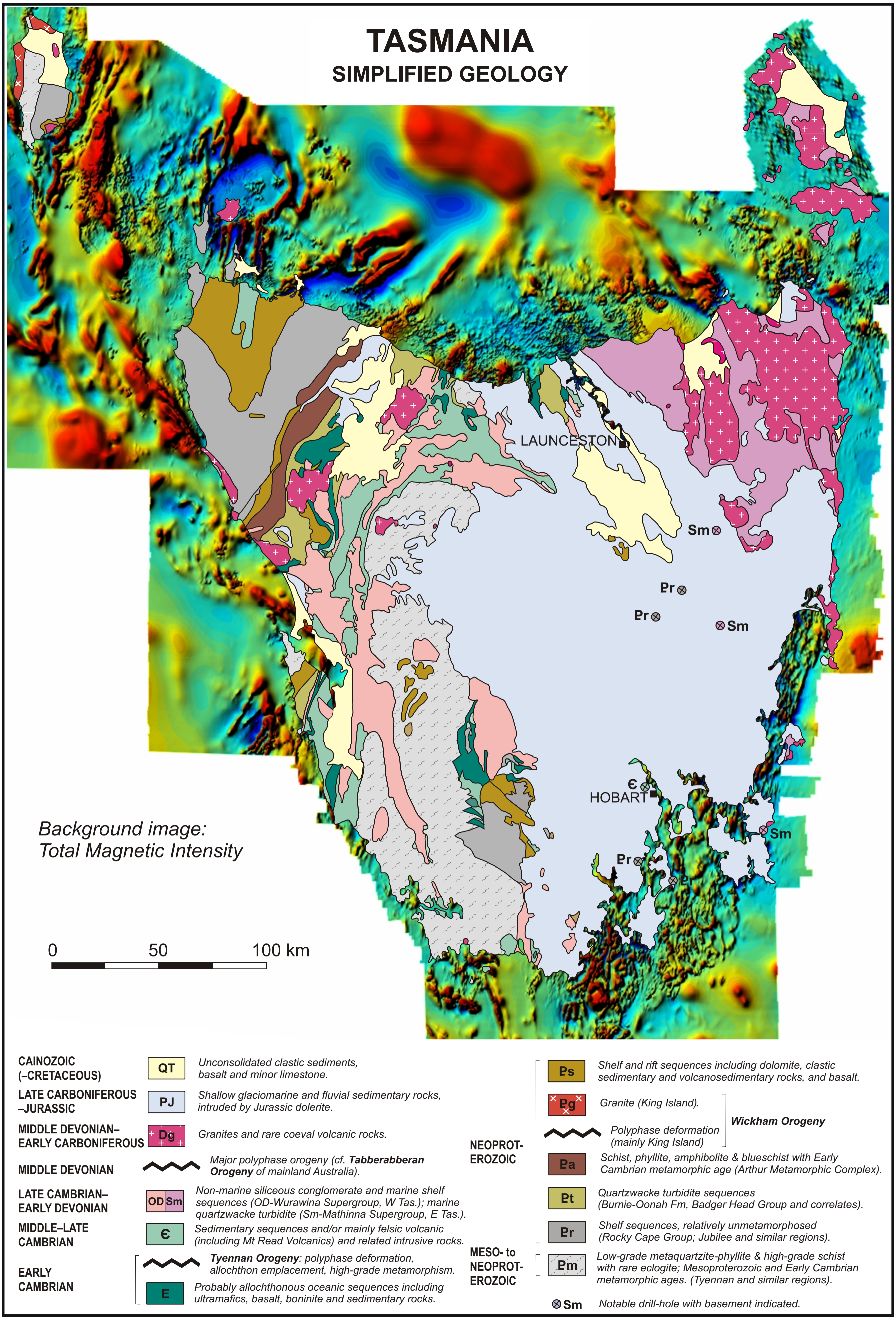
Tasmanian Geological Survey
Bulletin 72

# The Geology and Mineral Deposits of Tasmania: a summary

Tasmania

DEPARTMENT *of* INFRASTRUCTURE, ENERGY *and* RESOURCES
Mineral Resources Tasmania


TASMANIA
SIMPLIFIED GEOLOGY
LAUNCESTON
HOBART
Sm
Ꝑr
Sm
Ꝑr
Sm
Ꝑr
Є
Background image:
Total Magnetic Intensity
0 50 100 km
CAINOZOIC (–CRETACEOUS)
QT
Unconsolidated clastic sediments, basalt and minor limestone.
LATE CARBONIFEROUS –JURASSIC
PJ
Shallow glaciomarine and fluvial sedimentary rocks, intruded by Jurassic dolerite.
MIDDLE DEVONIAN– EARLY CARBONIFEROUS
Dg
Granites and rare coeval volcanic rocks.
MIDDLE DEVONIAN
Major polyphase orogeny (cf. Tabberabberan Orogeny of mainland Australia).
LATE CAMBRIAN– EARLY DEVONIAN
OD Sm
Non-marine siliceous conglomerate and marine shelf sequences (OD-Wurawina Supergroup, W Tas.); marine quartzwacke turbidite (Sm-Mathinna Supergroup, E Tas.).
MIDDLE–LATE CAMBRIAN
Є
Sedimentary sequences and/or mainly felsic volcanic (including Mt Read Volcanics) and related intrusive rocks.
EARLY CAMBRIAN
Tyennan Orogeny: polyphase deformation, allochthon emplacement, high-grade metamorphism.
E
Probably allochthonous oceanic sequences including ultramafics, basalt, boninite and sedimentary rocks.
NEOPROT- EROZOIC
Ꝑs
Shelf and rift sequences including dolomite, clastic sedimentary and volcanosedimentary rocks, and basalt.
Ꝑg
Granite (King Island).
Polyphase deformation (mainly King Island)
Wickham Orogeny
Ꝑa
Schist, phyllite, amphibolite & blueschist with Early Cambrian metamorphic age (Arthur Metamorphic Complex).
Ꝑt
Quartzwacke turbidite sequences (Burnie-Oonah Fm, Badger Head Group and correlates).
Ꝑr
Shelf sequences, relatively unmetamorphosed (Rocky Cape Group; Jubilee and similar regions).
MESO- to NEOPROT- EROZOIC
Ꝑm
Low-grade metaquartzite-phyllite & high-grade schist with rare eclogite; Mesoproterozoic and Early Cambrian metamorphic ages. (Tyennan and similar regions).
⊗Sm
Notable drill-hole with basement indicated.

MINERAL RESOURCES TASMANIA
DEPARTMENT *of* INFRASTRUCTURE, ENERGY *and* RESOURCES

## *Geological Survey Bulletin 72*

# The Geology and Mineral Deposits of Tasmania: a summary

*by D. B. Seymour, G. R. Green and C. R. Calver*

**February 2006**
**ISBN 0 7246 4017 7**

Mineral Resources Tasmania PO Box 56 Rosny Park Tasmania 7018
Phone: (03) 6233 8377 ● Fax: (03) 6233 8338
Email: info@mrt.tas.gov.au ● Internet: www.mrt.tas.gov.au

## CONTENTS



## Figures

## Introduction

Despite its small size, Tasmania has a remarkable geological diversity and abundance of mineral deposits. Rocks from every period of the Earth's history from the Middle Proterozoic are present and there have been at least four major episodes of economic mineralisation. Significant mineral deposits include Proterozoic iron ore, silica, dolomite and magnesite; Cambrian VHMS base metal-gold and ultramafic-related platinum group minerals (PGM) and chromite; Devonian slate-belt gold deposits; Devonian granite-related tin, tungsten, fluorite, magnetite, silver-lead-zinc and possibly nickel deposits; Triassic and Tertiary coal deposits; and Cainozoic alluvial gold, tin and PGMs, and residual iron oxide, silica and clay.

This brief summary of the current state of knowledge of the geology of Tasmania has, of necessity, relied heavily on several more lengthy syntheses, both published and in preparation. Sources include Seymour and Calver (1995), Bottrill *et al*. (1998), Stacey and Berry (2004), manuscripts in preparation by C. R. Calver, R. F. Berry, G.R Green, M. P. McClenaghan and D. B. Seymour for the upcoming *Geological Evolution of Tasmania* volume (Geological Society of Australia), and information compiled for the Western Tasmanian Regional Minerals Program conducted by the Australian and Tasmanian governments, and the 3-Dimensional Geological Model and Prospectivity Analysis of Tasmania. Products from the latter two sources are available from Mineral Resources Tasmania (MRT) and more information and free downloads are available at www.mrt.tas.gov.au. Parts of these sources have been reproduced or paraphrased in this summary, and many key references within their texts are also quoted here.

## Proterozoic

### *Mesoproterozoic sequences and events*

The western part of King Island is composed of amphibolite-grade metasedimentary rocks and minor amphibolite, known as the Surprise Bay Formation, intruded by 760 Ma granitoids (fig. 1). The succession is predominantly interbedded micaceous quartzite, schist and phyllite, derived from a variety of original sedimentary facies. The assemblage includes sandy turbidites, quartzofeldspathic schist in massive units up to several metres thick, mafic intrusive rocks and pods of marble. Evidence that these rocks are older than most other Tasmanian Proterozoic successions comes from dating of monazite in the metasedimentary rocks, which indicates an early regional Grenvillean-age (1287 ± 18 Ma) metamorphic event (Berry *et al*., 2005). The youngest detrital zircons are ca. 1350 Ma (Black *et al*., 2004), indicating a depositional age of ca. 1290–1350 Ma for the Surprise Bay Formation. The detrital zircon age profile is similar to many of the Tasmanian Lower Proterozoic rocks, suggesting that western King Island may have been part of the source area for

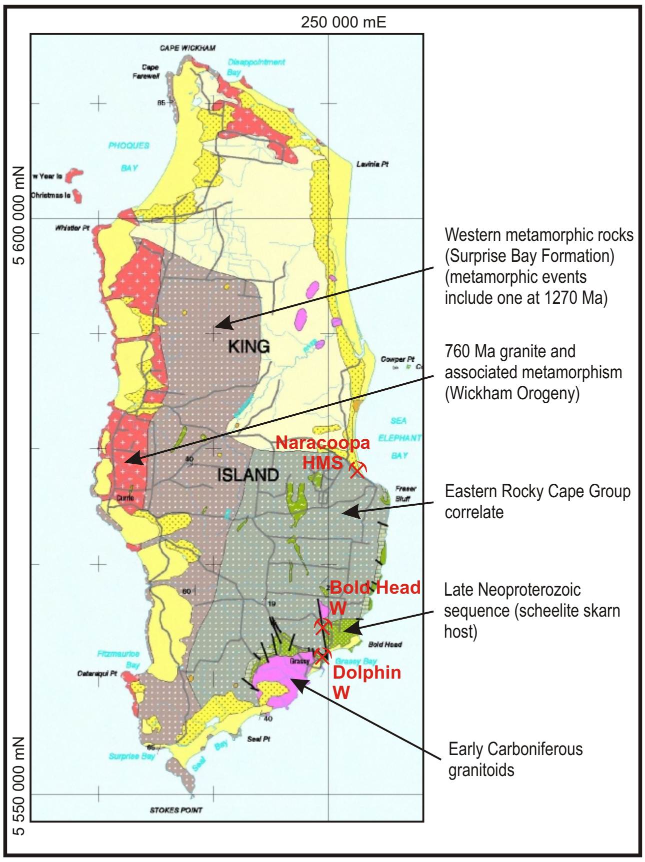



**Figure 1**

*Proterozoic rock units on King Island.*

sequences such as the Rocky Cape Group and the Oonah Formation (see below).

The earliest deformation phase ($D_1$) in the Surprise Bay Formation produced large, tight to isoclinal folds with penetrative axial surface cleavage, and is Mesoproterozoic (1287 ± 18 Ma) in age (Berry *et al*., 2005). Metamorphic conditions were 500°C and 300 MPa.

The only evidence for Mesoproterozoic basement on mainland Tasmania comes from dating of monazite in the cores of garnets in eclogite-bearing mylonitised gneiss of the Franklin Metamorphic Complex (MC), one of several high-grade metamorphic complexes in the Tyennan region (see below). The U-Th-Pb dating (R. F. Berry, University of Tasmania, personal communication) identified an early 1300 Ma metamorphic event, allowing the possibility that part of the Franklin MC is a Mesoproterozoic basement to Tasmania.

### *Tyennan region*

The largest area of exposed Proterozoic rocks on mainland Tasmania is in the Tyennan region, once thought to form a central, autochthonous basement core to Tasmania (fig. 2). Its constituent polydeformed metamorphic rocks are now considered to comprise a complex thrust stack of two metamorphic assemblages (one allochthonous), typically in mutual fault contact:

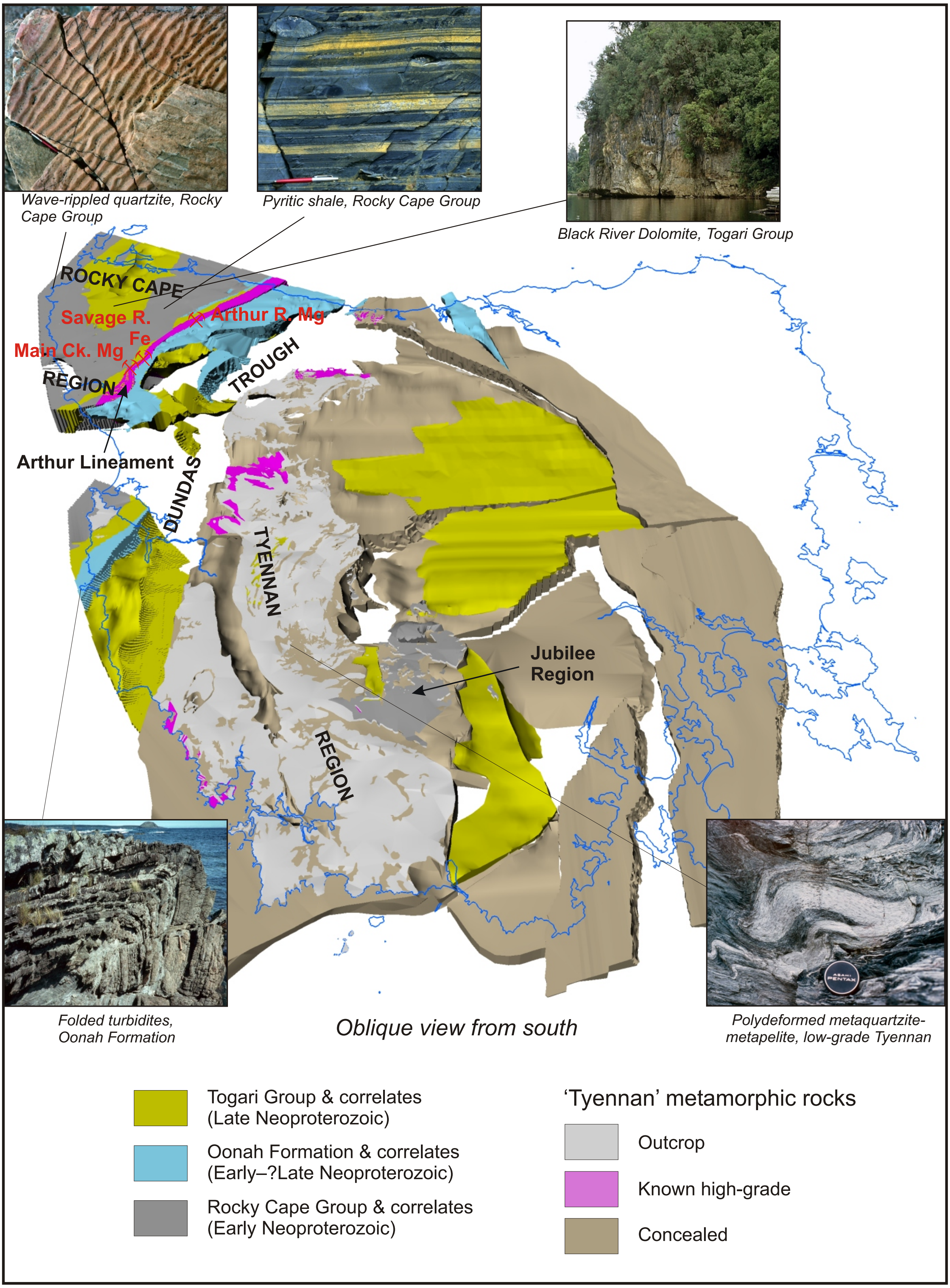



**Figure 2**

*3-D geological model of outcropping and concealed Proterozoic rock units of mainland Tasmania, with typical outcrop photographs of some of the sequences.*

- a low-grade (up to greenschist facies) assemblage of metaquartzite and graphitic and chloritic metapelite, derived from an early Neoproterozoic sedimentary sequence broadly similar to the Rocky Cape Group of northwest Tasmania (see below);
- a high-grade metamorphic assemblage of garnetiferous schist-quartzite-(amphibolite), including mafic meta-igneous rocks with metamorphic grades up to eclogite facies; the protolith of the high-grade assemblage was thought to have been Late Neoproterozoic in age, but as noted above, new dating from the Franklin MC favours a Mesoproterozoic age.

The high-grade metamorphism is attributed to the Early Cambrian Tyennan Orogeny (see below), which was probably also responsible for some of the metamorphism in the low-grade assemblage (Meffre *et al*., 2000). The high-grade rocks are believed to have been substantially translated both horizontally and vertically from their place of origin, i.e. they are allochthonous. The same interpretation is currently favoured for another belt of high-grade (up to blueschist facies) metamorphic rocks in the Arthur Lineament in northwest Tasmania (fig. 2) (see discussion of Tyennan Orogeny, below).

Currently, the high-grade metamorphic rocks are incompletely mapped within the Tyennan region and there may be more high-grade complexes to be found. Regionally, the exposed Tyennan rocks represent less than 50% of their full inferred distribution beneath Tasmania, as revealed by geophysical modelling and seismic profiling (fig. 2). The inferred distribution currently shown in the 3-D geological model may be conservative, as Tyennan-style crust may also underlie parts of northeastern Tasmania and the Palaeozoic Dundas Trough.

### *Rocky Cape Group (Lower Neoproterozoic; 1000–750 Ma)*

The Rocky Cape Group occupies a large area of northwestern Tasmania (fig. 2), where it is considered to represent a block of the autochthonous basement to Palaeozoic Tasmania, lying west of the limit of allochthon emplacement during the Tyennan Orogeny. It comprises a 10 km thick sequence (base unknown) of cross-bedded quartz sandstone (commonly indurated to quartzite), laminated siltstone, pyritic shale, and minor dolomite, deposited in an open marine shelf environment varying from low-energy and below storm wave base, to relatively high-energy between storm and fair-weather wave base, and perhaps to shallower marginal marine conditions at times. Common bipolar current patterns in cross-bedded quartz sandstone suggest tide-dominated settings, while hummocky cross-stratification in sandstone and gutter casts in siltstone attest to storm influence in parts of the succession. Rare, possible evaporite indicators occur in some siltstone units. Outcrop-scale growth faults in lower parts of the succession imply active extensional tectonics during sedimentation. The Rocky Cape Group is believed to be early Neoproterozoic in depositional age, with a maximum age limit of 1000 Ma (its youngest detrital zircon population), and a minimum age limit of 750 Ma (for the lower part of the overlying Togari Group).

As mentioned above, the low-grade metaquartzite-metapelite assemblage of the Tyennan region is considered to have been derived from lithologies similar to the Rocky Cape Group. Direct correlates of the Rocky Cape Group also occur in a thick, folded and faulted succession of quartz sandstone, siltstone, pelite and minor carbonate (the Clark Group) in the Jubilee region at the eastern margin of the Tyennan region (Calver *et al*., 1990; Calver *et al*., in press) (fig. 2). Other Rocky Cape Group correlates include a deformed succession of interbedded quartz sandstone, micaceous quartzite, phyllite and minor conglomerate in the Cape Sorell inlier (fig. 2), and a six kilometre thick un-named succession of interbedded pyritic shale, siltstone and fine-grained, muscovitic quartz sandstone forming most of the eastern half of King Island (fig. 1).

### *Oonah Formation and correlates (Lower Neoproterozoic; 1000–750 Ma)*

A thick, polydeformed Proterozoic quartzwacke turbidite succession, widespread in western and northern Tasmania, has variously been named the Oonah Formation, Burnie Formation, and 'Badger Head Group', which are regarded as correlates. The distribution suggests deposition in a basin roughly coincident with the Cambrian Dundas Trough (fig. 2), but the original basin configuration has been obscured by later events. The metamorphic grade is mostly low (sub-greenschist to low greenschist facies), with higher-grade and higher-strain schistose equivalents proximal to the Arthur Lineament and in the Dundas inlier. There is no known stratigraphic contact with the Rocky Cape Group and the base of the succession is unknown, but minimum thicknesses of two to three kilometres near Zeehan and five kilometres on the north coast have been determined. The depositional age is constrained by the youngest detrital zircons (1070 Ma) in the sequence, and the ca. 750 Ma age of the overlying Success Creek Group.

The Oonah Formation and correlates comprise two lithological associations. The predominant, quartzwacke turbidite association, which includes minor alkaline dolerite intrusions (Cooee Dolerite) and related lavas on the north coast, consists of monotonously interbedded quartz sandstone, quartzwacke, siltstone and grey-green to black pelite, with sedimentary structures typical of sandy turbidites. Derivation may have been from the Rocky Cape Group, and limited palaeocurrent data suggest northward or northeastward-directed turbidity currents, reworked by traction currents. The second lithological association, which may form an upper subdivision of the formation, is predominantly pelite and/or carbonate, including mafic volcanic rocks and conglomerate in some places. Near Zeehan, this association is host to a number of Devonian vein, skarn and replacement-tin deposits, and at Mt Bischoff a dolomitic unit hosted major Devonian tin lodes.

### *Wickham Orogeny (Middle Neoproterozoic; ca. 760 Ma)*

The schists of the Surprise Bay Formation of western King Island are intruded by syn-deformational granitoids associated with the Wickham Orogeny (Cox, 1973; Turner *et al.*, 1998). The granitoids (dominantly K-feldspar porphyritic monzogranite) are broadly synchronous with three phases of deformation which post-date the Mesoproterozoic $D_1$ fold generation in the country rocks, and which are localised to the contact aureoles of the granites. Away from King Island, the Wickham Orogeny has no clear correlate. The most likely equivalent structure in mainland Tasmania is the low angle unconformity between the lower Neoproterozoic siliciclastic sequences and the upper Neoproterozoic Togari Group and correlates, and possibly also a minor felsic intrusive phase dated at 777 ± 7 Ma within the Arthur Lineament (Turner *et al.*, 1998). However, there is substantial evidence that normal faulting accompanied late Neoproterozoic extension and the onset of Togari Group sedimentation, and the angular unconformity between the Rocky Cape Group and Togari Group might be the product of block rotation or 'roll-over' of hanging-wall strata above non-planar extensional faults (Everard *et al.*, 2001).

### *Togari Group and correlates (Upper Neoproterozoic–Lower Cambrian; 750–520 Ma)*

Late Neoproterozoic sedimentary and mafic volcanic successions of the Togari Group and its correlates (including the Ahrberg Group, Timbs Group, Success Creek Group, Crimson Creek Formation and Weld River Group) are widespread in Tasmania, with a large inferred concealed extent in the central part of the island (fig. 2). These rocks rest unconformably or disconformably upon older successions, a relationship that has been thought to reflect the Wickham Orogeny but may have other explanations. The most complete sequence, the Togari Group of the Smithton Basin, can be subdivided into four main phases of sedimentation:

1. A lower dolomitic succession with basal siliceous conglomerate-sandstone and diamictite near the top, which represents a widespread phase of mid-Cryogenian (ca. 750–700 Ma) shallow marine shelf sedimentation.

2. A phase of mafic rift volcanism and associated volcaniclastic and siliciclastic sedimentation, of late Cryogenian to early Ediacaran age (ca. 700–570 Ma).

3. A renewal of shallow-marine carbonate sedimentation later in the Ediacaran (ca. 570–545 Ma).

4. At the top, a Cambrian (probably Early Cambrian) phase of deep-water siliciclastic sedimentation.

The last two phases are apparently restricted to the Rocky Cape region. The age of the Togari Group and correlates is largely constrained by isotope stratigraphic evidence (Calver, 1996; 1998), supported by a U-Pb radiometric age of 582 ± 4 Ma from a minor felsic volcanic phase in the Kanunnah Subgroup (unit 2 mentioned above) (Calver *et al.*, 2004).

Carbonate sequences within some Togari Group correlates host economically important replacement mineralisation related to Late Devonian–Early Carboniferous granitoids. Prime examples include the Renison Bell tin deposit and the scheelite skarn deposits of southeast King Island. Togari Group dolomite is also mined for agricultural purposes and has potential for metallurgical use. Ultra-high purity silica flour is produced from weathered Togari Group dolomite near Corinna and other deposits are known in northwestern Tasmania.

## Tyennan Orogeny (Early Cambrian; ca. 515–510 Ma)

In the early phase of the Tyennan Orogeny, the east-facing Tasmanian passive margin collided with an oceanic arc, resulting in obduction of mafic-ultramafic complexes across much of Tasmania (Berry and Crawford, 1988; Crawford and Berry, 1992; Turner *et al.*, 1998) (fig. 3, 4). The original, shallowly-dipping geometry of the allochthonous sheets has been substantially disrupted by later, Cambrian and Devonian deformation, so that the present surface occurrences are typically steeply dipping and fault bounded. The late Early Cambrian emplacement age is constrained by a 513.6 ± 5.0 Ma radiometric age of crystallisation (Black *et al.*, 1997) of a minor felsic phase in one of the complexes, and the presence of ultramafic detritus in the basal beds of the overlying fossiliferous Middle Cambrian Dundas Group. An early westward-directed emplacement direction for the allochthons is preserved in some places in movement indicators in high-temperature mylonites on the soles of the mafic-ultramafic complexes (Stacey and Berry, 2004).

Three ultramafic-mafic rock associations are commonly in fault juxtaposition within the complexes: Layered Pyroxenite-Dunite (LPD); Layered Dunite-Harzburgite (LDH); and Layered Pyroxenite-Peridotite and associated Gabbro (LPG) (Brown, 1986). Igneous layering is common (fig. 4), and pseudo-sedimentary structures are present in the LPG succession. Pervasive serpentinisation is common. All of the ultramafic rocks are orthopyroxene-rich, which distinguishes them from the dominantly clinopyroxene-rich sequences which world-wide are usually associated with mid-ocean ridge and back-arc environments (Brown, 1989). Osmiridium has been mined from first and second-cycle alluvial and eluvial deposits derived from LDH-association rocks in a number of the ultramafic complexes.

Boninite and low-Ti tholeiite lavas are associated with the ultramafic complexes (Brown and Jenner, 1989) and

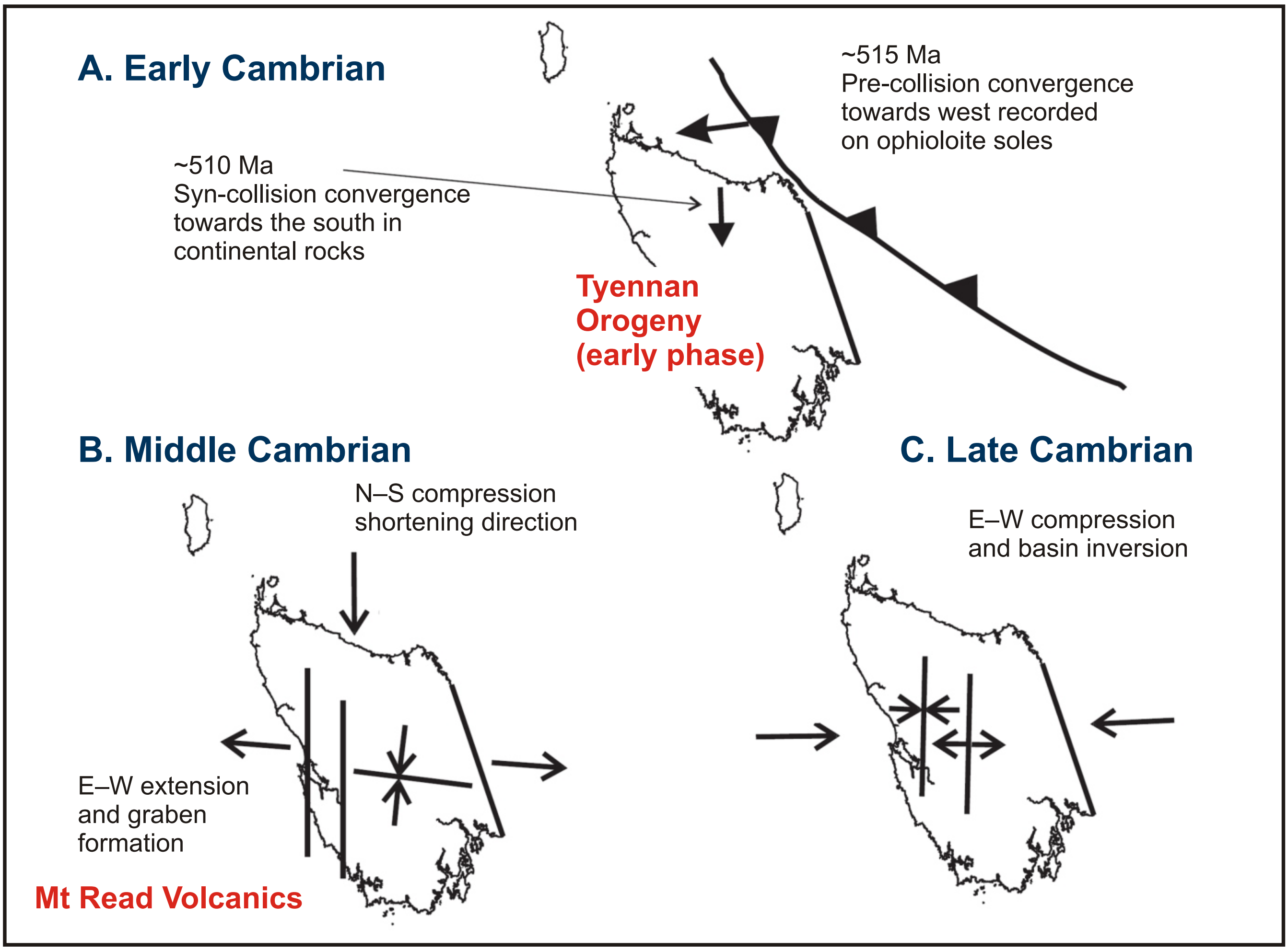



**Figure 3**

*Cambrian structural history of Tasmania, according to Stacey and Berry (2004).*

are indicative of formation in the forearc region of an oceanic island arc (Berry and Crawford, 1988).

In a possible second stage of the obduction process, fault-bounded Proterozoic units displaying anomalous high-grade metamorphism are also thought to have been emplaced (Berry, 1995; Meffre *et al*., 2000; Holm and Berry, 2002) (fig. 3). These include several metamorphic complexes within the Tyennan region and on the central north coast, and a belt of up to blueschist-facies metamorphic rocks within the Arthur Lineament (see below). They are thought to have been derived from the distal parts of the passive margin that became deeply buried or partially subducted during the collision event, then obducted or thrust westward or southward and emplaced at a high crustal level amongst the relatively unmetamorphosed autochthonous successions (Meffre *et al*., 2000; Turner and Bottrill, 2001; Holm and Berry, 2002). The allochthonous Proterozoic components thus typically comprise relict, high-pressure metamorphic rocks of Cambrian metamorphic age, as fault-bounded units with an inferred gently dipping initial structural attitude.

Other ?Early Cambrian sequences which may have been emplaced in the collisional phase of the Tyennan Orogeny show anomalous characteristics compared with adjacent autochthonous units, e.g. chert sequences of possible deep marine origin (Barrington Chert), and certain anomalous, clastic sedimentary and volcanosedimentary sequences (e.g. Cleveland–Waratah association, Ragged Basin Complex). The Cleveland–Waratah association comprises lithicwacke, red mudstone, chert, mafic volcanic rocks with Ocean Floor Basalt geochemical characteristics, and rare carbonate rocks which host Devonian skarn mineralisation at the Cleveland mine.

## Arthur Lineament

The term Arthur Lineament strictly refers to a narrow (110 km long by up to 10 km wide) tectonic feature which transects northwest Tasmania from the west coast to the north coast (fig. 2). The assemblage of rocks which form it is known as the Arthur Metamorphic Complex, which includes pelitic, dolomitic and quartzose schist with interbanded amphibolite derived from tholeiitic volcanic and related intrusive rocks (Turner and Bottrill, 1993). Metamorphic grades in the complex are typically greenschist facies, but relict blueschist and amphibolite facies mineralogies are preserved in metabasic rocks in the Bowry Formation. In its current geometry, the Bowry Formation forms part of a fault-bounded central unit of the Arthur Metamorphic Complex, flanked to the west by

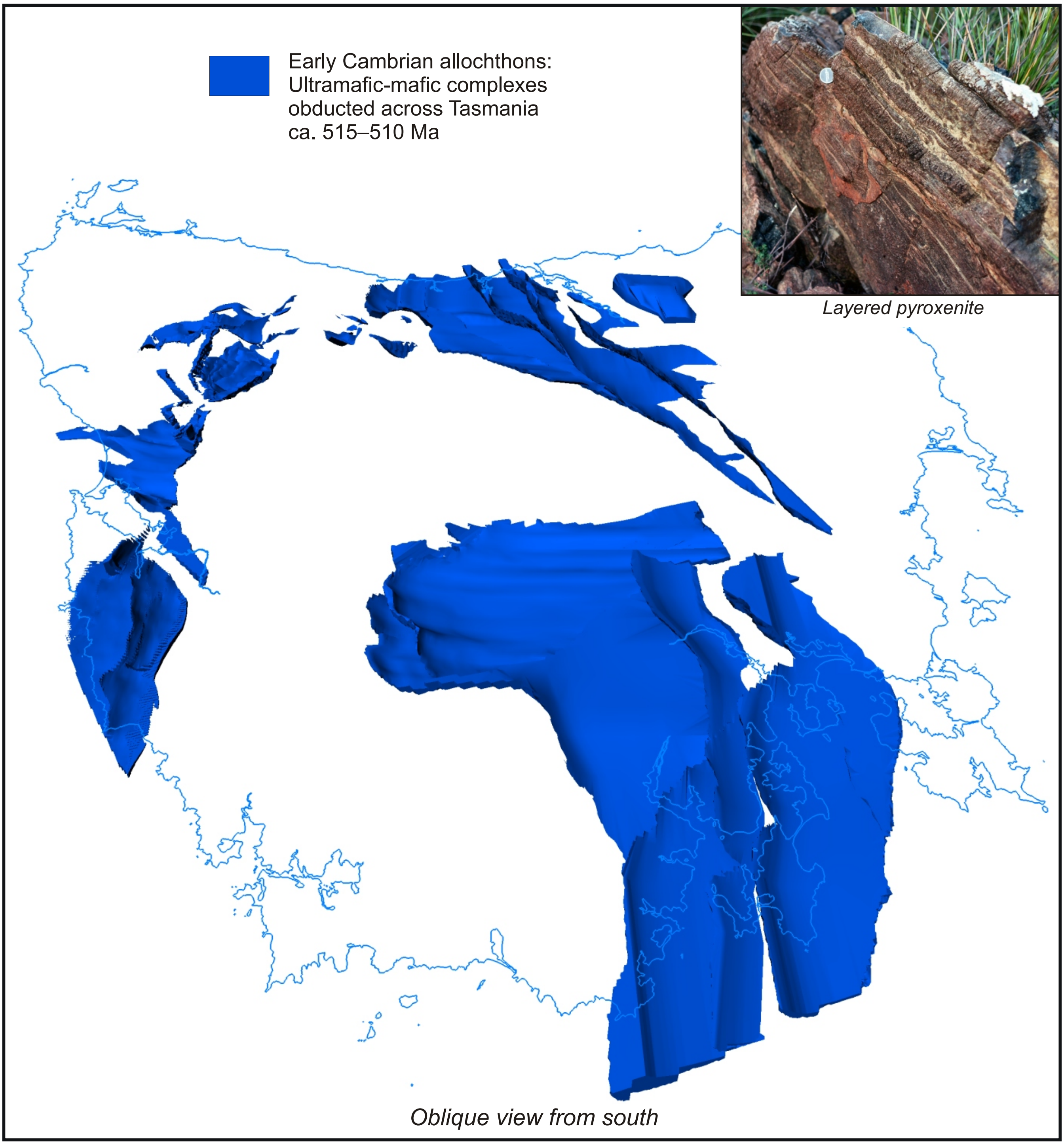



**Figure 4**

*3-D geological model showing extent of outcropping and concealed allochthonous Early Cambrian ultramafic-mafic complexes on mainland Tasmania.*

**Table 1**

*Non-alluvial deposits of Tasmania: pre-mining resources*

*Note: These data rely on past production figures and various resource estimations. Some of the latter do not comply with the Joint Ore Reserves Committee Code standards.*

| | |
|---|---|
| **Precambrian deposits in the Arthur Lineament** | |
| Savage River | 371 Mt @ 31.9% Fe |
| Arthur River | 29 Mt @ 42.8% Mg |
| Main Creek | 42.8 Mt @ 42.4% Mg |
| **Cambrian gabbro-hosted deposits** | |
| Nickel Reward (Cuni field) | 0.03 Mt @ 3% Ni |
| North Cuni–Genets Winze | 0.95 Mt @ 0.76% Ni, 0.94% Cu |
| **Cambrian deposits in the Mount Read Volcanics** | |
| Hellyer | 16.5 Mt @ 13.9% Zn, 7.2% Pb, 0.38% Cu, 169 g/t Ag, 2.55 g/t Au |
| Que River | 3.3 Mt @ 13.3% Zn, 7.4% Pb, 0.7% Cu, 195 g/t Ag, 3.3 g/t Au |
| Mount Charter | 13.0 Mt @ 0.6% Zn, 0.3% Pb, 16 g/t Ag, 0.9 g/t Au |
| Rosebery | 32.7 Mt @ 14.5% Zn, 4.4% Pb, 0.58% Cu, 145 g/t Ag, 2.2 g/t Au |
| Hercules | 3.33 Mt @ 17.3% Zn, 5.5% Pb, 0.4% Cu, 171 g/t Ag, 2.8 g/t Au |
| South Hercules | 0.56 Mt @ 3.7% Zn, 1.9% Pb, 0.1% Cu, 157 g/t Ag, 3.0 g/t Au |
| Mount Lyell | 311 Mt @ 0.97% Cu, 0.31 g/t Au |
| Henty–Mt Julia | 2.83 Mt @ 12.5 g/t Au |
| **Ordovician carbonate-hosted deposits** | |
| Oceana | 2.6 Mt @ 7.7% Pb, 2.5% Zn, 55 g/t Ag |
| Grieves Siding | ~0.7 Mt @ 8% Zn (primary); 0.15 Mt @ 5% Zn (secondary) |
| **Devonian orogenic gold deposits** | |
| Beaconsfield | 3.25 Mt @ 19.0 g/t Au |
| New Golden Gate | 0.51 Mt @ 15.6 g/t Au |
| Pinafore Reef (Lefroy) | 0.358 Mt @ 23.2 g/t Au |
| **Devonian granite-related tin deposits** | |
| ***Sulfide skarns*** | |
| Renison Bell | 24.54 Mt @ 1.41% Sn |
| Mount Bischoff | 10.54 Mt @ 1.1% Sn |
| Cleveland | 12.4 Mt @ 0.61% Sn, 0.25% Cu |
| Foley zone | 3.8 Mt @ 0.28% $WO_3$, 0.28% $MoS_2$, 0.05% Sn |
| Razorback | 0.34 Mt @ 0.9% Sn |
| Queen Hill | 3.6 Mt @ 1.2% Sn |
| ***Silicate skarns*** | |
| St Dizier | ~2.6 Mt @ 0.5% Sn, 0.05% $WO_3$ |
| ***Vein deposits*** | |
| Aberfoyle | 2.1 Mt @ 0.91% Sn, 0.28% $WO_3$ |
| Pieman vein (East Renison) | 0.43 Mt @ 1.0% Sn |
| ***Greisen deposits*** | |
| Anchor | 2.39 Mt @ 0.28% Sn |
| **Other Devonian granite-related deposits** | |
| ***Skarns*** | |
| King Island field | 17 Mt @ 0.85% $WO_3$ |
| Kara | 5.2 Mt @ >30% Fe, by-product $WO_3$ |
| Avebury | 11.59 Mt @ 1.02% Ni |
| West Viking (Avebury) | 1.4 Mt @ 1.3% Ni |
| Moina | 18 Mt @ 26% $CaF_2$, 0.1% Sn, 0.1% $WO_3$ |
| Hugo (Moina area) | 0.25 Mt @ 5.5% Zn, 1 g/t Au, 0.1% Bi |
| Stormont (Moina area) | 0.135 Mt @ 3.44 g/t Au, 0.21% Bi |
| ***Vein deposits*** | |
| Storeys Creek | 1.1 Mt @ 1.09% $WO_3$, 0.18% Sn |
| Magnet | 0.63 Mt @ 7.3% Zn, 7.3% Pb, 427 g/t Ag |
| Salmons Vein (East Renison) | 0.83 Mt @ 3.2% Pb, 2.2% Zn, 104 g/t Ag, 0.19% Sn, 0.61% Cu |
| New North Mount Farrell and North Mount Farrell | 0.908 Mt @ 12.5% Pb, 2.5% Zn, 408 g/t Ag |
| Lakeside | 0.75 Mt @ 0.2% Sn, 0.2% Cu, 4.0% As, 2.1 g/t Au, 20 g/t Ag |

metamorphosed equivalents of the Rocky Cape Group and Ahrberg Group (a correlate of the Togari Group), and to the east by a metamorphic equivalent of the Oonah Formation. These flanking units are transitional to their respective, relatively unmetamorphosed parent units with distance away from the axis of the lineament. The Arthur Lineament hosts economically important iron ore, magnesite and silica flour deposits (Table 1), and is considered to represent the current western limit of allochthonous units emplaced in the Tyennan Orogeny.

The Savage River magnetite mine, within the Bowry Formation, consists of sub-vertical, concordant lenses of massive magnetite ore with varying amounts of pyrite and trace chalcopyrite in a sequence of tholeiitic amphibolite of extrusive and intrusive origin, carbonates and serpentinite. This is enclosed by quartzite and carbonaceous schist which include carbonate units (dolomite and magnesite; Frost and Matzat, 1984) and minor amphibolite on the eastern side. The silicate gangue to the ore is dominated by either antigorite ± talc or tremolite-actinolite ± chlorite (Coleman, 1976; Spiller, 1974). Serpentinites have low Cr and Ni contents and it has been suggested that they may be metasomatised siliceous dolomite (Spiller, 1984). Other magnetite-rich iron formations lie to the south. The largest of these, the Long Plains deposit, is currently being explored. The ores were emplaced either before or early in the deformational history of the area. There has been no definitive genetic study of the deposits and a range of suggestions for the mode of formation includes volcanic exhalative (Coleman, 1976), magmatic segregation (Matzat, 1984) and hydrothermal replacement (Bottrill *et al*., 1998).

Magnesite deposits of high purity and substantial size are known at Arthur River, Lyons River, north of Savage River and Main Creek to the south, and there is a significant body to the east of the Northern deposit at Savage River. Frost (1982) considered the Main Creek body to be a metasomatic replacement of dolomite, but oxygen and carbon isotope analyses of the Savage River deposit and magnesite at Long Plains indicate a diagenetic origin, probably in a hypersaline environment (Matzat, 1984), and it is likely that all of the deposits share a common origin.

## Middle Cambrian post-collisional phase: Mt Read Volcanics

The most important metallogenic event in Tasmania coincided with the deposition of the Mount Read Volcanics (MRV). Various U-Pb zircon ages and numerous fossil localities constrain the bulk of the MRV to a narrow time range from early Middle Cambrian to early Late Cambrian. Generally the precision of the dating is insufficient to be useful in correlation, which relies on regional geological interpretation and relatively few fossil sites. Consequently aspects of MRV geology remain controversial.

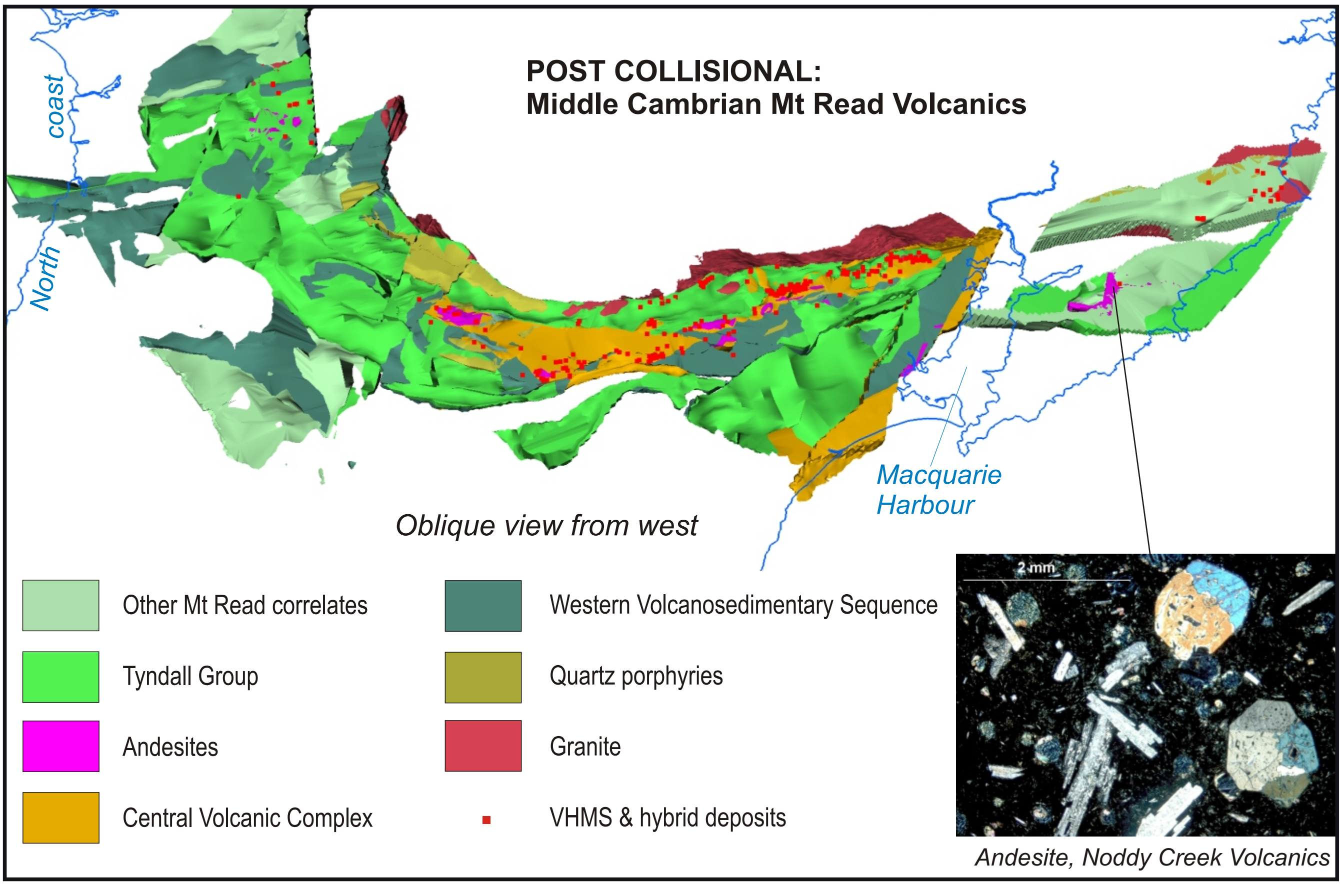



**Figure 5**

*3-D geological model showing distribution of the main component rock sequences of the post-collisional, Middle Cambrian Mt Read Volcanics.*

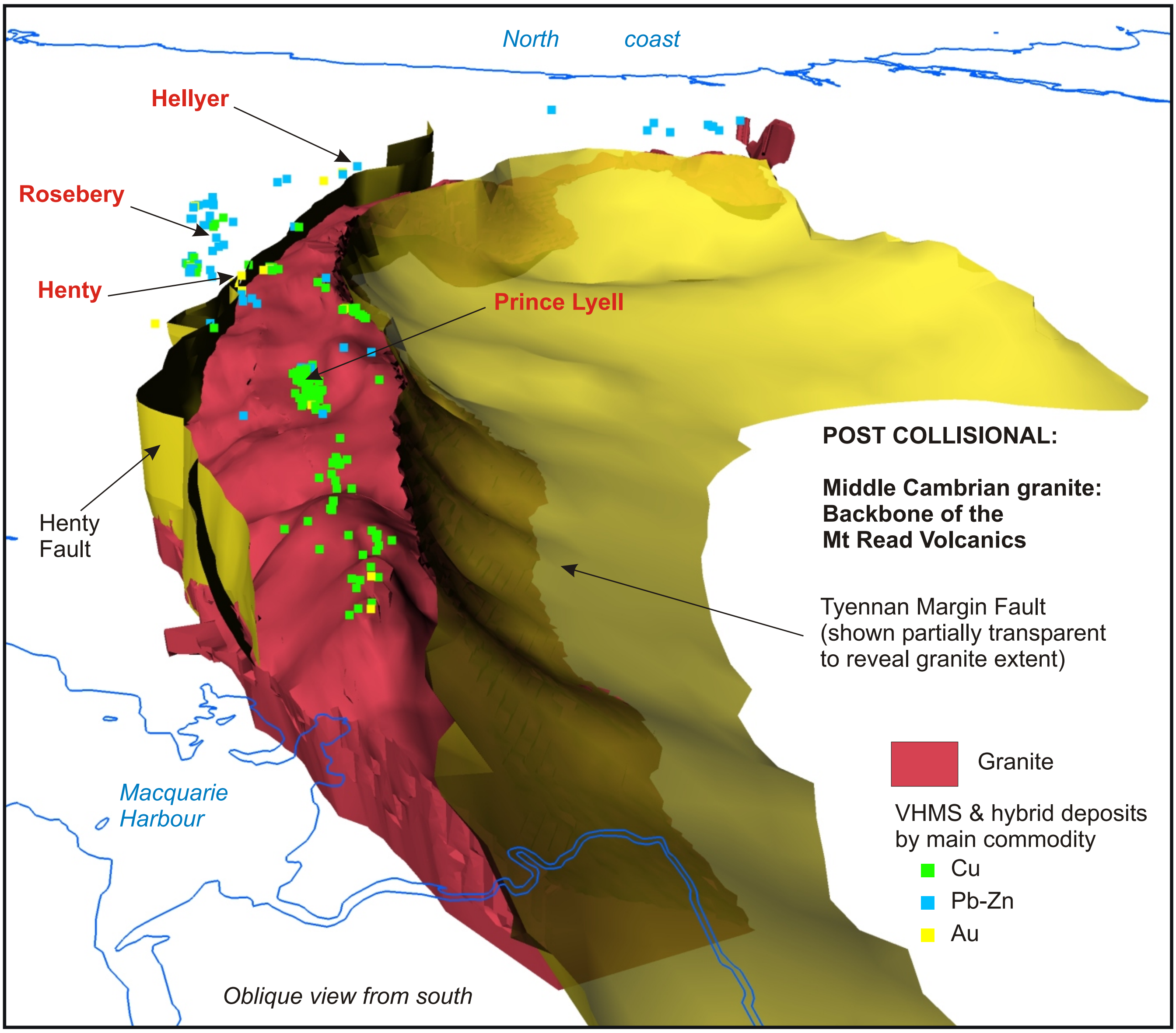



**Figure 6**

*3-D geological model of the Middle Cambrian granite spine beneath the Mt Read Volcanics.*

The main mineralised belt of the MRV between Mount Darwin and Hellyer is the Central Volcanic Complex (CVC; fig. 5), which is dominated by proximal volcanic rocks (rhyolite and dacite flows, domes and cryptodomes and massive pumice breccias) and andesite and rare basalt (lavas, hyaloclastites and intrusive rocks) deposited in a marine environment (Corbett, 1989; 1992; 2002; Gifkins and Kimber, 2003). This belt is flanked to the west by the coeval Western Volcano-Sedimentary Sequence (WVS; Corbett, 2002) of lithicwacke turbidite, mudstone (commonly rich in shards), siltstone, shale and subordinate intrusive rocks and lavas, commonly andesitic.

These rocks are overlain by the Tyndall Group, a unit of quartz-bearing volcaniclastic sandstone and conglomerate of mixed felsic and andesitic provenance, with the latter common towards the base, and minor felsic and andesitic lavas and intrusive rocks and welded ignimbrite (White and McPhie, 1996). Considerable erosion took place locally before deposition of the Tyndall Group. Clasts of granite and altered volcanic rocks occur in the basal Tyndall Group in the Mount Darwin area (Corbett, 2002; Morrison, 2002).

Flanking the CVC to the east and abutting the metasedimentary rocks of the Tyennan region is the Eastern Quartz-phyric Sequence (EQPS), which consists of quartz-feldspar-phyric lavas, intrusive porphyries and volcaniclastic sandstone, intruded by magnetite series granites (fig. 6). The basal unit of the EQPS consists of Precambrian-derived sandstone and conglomerate which passes upward gradationally into volcaniclastic sandstone. There is some controversy about the EQPS. Corbett (2002) considers that it is a time equivalent of the CVC, while Murphy *et al*. (2004) consider that it could be part of the Tyndall Group.

Tectonism was mostly near east-west extensional during Mount Read Volcanics deposition, as recorded by the orientation of hydrothermal veins (e.g. at the Hellyer deposit; Gemmell and Large, 1992) and basaltic

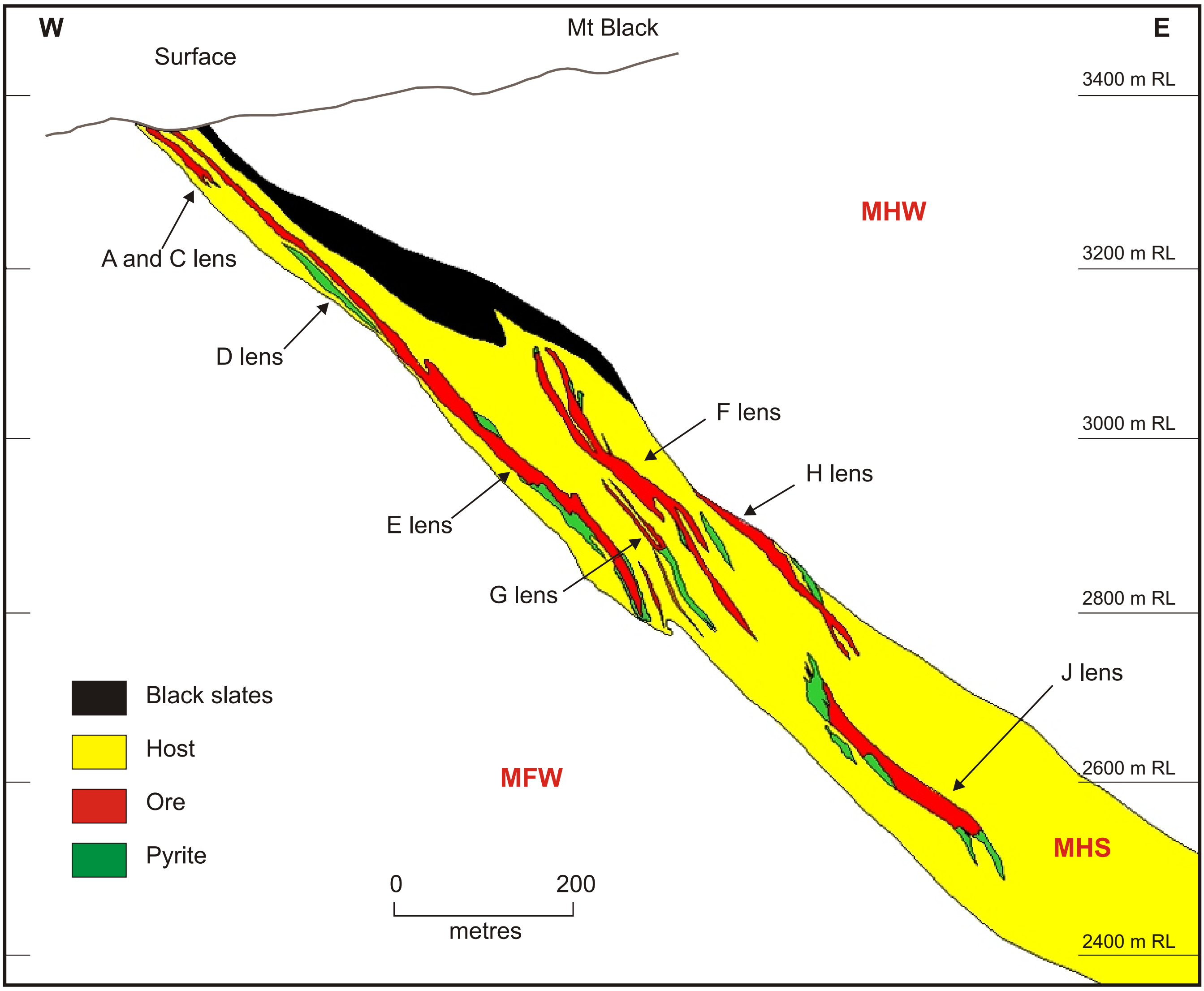



**Figure 7**

*Generalised cross section through the southern section of the Rosebery mine (courtesy A. W. McNeill, Zinifex Limited). Abbreviations — MFW: mine footwall sequence (pumiceous rhyolitic to dacitic mass flow deposits, commonly altered to quartz-sericite ± chlorite ± pyrite beneath the ore; MHS: mine host sequence (bedded volcaniclastic sandstone and siltstone, generally strongly altered to sericite ± quartz ± carbonate ± chlorite ± pyrite near ore); MHW: mine hangingwall sequence (commonly rhyolitic to dacitic volcaniclastic breccia). All units were emplaced in a subaqueous environment below wave base.*

dykes in the Henty Fault Zone. Further evidence of an extensional regime is provided by growth faulting and cauldron subsidence associated with the formation of the thick pumice-rich breccia underlying the Rosebery deposit (Green *et al*., 1981; Gifkins and Kimber, 2003). Solomon (*in* Solomon and Groves, 2000) has suggested that this episode of extensional tectonism and crustal thinning was related to eastward retreat of a west-dipping subduction zone following the collisional event.

Mineralisation was concentrated in a short time interval in the late Middle Cambrian (fossil ages summarised in Jago and Brown, 1989) at the top of the CVC and in places in the immediately overlying Tyndall Group rocks. Major alteration zones are dominantly of quartz-sericite mineralogy and restricted to the proximal CVC volcanic facies (Gifkins and Kimber, 2003; Herrmann and Kimber, 2003), but chlorite-rich cores are apparent at Hellyer (Gemmell and Large, 1992) and Hercules (Green and Taheri, 1992). Despite this, the WVS in areas of andesite proximally underlying Tyndall Group correlates offers under-explored targets (Corbett, 2002).

The Henty Fault Zone constitutes a fundamental metallogenic divide within the MRV. To the northeast, polymetallic Zn-Pb-Au-Ag-Cu massive sulfide deposits dominate (Hellyer, Que River, Rosebery, Hercules; fig. 7) together with disseminated deposits with lower base metal, but relatively high Au and Ag tenor (Mount Charter, South Hercules). Ore fluids have been considered to have been derived by convective circulation of seawater and interaction with volcanic and basement rocks, but there is evidence for a magmatic contribution to the ore fluids at Hellyer (Solomon and Groves, 2000). There is also debate about the extent to which massive sulfide ore deposition took

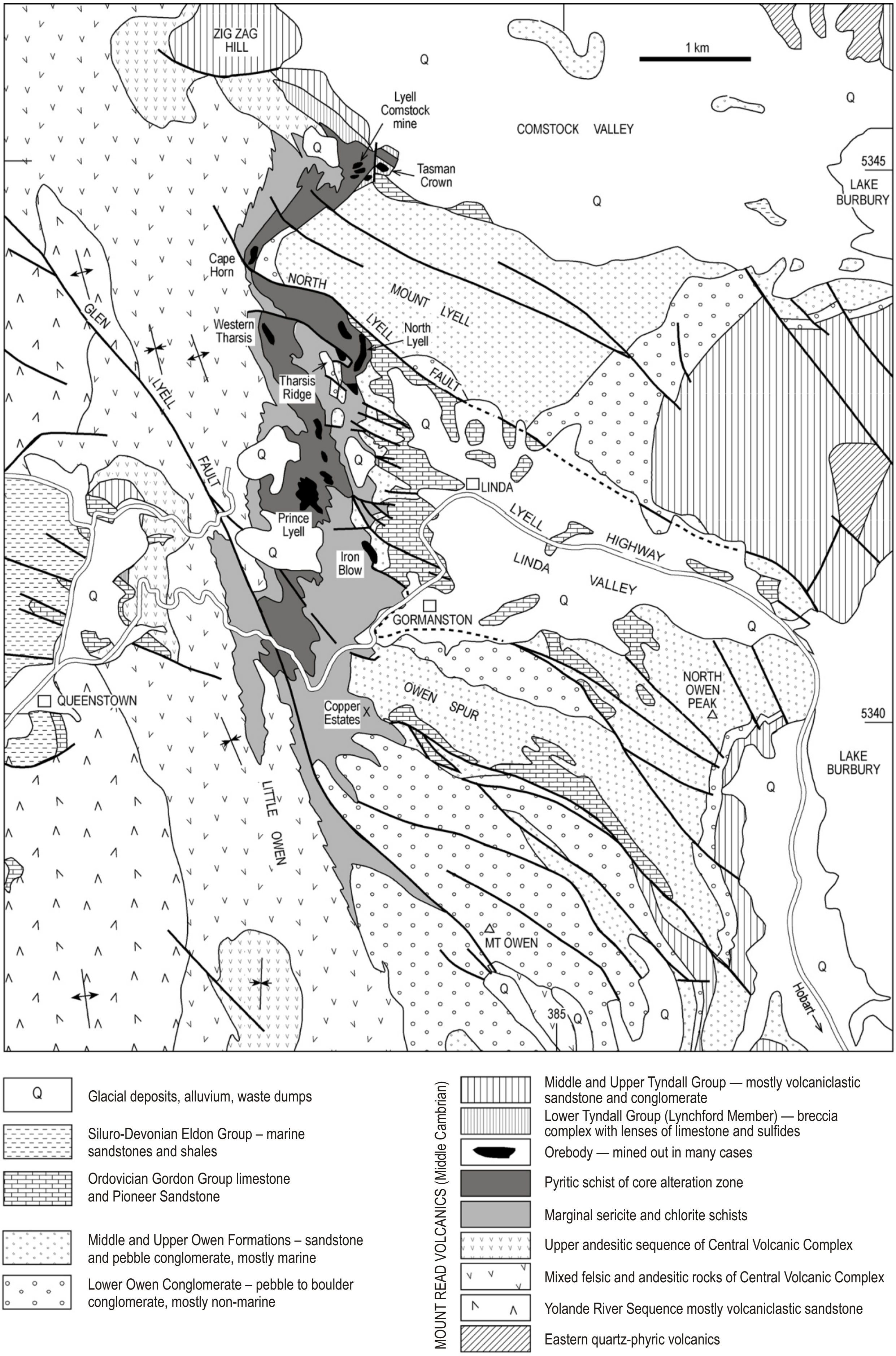



**Figure 8**

*Geological map of the Mount Lyell area (after Corbett, 2001).*

place in brine pools on the seafloor (Solomon and Groves, 2000), as opposed to in either Kuroko-type hydrothermally reworked seafloor mounds (e.g. Gemmell and Large, 1992) or sub-seafloor replacement or displacement (Allen, 1995). There is clear evidence that the sub-seafloor ore formation operated locally, for example at the Hercules deposit. The role of shifting loci of upward hydrothermal fluid flow in affecting the development of the Rosebery deposit, for example, has been argued (Green *et al*., 1981; Green, 1983).

Southeast of the Henty Fault copper-gold and gold deposits dominate, exemplified by the Mount Lyell field and the Henty gold deposit. The most economically important deposits in the Mount Lyell field (fig. 8) are disseminated chalcopyrite-pyrite orebodies in alteration assemblages dominated by quartz-sericite or quartz-chlorite-sericite (Prince Lyell, Cape Horn, Lyell Comstock in part, Western Tharsis; Walshe and Solomon, 1981). The Western Tharsis deposit consists of concentrically and vertically zoned alteration assemblages centred on the mineralisation, with a lower and central quartz-chlorite-sericite zone passing upward and outward in turn into pyritic quartz-pyrophyllite ± topaz ± fluorite ± zunyite ± woodhouseite with local bornite-bearing zones and a pyritic quartz-sericite assemblage (Huston and Kamprad, 2001). The highest grade ores are in the North Lyell area and consist of coarse-grained bornite with chalcopyrite and minor pyrite in brecciated chert ± pyrophyllite ± barite ± hematite or quartz-sericite schist at or near the contact between the altered CVC and the Owen Conglomerate, which may be locally hematised. The origin of these high-grade ores has been a matter of conjecture. Solomon *et al.* (1987) suggested that they may have formed during Devonian deformation by mixing of metamorphic fluids from the volcanic rocks and conglomerate, but most current opinion is that the ores of the field formed at the same time during late CVC or early Tyndall Group time (Corbett, 2001; Huston and Kamprad, 2001). The first deposit mined, the Mount Lyell or Iron Blow deposit, consisted mostly of massive pyrite-chalcopyrite. There are also small lenses of polymetallic pyritic sphalerite-galena rich massive sulfide in the Lyell Comstock area at the northern end of the field.

There are other disseminated copper deposits in the CVC further south. Some, such as the Garfield prospect, appear geologically similar to Prince Lyell. The Henty gold mine consists of a series of small high-grade lenses of gold mineralisation in quartz ± sericite-altered volcaniclastic and volcanic rocks that occupy a large sub-vertical quartz-sericite alteration zone that transects the CVC-Tyndall Group contact at a low angle. The deposit is regarded as a submarine epithermal system formed from a magmatic fluid (Callaghan, 2001). Current opinion also favours a significant or dominant magmatic component in the fluids that formed the Mount Lyell deposits (Large *et al*., 1996; Huston and Kamprad, 2001), with the fluids derived from Cambrian granite. The three dimensional geological model of Tasmania provides some empirical support for this, in that granite is interpreted to shallowly underlie the copper-gold metallogenic region east of the Henty Fault, but is considered to be at far greater depths to the west (fig. 6).

As part of the 3-D geological model development, the mineral potential was examined for a range of deposit styles in the context of the regional structural architecture (Murphy *et al*., 2004). Prospectivity analysis was empirically based on the use of potential field worms for determining the relative shape, depth extent and continuity of edges (e.g. faults, intrusive boundaries). This approach rests on the association between mineral deposits and geological structures.

The term 'worm' derives from an automated edge detection process that defines 3D arrays of maximum gradient points over a range of scales of upward continuation. The process relies on gradients in density or magnetic susceptibility, but structures may also be recognised by truncations or vacancies in the worm populations. For the central MRV, there is a clear association between the major deposits and the longer potential field worms (fig. 9).

Positive evidence of undiscovered VHMS deposits in western Tasmania exists in the form of debris flow deposits with rafts and clasts of high-grade ore.

## Wurawina Supergroup: Late Cambrian to Early Devonian

In the Late Cambrian, the final phase of the Tyennan Orogeny inverted earlier extensional faults (e.g. Henty Fault). Major reverse faults and upright open north-trending folds were formed in western Tasmania. This phase also caused uplift of the Tyennan region, with syn-orogenic sediments (Owen Group) accumulating in synclinal cores and other structural depressions (fig. 10). The Owen Group commonly rests with angular and/or erosional unconformity on older units. It typically includes large volumes of coarse siliciclastic conglomerate composed dominantly of metaquartzite clasts derived from the Tyennan Proterozoic sequences (fig. 2), but also includes turbidite and shallow marine sandstone units.

In a recent redefinition, an Ordovician pebble conglomerate and sandstone unit (Pioneer Sandstone), previously an upper unit of the Owen Group, has been promoted into the overlying Ordovician Gordon Group (Corbett, 2001; Noll and Hall, 2005). Because of Late Cambrian movements, the Pioneer Sandstone rests unconformably on older units of the Owen sequence in much of the Dundas Trough, west of the Machinery Creek Fault, a fundamental structure in central northern Tasmania. Thus, the relationship between the Owen and Gordon groups is largely unconformable west of the Machinery Creek Fault, and transitional and conformable east of it. The Gordon Group (fig. 10) above the Pioneer Sandstone is a shallow-marine to peritidal, platform succession of predominantly micritic, dolomitic limestone, up to

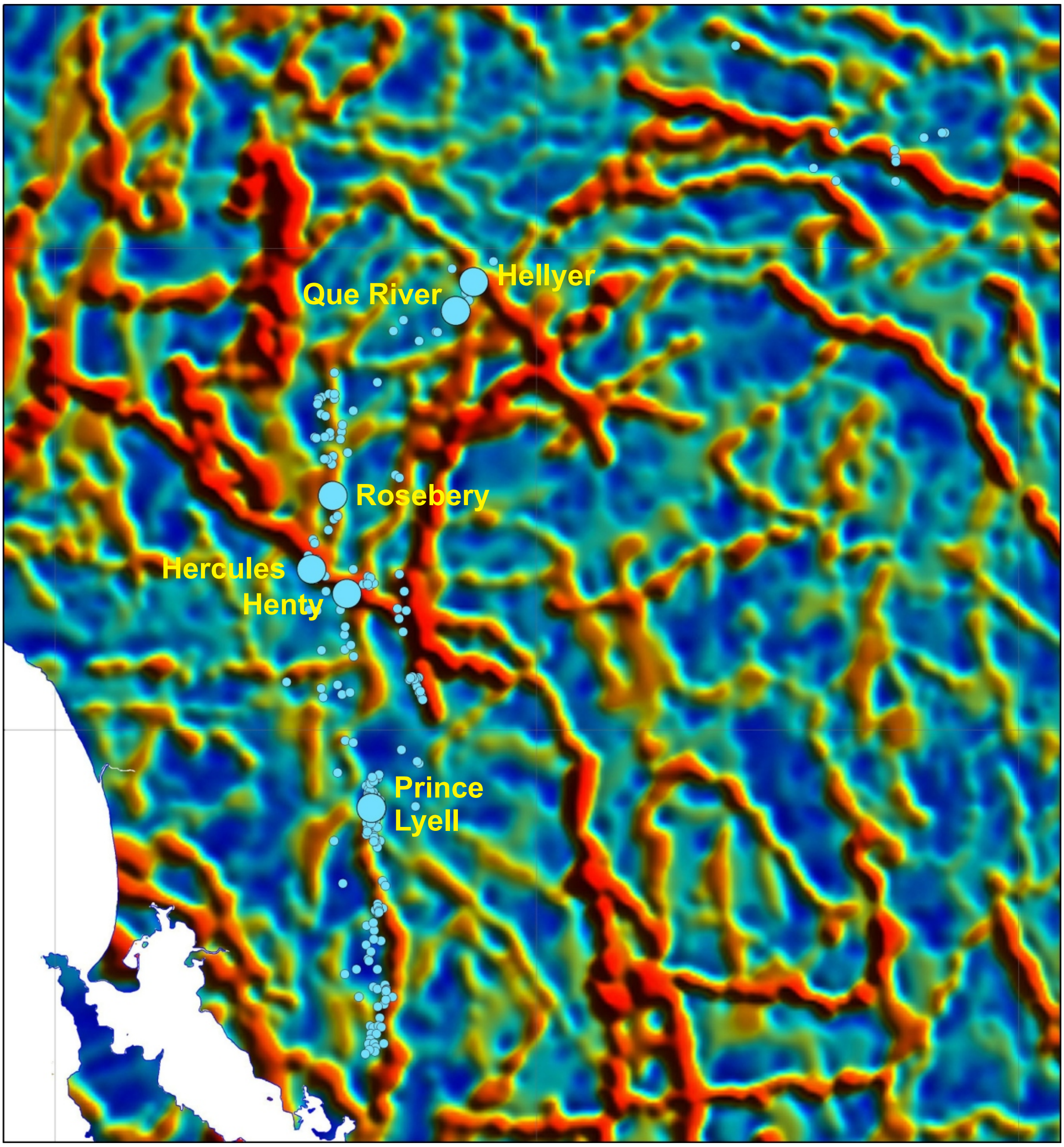



**Figure 9**

*Total (i.e. combined gravity and magnetic) Worm Length image for central western Tasmania, with locations of Middle Cambrian VHMS and hybrid mineral deposits overlain.*

1.8 km thick in central–southern Tasmania but considerably thinner in western Tasmania. The onset of carbonate sedimentation took place in the Middle Ordovician in western and northern Tasmania but was earlier (Early Ordovician) in the east (Banks and Burrett, 1980; Banks and Baillie, 1989; Laurie, 1995). Stratiform sulfide mineralisation and an associated breccia unit in the Zeehan area indicate local synsedimentary faulting and possible exhalative activity (Taylor and Mathison, 1990). An abrupt lateral transition into deep-water limestone in the far south is the only known indication of a platform edge (Burrett *et al.*, 1984). Gordon Group carbonate sequences became an important ore host for skarn mineralisation associated with intrusion of Late Devonian–Early Carboniferous granitoids. High purity limestone is mined at Mole Creek for metallurgical use and limestone from Railton and elsewhere is utilised for cement manufacture and other purposes.

The Silurian to Early Devonian Eldon Group is locally disconformable and erosional on the Gordon Group in western Tasmania but elsewhere the contact is conformable and transitional. Up to five kilometres thickness of Eldon Group rocks are preserved in the axial parts of major Devonian synclines (Banks and Baillie, 1989). The lower part of the succession is dominated by shallow-marine quartz sandstone (Crotty and Florence formations and correlatives); the upper by a thick, shelf-facies shale unit with minor limestone (Bell Shale and correlatives). The youngest known horizon in the Bell Shale is middle Early Devonian (Banks and Baillie, 1989).

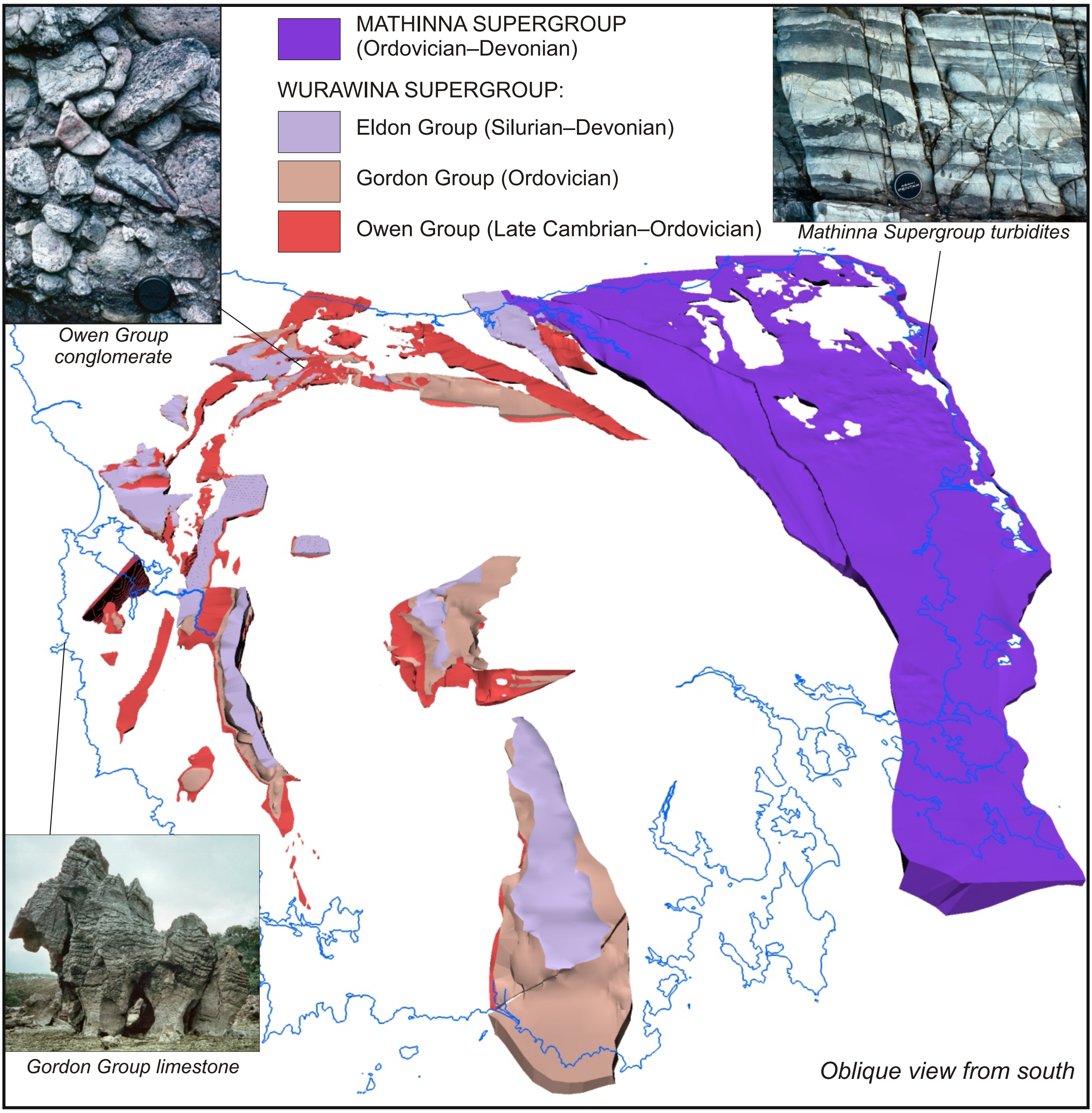



**Figure 10**

*3-D geological model showing distribution of the Wurawina Supergroup (western Tasmania) and the Mathinna Supergroup (eastern Tasmania).*

## Mathinna Supergroup: Ordovician to Early Devonian

Deposition of the Mathinna Supergroup in eastern Tasmania was approximately coeval with deposition of the Wurawina Supergroup between the uppermost parts of the Owen Group and the top of the Eldon Group in western Tasmania. The Mathinna Supergroup comprises a succession of turbiditic sandstone and mudstone and forms the pre-Carboniferous sedimentary basement to eastern Tasmania, from the Furneaux Group of islands in Bass Strait to the northeastern coast of the Tasman Peninsula in southeastern Tasmania (fig. 10). The proposed elevation to supergroup status is relatively recent (Reed, 2001), and based on an interpretation that the oldest deformation event affecting the sequence was Late Ordovician to Early Silurian (Benambran) in age and only affected the two lowermost formations, these being separated from the younger part of the sequence by an inferred unconformity. This interpretation is still controversial.

The base of the succession is not known. The oldest formation consists of a quartzose turbiditic sandstone succession ~1 km thick (Stony Head Sandstone) overlain by a 1–2 km thick black pelite (Turquoise Bluff Slate) containing an Early Ordovician graptolite (Banks and Smith, 1968). In the model of Reed (2001), these two formations are unconformably overlain by the rest of the succession, comprised of about a two kilometre thick succession of quartzose turbidites of sublithic composition (Bellingham Formation), followed by about two kilometres of feldspathic turbidites (Sidling Sandstone; Powell *et al.*, 1993). This part of the sequence contains Late Silurian graptolites (Rickards *et al.*, 1993) and Early Devonian shelly fossils (Rickards and Banks, 1979). There is a regional younging of the supergroup from west to east based on the sparse fossil data.

Palaeocurrent data indicate that in the Silurian to Early Devonian the Mathinna Supergroup was deposited in a basin elongated NNW-SSE and with a quartzose, cratonic source to the southwest (Banks and Baillie, 1989), but detrital zircon dating indicates this source was not the western Tasmanian terrane (Black *et al.*, 2004). Current models suggest that northeastern Tasmania may have been substantially separated from the rest of Tasmania at the time of Mathinna Supergroup deposition, and that it was finally docked with western Tasmania during Middle Devonian orogenesis.

## Middle Devonian orogenesis, granite emplacement and mineralisation

In the Middle Devonian, most of Tasmania was affected by polyphase deformation, characterised by a complexity of fold orientations, due in part to reactivation of older structures (fig. 11). The fold geometry was commonly controlled by the trends of Cambrian folds which were tightened during the Devonian, and as a result, in places Devonian cleavage orientations transect the axial planes of associated folds.

The folding occurred in two main phases in western Tasmania. The early phase produced NNW-trending folds in areas where reactivation effects were not significant, but also tightened north-trending Cambrian folds (which became transected in some cases by NNW-trending Devonian cleavage), and produced NE-trending folds in the vicinity of the Henty Fault due to reactivation of that structure (fig. 11). The subsequent second main phase of compression produced NW to WNW-trending folds and thrusts, with regionally compartmentalised strain, the strongest effects being in central northern Tasmania and in the Linda Zone which passes through Queenstown and transects the Tyennan region (fig. 11). East-west trending Devonian folds in the central north were probably generated by reactivation of similarly trending Cambrian folds in the early stages of this second Devonian event. The age of Devonian deformation in western Tasmania is constrained by the age of the youngest rocks affected (the fossiliferous upper Eldon Group, of Early Devonian age; Banks and Baillie, 1989) and by undisturbed, spore-rich, upper Middle Devonian terrestrial cavern fillings at Eugenana, which contain disoriented blocks of deformed Gordon Group limestone that collapsed from the cavern walls (Balme, 1960; Burns, 1964).

The Mathinna Supergroup of northeastern Tasmania also shows evidence of at least two, and perhaps three, pre-Carboniferous compressional deformation events, but whether these are all Devonian remains controversial. In contrast to western Tasmania, the fold generations are essentially coaxial (and of generally northwest trend). Currently there are two main competing models. Patison *et al.* (2001) produced a relatively thin-skinned tectonic model, in which an eastward-tapering tectonic wedge was subsequently exhumed by westward back-thrusting, yielding a total crustal thickening of about ten kilometres. In a two-phase Devonian structural history, $D_1$ (pre-388 Ma) involved east-directed thrusting and tectonic thickening, and a foreland-propagating thrust wedge. Peak metamorphism (200–300°C, sub-greenschist facies) occurred after the east-directed thrusting as a result of crustal thickening. The west-vergent $D_2$ event (ca. 390 Ma; Black *et al.*, 2005) refolded $D_1$ structures and created a back-thrust zone in the Beaconsfield area near the boundary of the Eastern Tasmanian and Western Tasmanian terranes (fig. 11). The alternative model of Reed (2001) proposed that the $D_1$ event represents a Tasmanian effect of the Benambran Orogeny (455–424 Ma) of mainland Australia, and produced east-vergent recumbent folds and thrusts confined to the western part of the terrane (fig. 11). This was followed by two Devonian deformation events, the last of which is essentially the same as $D_2$ of Patison *et al.* (2001). Problems with the Reed model include the recent discovery of early east-directed thrusts in the eastern parts of the terrane, but the jury may still be out on the general question of Benambran effects in Tasmania.

Upper correlates of the Owen Group in the Beaconsfield area and the Mathinna Supergroup further east host economically important orogenic vein-style gold mineralisation, which is localised on structures associated with the last main Devonian deformation event ($D_2$ of Patison *et al.*, 2001; $D_3$ of Reed, 2001). In the western part of the Eastern Tasmanian terrane, notably in the Beaconsfield and Lefroy districts (fig. 12), steeply-dipping reefs have an east to ENE orientation and formed near-parallel to the axis of maximum principal stress (Powell, 1991; Reed, 2002), whereas to the east along the Mangana–Mathinna–Waterhouse gold lineament and elsewhere, the dominant orientation is NNW and orthogonal to this axis. The latter structures form as a response to failure on the steep eastern limbs of anticlines formed during the earlier fold event. In detail, gold-rich shoots within the reefs tend to pitch steeply, as a response to either favourable lithology for reef formation (Beaconsfield) or due to gold mineralisation during late transcurrent movement on the structures (Mathinna; Keele, 1994).

An extended period of large-scale intrusion of granitoids at relatively high crustal levels commenced in eastern Tasmania at ca. 400 Ma with the emplacement of unfractionated I-type granodiorites

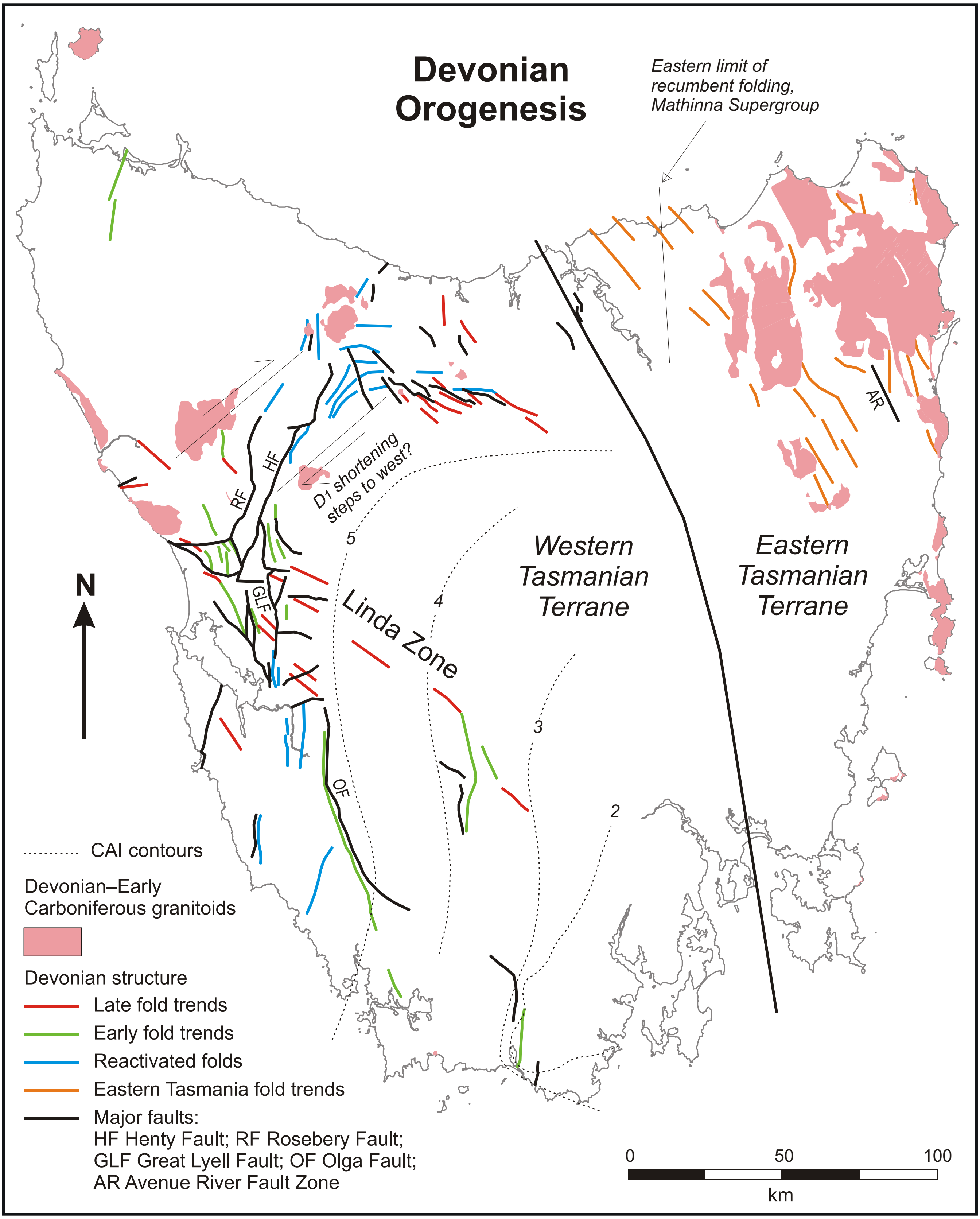



**Figure 11**

*Devonian structural trends in Tasmania, modified and updated from Stacey and Berry (2004) and Williams et al. (1989). Contours of Conodont Alteration Index (CAI) from Burrett (1992) shown for the Western Tasmanian terrane.*

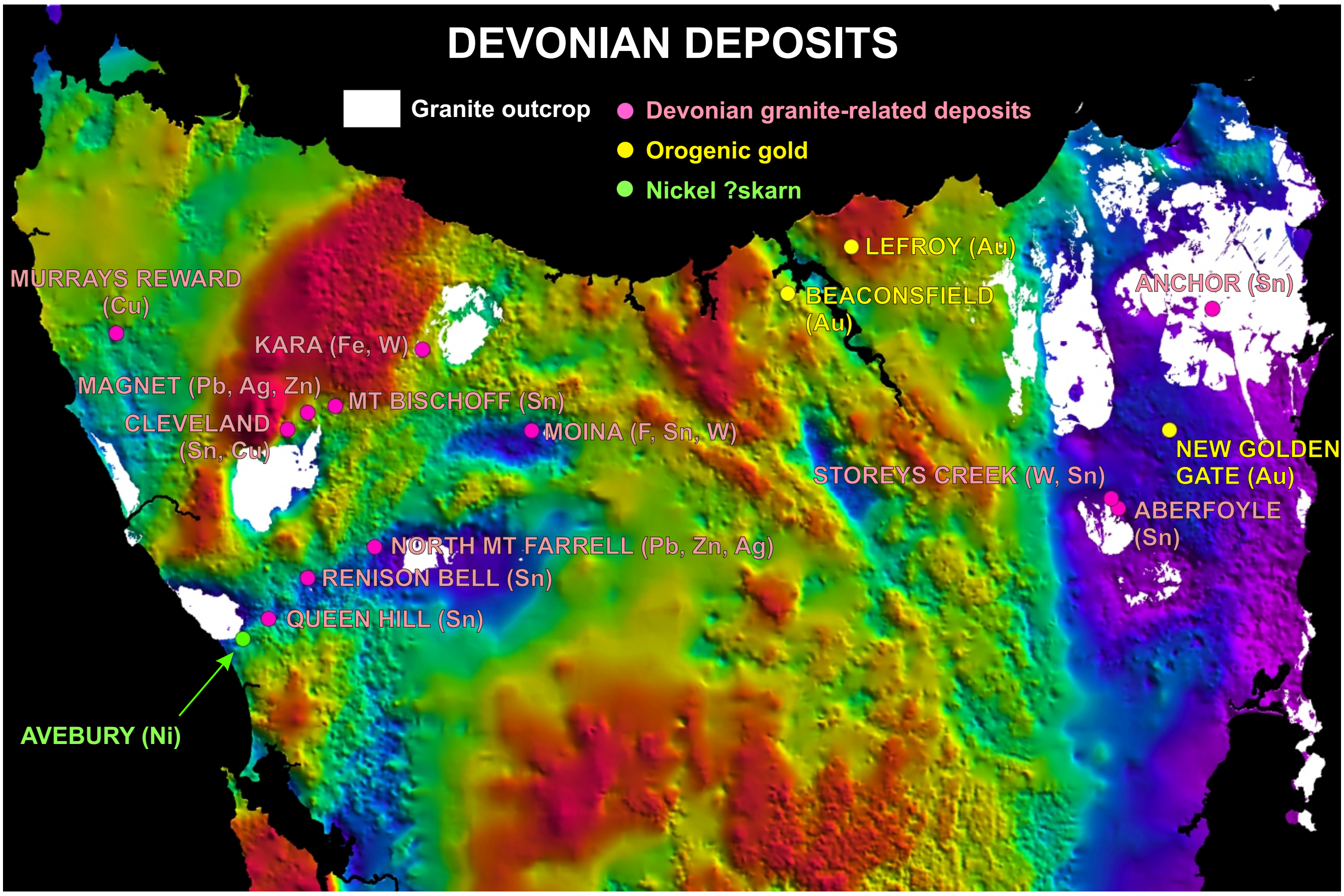



**Figure 12**

*Granitoids and major Devonian mineral deposits of Tasmania shown on an image of terrain-corrected gravity.*

prior to the Devonian deformation events, and continued after the close of Devonian deformation, with the youngest granitoid at ca. 350 Ma on King Island. Most of the western Tasmanian granitoids post-date Devonian folding events. On mainland Tasmania the granitoids form three large complexes; one in the east, another in the northwest, and a third largely concealed beneath the southwest (fig. 13). Crystallisation age data show a statistical westward younging of the granitoids regionally across Tasmania, and there is also a statistical compositional trend towards felsic, fractionated I-type and S-type granite and monzogranite with decreasing age (Black *et al.*, 2005).

All significant Devonian granite-related deposits lie within the four kilometre granite isobath (fig. 13). Contrast in the host rocks in western and northeastern Tasmania is responsible for fundamental differences in the Late Devonian to Early Carboniferous metallogenesis of the regions. There is a more restricted range of deposits in the northeast due mostly to a lack of reactive host lithologies, and perhaps also to a greater degree of unroofing (Solomon and Groves, 2000). There is current exploration interest in intrusion-related gold that occurs as part of the Au-As-Bi-Mo association within, and in the aureoles of, unfractionated I-type moderately oxidised to moderately reduced granodiorite in the Lisle–Golconda area. Significant mineral deposits (fig. 12) are cassiterite and wolframite-bearing vein deposits (e.g. Aberfoyle, Storeys Creek) and cassiterite-bearing greisen deposits associated with post-tectonic reduced, strongly fractionated, S-type granite (e.g. Anchor, Royal George). Base metal vein deposits are known but are insignificant. Kaolinite is found in the haloes of greisen tin deposits and within alluvial tin deposits. Kaolinite of debatable hydrothermal or supergene origin from Tonganah, near Scottsdale, has been exploited as paper filler.

In contrast, western Tasmania contains a wide variety of reactive host rocks including dolomite, limestone and ultramafic rocks. This has led to a much more diverse suite of deposit styles, the most important of which are sulfide and silicate skarns.

World-class calcic scheelite skarns on King Island (Dolphin, Bold Head; fig. 1) are in the contact aureole of I-type, moderately oxidised unfractionated granodiorite, whereas at the Kara deposit calcic magnetite-scheelite bearing skarn abuts moderately fractionated, moderately to strongly oxidised granite. Calcic magnetite-fluorite skarn with minor tin and tungsten is associated with strongly fractioned, moderately reduced I-type granite at Moina, where there are also distal Au ± Bi ± Cu skarns.

Distal sulfide skarn tin deposits (Renison Bell, Mount Bischoff, Cleveland, Queen Hill, Razorback; fig. 12) are the most economically important and the former two,

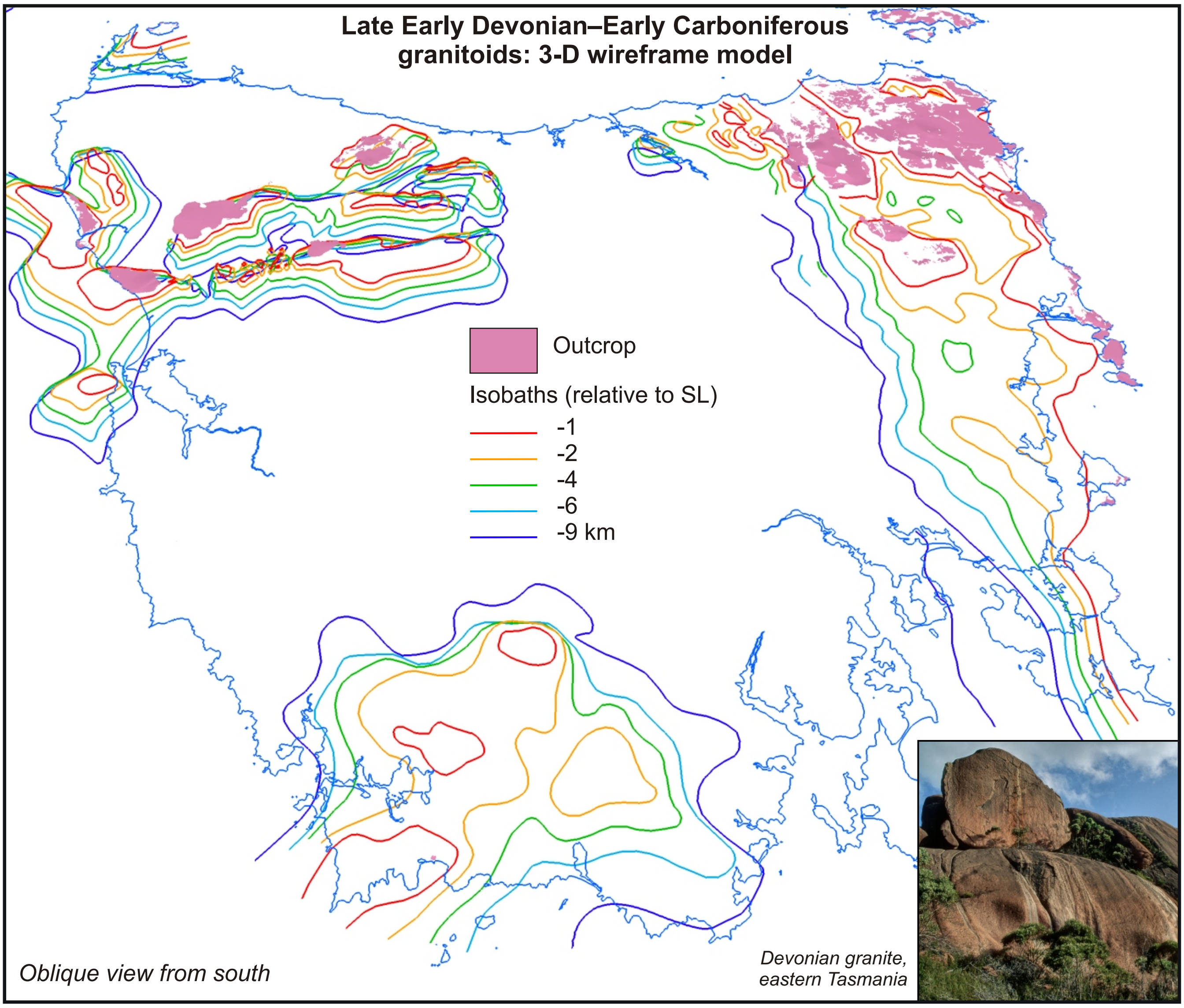



**Figure 13**

*3-D geological model showing outcrop and subsurface form of Devonian–Early Carboniferous granitoids on mainland Tasmania. Isobaths are from Leaman and Richardson (2003).*

both world-class deposits, are associated with reduced, moderately to strongly fractionated I-type granite. Fracture filling quartz-arsenopyrite-pyrrhotite-cassiterite-fluorite veins occupy faults that were active during ore formation at Renison Bell. These were the hydrothermal feeders to the dolomite-replacement stanniferous pyrrhotite horizons (Kitto, 1994; Patterson *et al*., 1981) and are the dominant source of ore. At Mount Bischoff, cassiterite-bearing greisenised quartz porphyry dykes were the conduits for ore formation and form a significant part of the tin resource (Halley and Walshe, 1995; Solomon and Groves, 2000). At the Cleveland mine the subeconomic dyke-hosted deposit (Foley's Zone) contains significant Mo, W and Bi as well as Sn (Collins, 1981; Jackson, 1992) and also appears to have been a fluid conduit.

Tin silicate skarns lie in the immediate contact zones of reduced, fractionated I-type and indeterminate-type granites. These deposits may be metallurgically difficult depending upon the partitioning of tin among cassiterite and a variety of silicate, oxide and borate minerals (Kwak, 1987).

The most contentious deposit is the essentially unstudied Avebury nickel deposit (fig. 12), due to go into production in 2006. Avebury lies in the aureole of the strongly fractionated, reduced Heemskirk white granite and there is debate on whether it is a skarn or a hydrothermally-altered more conventional type of magmatic nickel deposit. Disseminated nickel mineralisation, mostly as pentlandite, occurs within Cambrian serpentinite adjacent to structurally overlying basalt and mudstone-lithicwacke-minor carbonate sequences. There are two main gangue assemblages; antigorite-magnetite-chromite and tremolite-diopside-magnetite, both with pentlandite-pyrrhotite-millerite-arsenides. There are also low grade pentlandite-bearing calcic carbonate-replacement skarn assemblages within the country rock, indicative that nickeliferous skarn formation may occur in this environment. The nearby Burbank deposit is a weathered magnetite-silicate skarn, which contains

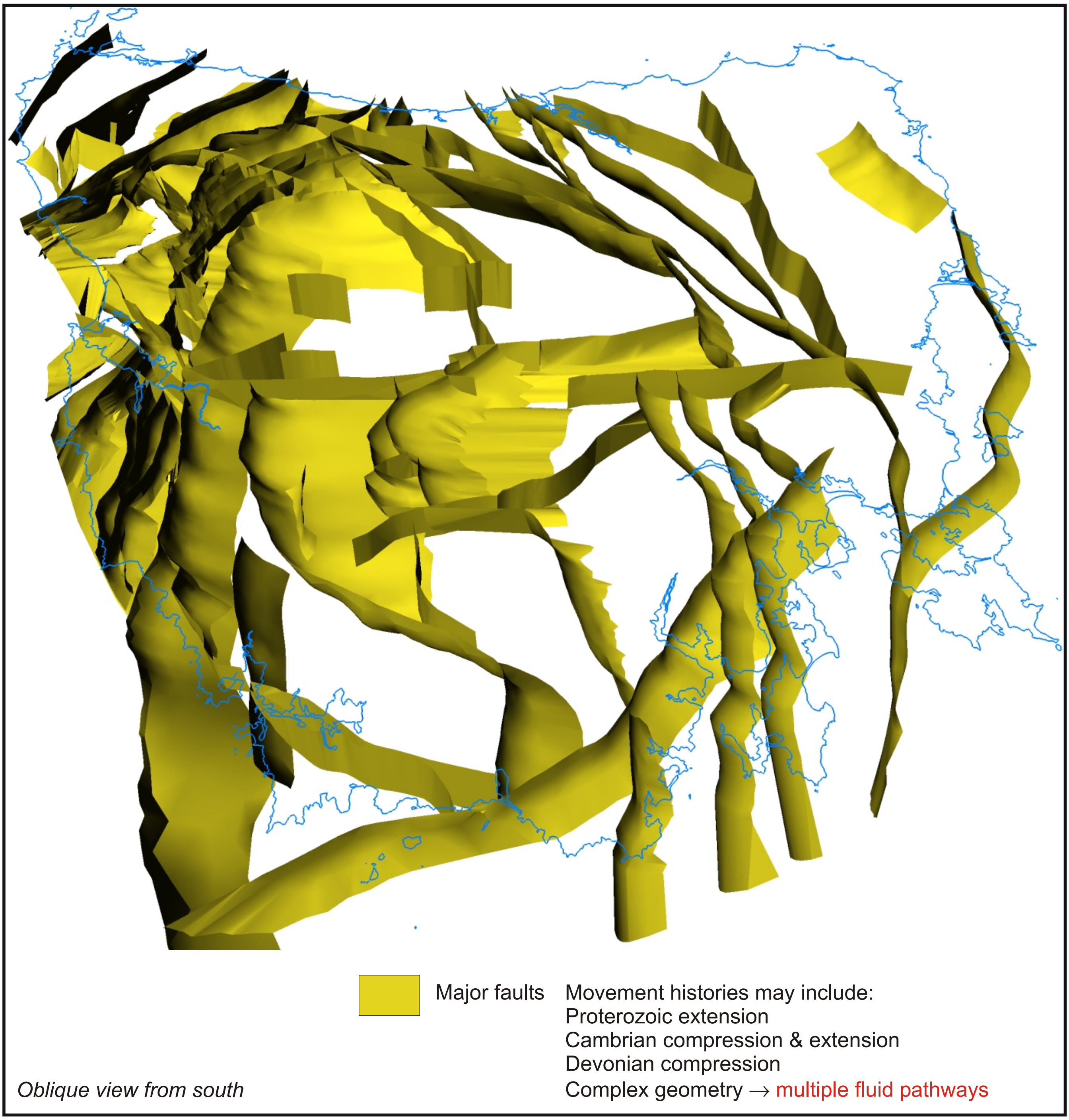



**Figure 14**

*3-D geological model showing major faults on mainland Tasmania.*

the unusual nickel-zinc association. A drill hole intersected 24 metres of 0.77% Ni and 0.28% Zn. It may be significant that magnetite is a major constituent of other several million tonne magnetite-rich, carbonate-replacement skarns in the area; Tenth Legion, which contains minor Sn and Zn and the Sylvester (Comstock) Zn-Pb-Ag deposit. If Avebury represents a new style of nickel skarn deposit, the ingredients are present for repetitions of this deposit type elsewhere in western Tasmania.

Tin greisen (e.g. Sweeneys, Federation) and vein deposits of tin (Pieman vein at East Renison), tungsten (Oakleigh Creek), copper (Murrays Reward), and lead-silver-zinc ± gold (Zeehan and Mount Farrell fields, Magnet; fig. 12) are historically important producers in western Tasmania and are current exploration targets. The latter form conspicuous haloes around the reduced, fractionated granites associated with tin mineralisation, but are relatively rare around the moderately or unfractionated granites with marginal scheelite skarns.

The movement of mineralising fluids outward from the granitoids to favourable sites for ore formation was assisted by the complex pre-existing network of major faults which was largely the net result of Cambrian and Devonian orogenesis (fig. 14). Mineralised structures were initiated or reactivated at the time of granite emplacement and occur in a variety of settings including fractures formed in a regional stress field above granite cupolas (Zeehan Pb-Ag veins) or ridges (Lakeside, Mount Farrell field), faults tangential to margins of granite intrusions (Aberfoyle, Renison Bell) and porphyry dykes that also acted as hydrothermal conduits and form a radiating array above a supposed granite cupola (Mount Bischoff).

Conodont colour alteration index (CAI) values from Gordon and Eldon group carbonate sequences in western Tasmania have been interpreted as dominantly due to Devonian regional metamorphism and granitoid intrusion (Burrett, 1992). Maximum temperatures were in the Dundas Trough and equivalent Palaeozoic depocentres in central northern Tasmania, and in the Olga valley within the Tyennan region, all of which show CAI = 5 indicating temperatures of 300–480°C (fig. 11). It has been suggested that the data imply a very high heat flow during Devonian orogenesis, with a suggested geothermal gradient of 7.5°C per 100 m (Burrett, 1992).

A further event which post-dated the granites, prior to development of the Tasmania Basin, is a possible period of Middle Carboniferous megakinking affecting the Mathinna Supergroup in northeastern Tasmania (Goscombe *et al.*, 1994). This may correlate with a similar event in the Lachlan Fold Belt of mainland Australia.

## Late Carboniferous–Triassic: Tasmania Basin

This and subsequent sections are largely sourced from Stacey and Berry (2004).

Large-scale erosion followed the close of Devonian orogenesis. Deposition restarted in the Late Carboniferous with 1.5 km of generally flat-lying sedimentary rocks of Late Carboniferous to Late Triassic age deposited in the Tasmania Basin. The basin is unconformable on the Late Devonian–Early Carboniferous granites and older folded rocks, with up to one kilometre of basement relief, and is subdivided into two broad lithological and environmental associations, the Lower Parmeener Supergroup and Upper Parmeener Supergroup. The Lower Parmeener Supergroup consists of glacial and glaciomarine and subordinate terrestrial sedimentary rocks, while the Upper Parmeener Supergroup comprises fluvial and lacustrine sedimentary rocks. Both units contain subordinate coal measures. The basin, covering an area greater than 30 000 km$^2$ (fig. 15), is best preserved in northern and southeastern Tasmania. The present basin limits are erosional rather than depositional, and the basin was once considerably larger (Bacon *et al.*, 2000).

The Tasmania Basin has been interpreted as a sag basin (Veevers, 1984), or a foreland basin (Collinson *et al.*, 1987). The sedimentation rate of 1.5 km in 100 million years is substantially slower than in classic foreland

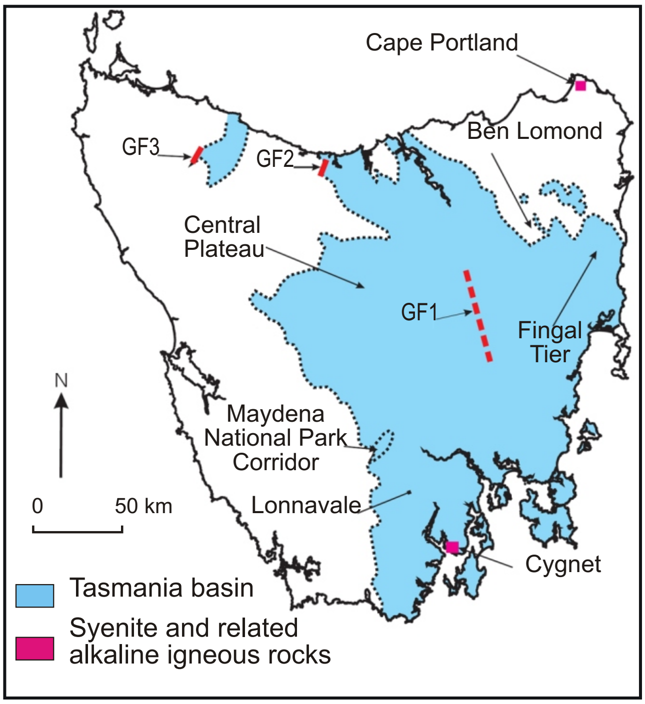



**Figure 15**

*Current extent of the Tasmania Basin, locations of alkaline igneous intrusions and growth faults associated with early basin structure (after Stacey and Berry, 2004).*

basins (Schwab, 1986). Minor normal faulting occurred early in the basin history, but its limited nature and the relatively slow sedimentary accumulation rate is not typical of rift basins. The early development of a marine basin and its subsequent infilling may relate to a eustatic sea-level rise rather than tectonic events, or thermal subsidence. The history of sedimentary accumulation is more typical of a continental margin ('pericratonic') basin. The variable development of lower glacigene beds has been interpreted as the infilling of glacial valleys. Evidence for structural control on the basin evolution is in the form of possible growth faults in the north, northwest and centre of the basin (fig. 15).

The Tasmania Basin has yet to be fully explored using modern hydrocarbon exploration methods. Potential source, reservoir and seal rocks occur within the Lower Parmeener Supergroup, and the main body of the Tasmania Basin is considered mature (Reid *et al.*, 2003). Maturation and initial migration probably resulted from an elevated geothermal gradient in the Middle to Late Cretaceous, however subsequent restructuring has probably modified the trap geometries and complicated the recognition of migration pathways. Reports of oil and gas seeps in Tasmania have almost all proved to be erroneous. A single confirmed bitumen seep at Lonnavale, WSW of Hobart (fig. 15), was derived from a *Tasmanites*-bearing oil shale (Revill, 1996). A suggested mechanism is heat from intruding dolerite generating hydrocarbons from an oil shale source, with migration facilitated by fractures in the dolerite (Bacon *et al.*, 2000).

**Figure 16**

*Form of Jurassic dolerite intrusion in Devonian granite basement (lower part of background cliff) and in Permian sedimentary rocks of the Tasmania Basin (foreground cliff), Cape Surville, southeast Tasmania. (Photo: David Seymour).*

## Jurassic dolerite

By the early Jurassic the Parmeener Supergroup formed a shallow SSE-plunging syncline, with possibly some gentle folding in an otherwise sub-horizontal succession (Hergt *et al*., 1989). Large volumes of tholeiitic dolerite, intruded into the Tasmanian crust during the Middle Jurassic, are probably related to a major thermal anomaly occurring along the eastern margin of Gondwana. The dolerite formed mainly as sills in the Tasmania Basin, and has an estimated volume of 15 000 $km^3$, exposed over an area of 30 000 $km^2$ (Hergt *et al*., 1989). The dolerite has been a major deterrent to petroleum exploration. Most dolerite intrusions have the form of a flattened cone connected to a source or sources at the deepest point, while the limbs are concordant (sills) or approximately concordant with steep transgressions when rising to higher levels (Leaman, 1976) (fig. 16). Metamorphic effects are usually confined to within a few metres of the intrusion margins, although these may be more extensive locally above intrusive feeders. At the Forster prospect in southern Tasmania a small low-grade gold deposit may be a product of metasomatic activity associated with dolerite emplacement, with gold being upgraded locally by later supergene weathering (Bottrill *et al*., 1999).

## Late Jurassic–Cretaceous basin development

Extension related to rifting between Australia and Antarctica initiated the Bass, Durroon and Sorell basins in the latest Jurassic to Early Cretaceous (fig. 17). Extension in the Bass Basin resulted in the development of a series of northwest-trending half-grabens with faults dipping to the southwest (Young *et al*., 1991; Etheridge *et al*., 1985; Hill *et al*., 1995). The Durroon Basin comprises three sub-basins (Baillie and Pickering, 1991) (fig. 17). It adjoins the southeast Bass Basin and its earliest development is related to the same extensional forces, but has a different structural history (Baillie and Pickering, 1991). Sediments of the Sorell Basin, and further north, the Otway Basin, cover the western continental margin of Tasmania (Willcox *et al*., 1989). The Sorell Basin contains four depocentres (fig. 17), each with over four kilometres of sediment thickness. These sub-basins were probably developed in the latest Jurassic to earliest Cretaceous (Hill *et al*., 1997), and have been interpreted as 'relieving bends' on a major left-lateral, strike-slip fault, associated with extension along Australia's southern margin (Moore *et al*., 1992).

Petroleum exploration is active in the offshore basins. Production is due to commence in 2006 from both the

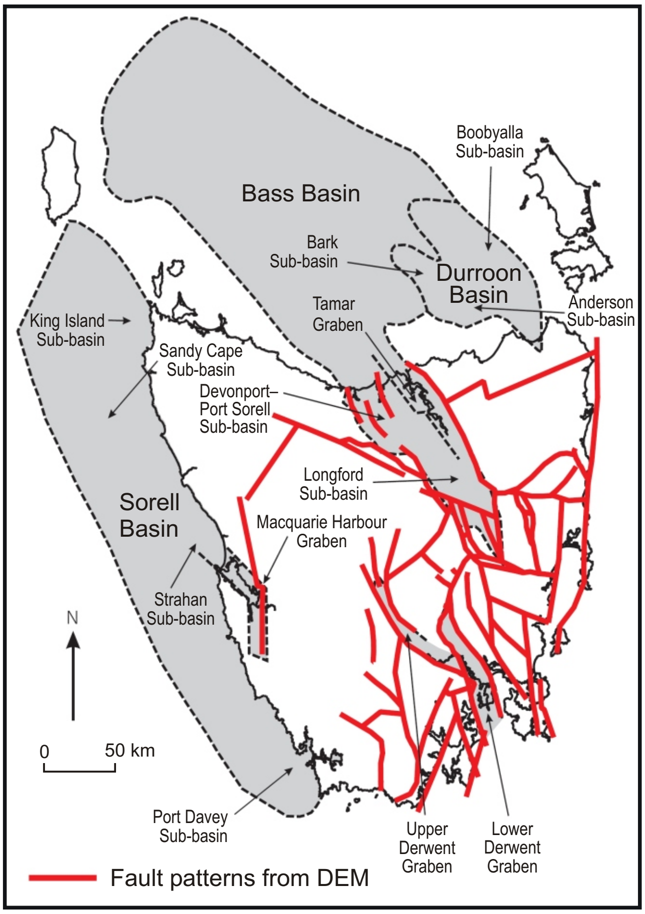



**Figure 17**

*Late Jurassic to Middle Tertiary basins and fault patterns interpreted from high resolution DEM (after Stacey and Berry, 2004).*

Yolla gas-condensate field (Bass Basin) and the Thylacine gas field (Otway Basin).

## Middle to Late Cretaceous

By the mid-Cretaceous, rifting along the southern margin of Australia gave way to spreading, an event recorded in the Bass Basin as an angular unconformity at the top of the Otway Group. In the Sorell Basin rifting was followed by low-energy sag fill or late rift deposition, succeeded by uplift and erosion (Hill *et al.*, 1997). Extension coupled with lithospheric cooling in the Durroon Basin split the basin into a northwest-trending graben and half-graben separated by major listric faults (Baillie and Pickering, 1991). The Durroon Basin experienced increased geothermal gradients (up to 55°C/km) from 100 Ma to 90 Ma, followed by uplift and erosion (Duddy, 1992). A 110 Ma to 90 Ma cooling event is recorded throughout eastern Tasmania, partly in response to kilometre-scale erosion of a thick overlying succession, but also influenced by a Cretaceous magmatic pulse.

Onshore, alkaline igneous intrusions occur in the southeast (Cygnet–Oyster Cove area) and the far northeast (Cape Portland) (fig. 15). Syenite (100.5 ± 3 Ma K-Ar; Evernden and Richards, 1962; McDougall and Leggo, 1965) is part of a felsic, unfractionated felsic alkaline porphyry complex which is composed of a number of sheet-like bodies, sills, dykes and pipes (Farmer, 1985), mainly intruding the Lower Parmeener Supergroup, which at Cygnet is domed above a 750 m thick laccolith lying close to the surface (Leaman, 1990). Gold occurs within the Permian Tasmania Basin country rock and silica-saturated quartz monzodiorite to alkali feldspar syenite porphyries, but there are also feldspathoid-bearing undersaturated rocks. High temperature, highly saline fluids of probable magmatic origin accompanied early stages of gold mineralisation (Taheri and Bottrill, 1999). The Cape Portland alkaline rocks (101.3–102.3 ± 2.6 Ma K-Ar; McDougall and Green, 1982) include small flows, dykes and plug-like intrusions (Jennings and Sutherland, 1969).

Tasmania experienced an elevated geothermal gradient in the Middle to Late Cretaceous. Organic matter in the Parmeener Supergroup at Cygnet was totally carbonised by heating associated with the syenite intrusions (Farmer, 1985), while the Gordon Group at Ida Bay was remagnetised by a Late Cretaceous heating event that persisted for about ten million years (Sharples and Klootwijk, 1981). This Cretaceous thermal event is widely recognised across adjacent parts of Antarctica and New Zealand (Tessensohn, 1994; Veevers, 2000).

Active spreading in the Tasman Sea had begun by the Late Cretaceous. High rates of subsidence associated with extension continued in the Bass Basin (Young *et al.*, 1991). In the Durroon Basin there was growth on northeast-trending extensional faults in the early Late Cretaceous, resulting in thick (up to 2.5 km) wedges of Durroon Formation unconformably overlain by a sag sequence (Hill *et al.*, 1995). By the latest Cretaceous, subsidence in the basin had diminished, with regional subsidence migrating towards the present day depocentre of the Bass Basin (Baillie and Pickering, 1991).

A series of extensional structures began to develop across northeast Tasmania in the latest Cretaceous. The first of these structures, the northwest-striking Tamar Graben (fig. 17), is defined by a series of west-dipping graben and half-graben structures, containing several hundred metres of latest Cretaceous to Tertiary non-marine sediments (Forsyth, 1989).

A phase of minor, dominantly dextral strike-slip faulting in western and southern Tasmania affects rocks as young as 100 Ma and pre-dates normal faulting in the Derwent Graben (Berry and Banks, 1985). It is interpreted as due to a short period of dextral transtension (lateral east-side south plus apart movement) during continental separation of Australia and Antarctica.

## Tertiary

In the late Palaeocene to early Eocene Tasmania was moving NNE along a left-lateral transform which ran along the eastern margin of Antarctica. At about this time the western domain of the South Tasman Rise detached from Antarctica and joined the Australian

Plate, and sea-floor spreading propagated south of the South Tasman Rise and changed to a more southerly direction (Royer and Rollet, 1997). Spreading in the Tasman Sea ceased in the early Eocene (55–50 Ma) (Royer and Rollet, 1997). Major structuring related to these events ceased by the end of the Eocene. The vertical displacement of the Early Tertiary faults can still be recognised in the modern topography (fig. 17).

The Devonport–Port Sorell Sub-basin contains interbedded Palaeocene to late Oligocene non-marine sediments and basalt flows. The dominant structures are NNW–trending, sub-vertical, *en echelon* normal faults. Gravity data suggest that the basin was land-locked, until at least the end of the Eocene (Cromer, 1989).

The Longford Sub-basin has been variously described as a graben with a central horst (Carey, 1947), a southwesterly-dipping surface fractured by normal faults downthrown to the east (Longman, 1966; Longman and Leaman, 1971), and an asymmetrical depression developed on multiple, fluvially incised blocks exhibiting half-graben rotation (Direen and Leaman, 1997). Regional seismic data (Lane, 2002) indicate that the basin consists of a northwest-trending western graben and eastern half-graben separated by a central horst. Many of the faults were continuously active during deposition. The Palaeocene to Eocene basin fill was largely derived from the erosion of basement rocks and deposited under lacustrine/ fluviatile conditions (Matthews, 1989; Lane, 2002).

Palaeocene to mid-Oligocene strike-slip related structures have been interpreted in the Strahan Sub-basin (Willcox *et al*., 1989; Moore *et al*., 1992). An *en echelon*, onshore extension of the Strahan Sub-basin, the Macquarie Harbour Graben (fig. 17), contains about 500 m of Eocene marginal marine sediments (Baillie and Hudspeth, 1989).

The Derwent Graben consists of two linked structures, the lower and upper Derwent Graben. The Lower Derwent Graben is a narrow northwest-trending structure containing only a few hundred metres of sediment, the oldest being of Palaeocene age, indicating faulting was initiated in the Early Tertiary (Colhoun, 1989). A complex transfer zone dominated by north-striking and east-striking faults links the Longford and Derwent grabens (fig. 17).

Tertiary volcanic rocks, dominantly basaltic lavas, are widespread in mainland Tasmania, and extend offshore to Bass Basin, the South Tasman Rise and East Tasman Plateau (Everard *et al*., 2004). Centres range from small plugs and dykes to large necks and probable crater fills, and flows are rarely up to 50 km long. In some cases lava piles have largely buried the pre-volcanic topography. Pyroclastic deposits are associated with some shield volcanos and diatremes. Aquagene volcaniclastic rocks are well-developed in the far northwest, and inland in some highland lakes or where flows have dammed major streams. Radiometric ages range from 16.3 Ma to 64 Ma, but the age distribution cannot be clearly linked to any plume beneath the northward-moving Australian plate (Everard *et al*., 2004). The majority of basalt compositions range from olivine melilitites to quartz tholeiites ($SiO_2$ 37–54%), and are attributed to progressively larger degrees of partial melting. Some undersaturated types have fractionated at mantle depths, producing evolved hawaiite, nepheline hawaiite and nepheline mugearite (Everard *et al*., 2004).

Placer cassiterite deposits have been a significant contributor to the economy of northeastern Tasmania. The bulk of production has come from late Oligocene braidplain deposits around the southeastern edge of the Ringarooma Valley (Morrison, 1989). Production is due to resume in 2006 at the Scotia and Endurance deposits and for the first time sapphire will be a significant co-product.

## Miocene–Pliocene

Mio-Pliocene compressional tectonics are widely recognised in Victoria, and measurements of the modern stress field indicate that Tasmania is also affected by a NNW compressional stress. However there is very little evidence for neotectonic structures in Tasmania, with only one active fault scarp (the Lake Edgar Fault). Tasmania is actively rising at present (Murray-Wallace and Goede, 1995) but this appears to be a regional response to high heat flow rather than a structural response to regional stress.

## Pleistocene

Tasmania was subjected to several major phases of glaciation in the Pleistocene, and valley glaciers descended to sea level in some cases. Valleys and plateaus in central and western Tasmania, particularly in the highland areas, commonly carry superficial deposits of moraine, glaciofluvial outwash gravel and other glaciogene deposits. In some cases these obscure highly prospective Palaeozoic rock units, requiring the use of various deep sounding techniques in mineral exploration.

## Pleistocene to Holocene

There are a number of heavy mineral sands in dune systems and active beaches of Pleistocene to Holocene age. The most important is at Naracoopa on eastern King Island where zircon and rutile have been produced.

## References


ALLEN, R. L. 1995. Synvolcanic, subseafloor replacement model for Rosebery and other massive sulphide ores, *in:* COOKE, D. R.; KITTO, P. A. (ed.). Contentious issues in Tasmanian geology. *Abstracts Geological Society of Australia* 39:107–108.

BACON, C. A.; CALVER, C. R.; BOREHAM, C. J.; LEAMAN, D. E.; MORRISON, K. C.; REVILL, A. T.; VOLKMAN, J. K. 2000. The petroleum potential of onshore Tasmania: a review. *Bulletin Geological Survey Tasmania* 71.

BAILLIE, P. W.; HUDSPETH, J. W. 1989. West Tasmania Region, *in:* BURRETT, C. F.; MARTIN, E. L. (ed.). Geology and mineral resources of Tasmania. *Special Publication Geological Society of Australia* 15:361–365.

BAILLIE, P.; PICKERING, R. 1991. Tectonic evolution of the Durroon Basin, Tasmania. *Exploration Geophysics* 22:13–17.

BALME, B. E. 1960. Palynology of a sediment from Halletts Quarry, Melrose, Tasmania. *Palynology Report Geology Department University of Western Australia* 62:1–14.

BANKS, M. R.; BAILLIE, P. W. 1989. Late Cambrian to Devonian, *in:* BURRETT, C. F.; MARTIN, E. L. (ed.). Geology and mineral resources of Tasmania. *Special Publication Geological Society of Australia* 15:182–237.

BANKS, M. R.; BURRETT, C. F. 1980. A preliminary Ordovician biostratigraphy of Tasmania. *Journal of the Geological Society of Australia* 26:363–376.

BANKS, M. R.; SMITH, E. A. 1968. A graptolite from the Mathinna Beds, north-eastern Tasmania. *Australian Journal of Science* 31:118–119.

BERRY, R. F. 1995. Tectonics of western Tasmania: Late Precambrian–Devonian, *in:* COOKE, D. R.; KITTO, P. A. (ed.). Contentious issues in Tasmanian geology. *Abstracts Geological Society of Australia* 39:6–8.

BERRY, R. F.; BANKS, M. R. 1985. Striations on minor faults and the structure of the Parmeener Super-group near Hobart, Tasmania. *Papers and Proceedings Royal Society of Tasmania* 119:23–29.

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.

BERRY, R. F.; HOLM, O. H.; STEELE, D. A. 2005. Chemical U-Th-Pb dating and the Proterozoic history of King Island, southeast Australia. *Australian Journal of Earth Sciences* 52:461–471.

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 Australian Geological Survey Organisation* 1997/15.

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.

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.

BOTTRILL, R. S.; BROWN, A. V.; CALVER, C. R.; CORBETT, K. D.; GREEN, G. R.; MCCLENAGHAN, M. P., PEMBERTON, J.; SEYMOUR, D. B.; TAHERI, J. 1998. A summary of the economic geology and mineral potential of late Proterozoic and Palaeozoic provinces in Tasmania. *AGSO Journal of Australian Geology and Geophysics* 17:123–143.

BOTTRILL, R. S.; TAHERI, J.; CALVER, C. R. 1999. The nature and origin of gold mineralisation at the Forster Prospect, Glovers Bluff/Weld River area. *Record Tasmanian Geological Survey* 1999/06.

BROWN, A. V. 1986. Geology of the Dundas–Mt Lindsay–Mt Youngbuck region. *Bulletin Geological Survey Tasmania* 62.

BROWN, A. V. 1989. Eo-Cambrian–Cambrian, *in:* BURRETT, C. F.; MARTIN, E. L. (ed.). Geology and mineral resources of Tasmania. *Special Publication Geological Society of Australia* 15:47–83.

BROWN, A. V.; JENNER, G. A. 1989. Geological setting, petrology and chemistry of Cambrian boninite and low Ti-tholeiite lavas in western Tasmania, *in:* CRAWFORD, A. J. (ed.). *Boninites and related rocks*. 232–263. Unwin Hyman : London.

BURNS, K. L. 1964. Geological Atlas 1 Mile Series. Sheet K/55-6-29. Devonport. *Explanatory Report Geological Survey Tasmania*.

BURRETT, C. F. 1992. Conodont geothermometry in Palaeozoic carbonate rocks of Tasmania and its economic implications. *Australian Journal of Earth Sciences* 39:61–66.

BURRETT, C, F.; STAIT, B.; SHARPLES, C.; LAURIE, J. 1984. Middle–Upper Ordovician shallow platform to deep basin transect, southern Tasmania, Australia, *in:* BRUTON, D. L. (ed.). *Aspects of the Ordovician System*. 149–157. Universitetsforlaget : Oslo.

CALLAGHAN, T. 2001. Geology and host-rock alteration of the Henty and Mount Julia gold deposits, western Tasmania. *Economic Geology* 96:1073–1088.

CALVER, C. R. 1996. Reconnaissance isotope chemostratigraphy of Neoproterozoic carbonate rocks in western Tasmania. *Record Geological Survey Tasmania* 1996/10.

CALVER, C. R. 1998. Isotope stratigraphy of the Neoproterozoic Togari Group, Tasmania. *Australian Journal of Earth Sciences* 45:865–874.

CALVER, C. R., TURNER, N. J.; MCCLENAGHAN, M. P.; MCCLENAGHAN, J. 1990. Geological Atlas 1:50 000 Series. Sheet 80 (8112S). Pedder. *Explanatory Report Geological Survey Tasmania*.

CALVER, C. R.; FORSYTH, S. M.; EVERARD, J. L. In press. Explanatory report for the Maydena, Skeleton, Nevada, Weld and Picton 1:25 000 scale digital geological map sheets. *Explanatory Report Tasmanian Geological Survey.*

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.

CAREY, S. W. 1947. Geology of the Launceston district, Tasmania. *Record Queen Victoria Museum* 2:31–46.

COLEMAN, R. J. 1976. Savage River magnetite deposits, *in:* KNIGHT, C. L. (ed.). Economic geology of Australia and Papua New Guinea. *Monograph Serial Australasian Institute of Mining and Metallurgy* 5:598–604.

COLHOUN, E. A. 1989. Cainozoic geomorphology, *in:* BURRETT, C. F.; MARTIN, E. L. (ed.). Geology and mineral resources of Tasmania. *Special Publication Geological Society of Australia* 15:403–409.

COLLINS, P. L. F. 1981. The geology and genesis of the Cleveland tin deposit, western Tasmania: Fluid inclusion and stable isotope studies. *Economic Geology* 76:365–392.

COLLINSON, J. W.; KEMP, N. R.; EGGERT, T. J. 1987. Comparison of the Triassic Gondwana sequences in the Transantarctic mountains and Tasmania, *in:* MCKENZIE, G. D. (ed.). Gondwana Six: Stratigraphy, sedimentology and paleontology. *Geophysical Monograph American Geophysical Union* 41:51–61.

CORBETT, K. D. 1989. Stratigraphy, palaeogeography and geochemistry of the Mt Read Volcanics, *in:* BURRETT, C. F.; MARTIN, E. L. (ed.). Geology and mineral resources of Tasmania. *Special Publication Geological Society of Australia* 15:86–119.

CORBETT, K. D. 1992. Stratigraphic-volcanic setting of massive sulfide deposits in the Cambrian Mount Read Volcanics, Tasmania. *Economic Geology* 87:564–586.

CORBETT, K. D. 2001. New mapping and interpretations of the Mount Lyell mining district, Tasmania: A large hybrid Cu-Au system with an exhalative Pb-Zn top. *Economic Geology* 96:1089–1122.

CORBETT, K. D. 2002. Western Tasmanian Regional Minerals Program. Mount Read Volcanics Compilation. Updating the geology of the Mount Read Volcanics belt. *Record Geological Survey Tasmania* 2002/19.

COX, S. F. 1973. *The structure and petrology of the Cape Wickham area, King Island*. B.Sc.(Hons) thesis, University of Tasmania.

CRAWFORD, A. J.; BERRY, R. F. 1992. Tectonic implications of Late Proterozoic–Early Palaeozoic igneous rock associations in western Tasmania. Tectonophysics 214:37–56.

CROMER, W. C. 1989. Devonport-Port Sorell Sub-basin, *in:* BURRETT, C. F.; MARTIN, E. L. (ed.). Geology and mineral resources of Tasmania. *Special Publication Geological Society of Australia* 15:361.

DIREEN, N. G.; LEAMAN, D. E. 1997. Geophysical modelling of structure and tectonostratigraphic history of the Longford Basin, northern Tasmania. *Exploration Geophysics* 28:29–33.

DUDDY, I. R. 1992. Assessment of thermal history data from the Durroon-1 well, Bass Basin. *Report Geotrack International Pty Ltd* 386 [OR 336F].

ETHERIDGE, M. A.; BRANSON, J. C.; STUART-SMITH, P. G. 1985. Extensional basin-forming structures in Bass Strait and their importance for hydrocarbon exploration. *Journal Australasian Petroleum Exploration Association* 25:344–361.

EVERARD, J. L.; SEYMOUR, D. B.; REED, A. R.; MCCLENAGHAN, M. P.; GREEN, D. C.; CALVER, C. R.; BROWN, A. V. 2001. Regional geology of the southern Smithton Synclinorium. Explanatory report for the Roger, Sumac and Dempster 1:25 000 scale map sheets, far northwestern Tasmania. *Explanatory Report Tasmanian Geological Survey* [unpublished].

EVERARD, J. L.; ZHANG, M.; LO, C.-H.; O'REILLY, S.; FORSYTH, S. M. 2004. Overview of Tasmanian Tertiary basalts *Abstracts Geological Society of Australia* 73:74.

EVERNDEN, J. F.; RICHARDS, J. R. 1962. Potassium-argon ages in eastern Australia. *Journal of the Geological Society of Australia* 9:1–37.

FARMER, N. 1985. Geological Atlas 1:50 000 Series. Sheet 88 (8311N). Kingborough. *Explanatory Report Geological Survey Tasmania*.

FORSYTH, S. M. 1989. The Tamar Graben, *in:* BURRETT, C. F.; MARTIN, E. L. (ed.). Geology and mineral resources of Tasmania. *Special Publication Geological Society of Australia* 15:358–361.

FROST, M. T. 1982. The magnesite deposit at Main Creek, Savage River, Tasmania. *Economic Geology* 77:1901–1911.

FROST, M. T.; MATZAT, H. W. 1984. A further large magnesite deposit along the Savage River in northwestern Tasmania. *Economic Geology* 79:404–408.

GEMMELL, J.B.; LARGE, R. R. 1992. Stringer system and alteration zones underlying the Hellyer volcanogenic massive sulfide deposit, Tasmania, Australia. *Economic Geology* 87:620–649.

GIFKINS, C. C.; KIMBER, B. 2003. *3-D geological modelling and mineral systems in Tasmania. Explanatory notes for the map of volcanic facies associations and volcanic centres in the Mount Read Volcanics, Tasmania*. Centre for Ore Deposit Research University of Tasmania [unpublished].

GOSCOMBE, B. D.; FINDLAY, R. H.; MCCLENAGHAN, M. P.; EVERARD, J. L. 1994. Multi-scale kinking in northeast Tasmania: crustal shortening at shallow crustal levels. *Journal of Structural Geology* 16:1077–1092.

GREEN, G. R. 1983. *The geological setting and formation of the Rosebery volcanic-hosted massive sulphide orebody, Tasmania*. Ph.D. Thesis, University of Tasmania.

GREEN, G. R.; SOLOMON, M.; WALSHE, J. L. 1981. The formation of the Rosebery volcanic-hosted massive sulfide ore deposit at Rosebery, Tasmania. *Economic Geology* 76:304–338.

GREEN, G. R.; TAHERI, J. 1992. Stable isotopes and geochemistry as exploration indicators. *Bulletin Geological Survey Tasmania* 70:84–91.

HALLEY, S. W.; WALSHE, J. L. 1995. A reexamination of the Mount Bischoff cassiterite sulfide skarn, western Tasmania. *Economic Geology* 90:1676–1693.

HERGT, J. M.; MCDOUGALL, I.; BANKS, M. R.; GREEN, D. H. 1989. Jurassic dolerite, *in:* BURRETT, C. F.; MARTIN, E. L. (ed.). Geology and mineral resources of Tasmania. *Special Publication Geological Society of Australia* 15:375–381.

HERRMANN, W.; KIMBER, B. 2003. *Map of altered facies in the Mount Read Volcanics, Tasmania*. Centre for Ore Deposit Research University of Tasmania [unpublished].

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, *in:* BUCHANAN, J. G.; BUCHANAN, P. G. (ed.). Basin inversion. *Special Publication Geological Society* 88:525–547.

HILL, P. J.; MEIXNER, A. J.; MOORE, A. M. G.; EXON, N.F. 1997. Structure and development of the west Tasmanian offshore sedimentary basins: results of recent marine and aeromagnetic surveys. *Australian Journal of Earth Sciences* 44:579–596.

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.

HUSTON, D. L.; KAMPRAD, J. 2001. Zonation of alteration facies at Western Tharsis: Implications for the genesis of Cu-Au deposits, Mount Lyell field, western Tasmania. *Economic Geology* 96:1123–1132.

JACKSON, P. 1992. *The evolution of Foley's Zone, Cleveland Mine, Tasmania, Sn-Mo-Bi mineralization, geology, petrology, fluid inclusions and stable isotopes*. Ph.D. thesis, LaTrobe University.

JAGO, J. B.; BROWN, A. V. 1989. Middle to Upper Cambrian fossiliferous sedimentary rocks, *in:* BURRETT, C. F.; MARTIN, E. L. (ed.). Geology and mineral resources of Tasmania. *Special Publication Geological Society of Australia* 15:74–83.

JENNINGS, D. J.; SUTHERLAND, F. L. 1969. Geology of the Cape Portland area with special reference to the Mesozoic(?) appinitic rocks. *Technical Report Department of Mines Tasmania* 13:45–82.

KEELE, R. A. 1994. Structure and veining in the Devonian-aged Mathinna–Alberton gold lineament, northeast Tasmania. *Report Mineral Resources Tasmania* 1994/06.

KITTO, P. A. 1994. *Structural and geochemical controls on mineralisation at Renison, Tasmania*. Ph.D. Thesis, University of Tasmania.

KWAK, T. A. P. 1987. W-Sn deposits and related metamorphic skarns and granitoids. *Developments in Economic Geology* 24.

LANE, P. 2002. *Seismic interpretation and basin analysis of the Longford Sub-Basin*. B.Sc. (Hons) Thesis, University of Tasmania.

LARGE, R. R.; DOYLE, M. G.; RAYMOND, O. L.; COOKE, D. R.; JONES, A. T.; 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.

LAURIE, J. R. 1995. Examination of macrofossils from the Pioneer Beds from near Queenstown, Tasmania. *Professional Opinion Australian Geological Survey Organisation* 1995/018.

LEAMAN, D. E. 1976. Geological Atlas 1:50 000 series. Sheet 82 (8312S). Hobart. *Explanatory Report Geological Survey Tasmania*.

LEAMAN, D. E. 1990. Inferences concerning the distribution and composition of pre-Carboniferous rocks in southeastern Tasmania. *Papers and Proceedings Royal Society of Tasmania* 124:1–12.

LEAMAN, D. E.; RICHARDSON, R. G. 2003. A geophysical model of the major Tasmanian granitoids. *Record Tasmania Geological Survey* 2003/11.

LONGMAN, M.J. 1966. One Mile Geological Map Series. K/55-7-39. Launceston. *Explanatory Report Geological Survey Tasmania*.

LONGMAN, M. J.; LEAMAN, D. E. 1971. Gravity survey of the Tertiary Basins in northern Tasmania. *Geological Survey Bulletin Tasmania* 51.

MATTHEWS, W. L. 1989. Longford Sub-Basin, *in:* BURRETT, C. F.; MARTIN, E. L. (ed.). Geology and mineral resources of Tasmania. *Special Publication Geological Society of Australia* 15:370–372.

MATZAT, H. W. 1984. *Mineralogical, petrological and geochemical features of ore formation of Savage River, Tasmania (Australia)*. Ph.D. thesis, University of Heidelberg.

MEFFRE, S.; BERRY, R. F.; HALL, M. 2000. Cambrian metamorphic complexes in Tasmania: tectonic implications. *Australian Journal of Earth Sciences* 47:971–985.

MCDOUGALL, I.; GREEN, D. C. 1982. Cretaceous K/Ar ages from northeastern Tasmania. Appendix 3 *in:* MCCLENAGHAN, M. P.; TURNER, N. J.; BAILLIE, P. W.; BROWN, A. V.; WILLIAMS, P. R.; MOORE, W. R. Geology of the Ringarooma–Boobyalla area. *Bulletin Geological Survey Tasmania* 61:179–181.

MCDOUGALL, I.; LEGGO, P. J. 1965. Isotopic age determinations on granitic rocks from Tasmania. *Journal of the Geological Society of Australia* 12:295–332.

MOORE, A. M. G.; WILLCOX, J. B.; EXON, N. F.; O'BRIEN, G. W. 1992. Continental shelf basins on the west Tasmania margin. *Journal Australasian Petroleum Exploration Association* 32:231–250.

MORRISON, K. C. 1989. Tin fields of northeastern Tasmania, *in:* BURRETT, C. F.; MARTIN, E. L. (ed.). Geology and mineral resources of Tasmania. *Special Publication Geological Society of Australia* 15:369.

MORRISON, K. C. 2002. Western Tasmanian Regional Minerals Program. Mount Read Volcanics Compilation. Report on field investigations, Mt Darwin–Mt Murchison region. *Record Tasmania Geological Survey* 2002/18.

MURPHY, F. C.; DENWER, K.; KEELE, R. A.; STAPLETON, P.; SEYMOUR, D. B.; GREEN, G. R. 2004. Tasmania mineral province: geoscientific database, 3D geological modelling, mines and mineral prospectivity. Mineral Resources Tasmania [unpublished].

MURRAY-WALLACE, C. V.; GOEDE, A. 1995. Aminostratigraphy and electron spin resonance dating of Quaternary coastal neotectonism in Tasmania and the Bass Strait islands. *Australian Journal of Earth Sciences* 42:51–67.

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.

PATISON, N. L.; BERRY, R. F.; DAVIDSON, G. J.; TAYLOR, B. P.; BOTTRILL, R. S.; MANZI, B.; RYBA, J.; SHEPHERD, R. E. 2001. Regional metamorphism of the Mathinna Group, northeast Tasmania. *Australian Journal of Earth Sciences* 48:281–292.

PATTERSON, D. J.; OHMOTO, H.; SOLOMON, M., 1981. Geologic setting and genesis of cassiterite-sulfide mineralization at Renison Bell, western Tasmania. *Economic Geology* 76:393–438.

POWELL, C. McA. 1991. Structure of the Beaconsfield and Lefroy goldfields. *Report Division of Mines and Mineral Resources Tasmania* 1991/16.

POWELL, C. McA.; 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.

REED, A. R. 2001. Pre-Tabberabberan deformation in eastern Tasmania: a southern extension of the Benambran Orogeny. *Australian Journal of Earth Sciences* 48:785–796.

REED, A. R. 2002. Formation of lode-style gold mineralisation during Tabberabberan wrench faulting at Lefroy, eastern Tasmania. *Australian Journal of Earth Sciences* 49:879–890.

REID, C. M.; CHESTER, A. D.; STACEY, A. R.; BURRETT, C. F. 2003. Stratigraphic results of diamond drilling of the Hunterston Dome, Tasmania: implications for palaeogeography and hydrocarbon potential. *Papers and Proceedings Royal Society of Tasmania* 137:87–94.

REVILL, A. M. 1996. Hydrocarbons isolated from Lanna Vale seep. Swab and bitumen samples. *Report CSIRO Division of Oceanography* TDR-1 [TCR96-3900].

RICKARDS, R. B.; BANKS, M. R. 1979. An Early Devonian monograptid from the Mathinna Beds, Tasmania. *Alcheringa* 3:307–311.

RICKARDS, R. B.; DAVIDSON, G.; BANKS, M. R. 1993. Silurian (Ludlow) graptolites from Golden Ridge, NE Tasmania. *Memoir Association of Australasian Palaeontologists* 15:125–135.

ROYER, J. Y.; ROLLET, N. 1997. Plate-tectonic setting of the Tasmanian region. *Australian Journal of Earth Sciences* 44:543–560.

SCHWAB, F. L. 1986. Sedimentary 'signatures' of foreland basin assemblages: real or counterfeit? *in:* ALLEN, P. A.; HOMEWOOD, P. (ed.). Foreland basins. *Special Publication International Association of Sedimentologists* 8:395–410.

SEYMOUR, D. B.; CALVER, C. R. 1995. Explanatory notes for the Time-Space Diagram and Stratotectonic Elements Map of Tasmania. *Record Tasmanian Geological Survey* 1995/01.

SHARPLES, C.; KLOOTWIJK, C. T. 1981. Palaeomagnetic results from the Gordon Subgroup of Tasmania: further evidence for a Late Cretaceous magnetic overprint in southeastern Australia. *Papers and Proceedings Royal Society of Tasmania* 115:85–91.

SOLOMON, M.; GROVES, D. I. 2000. *The geology and origin of Australia's mineral deposits, reprinted with additional material*. Centre for Ore Deposit Research, University of Tasmania and Centre for Global Metallogeny, University of Western Australia.

SOLOMON, M.; VOKES, F. M.; WALSHE, J. L. 1987. Chemical remobilization of volcanic-hosted massive sulphide deposits at Rosebery and Mt Lyell, Tasmania. *Ore Geology Reviews* 2:173–190.

SPILLER, A. R. 1974. *Petrology of the Savage River iron ore deposit, Tasmania*. B.A. (Hons.) thesis, Macquarie University.

STACEY, A. R.; BERRY, R. F. 2004. The structural history of Tasmania: a review for petroleum explorers. *PESA Eastern Australasian Basins Symposium II, Adelaide*.

TAHERI, J.; BOTTRILL, R. S. 1999. Porphyry and sediment-hosted gold deposits near Cygnet: new styles of gold mineralisation in Tasmania. *Record Geological Survey Tasmania* 1999/01.

TAYLOR, S.; MATHISON, I. J. 1990. Oceana lead-zinc-silver deposit, *in:* HUGHES, F. E. (ed.). Geology of the mineral deposits of Australia and Papua New Guinea. *Monograph Serial Australasian Institute of Mining and Metallurgy* 14:1253–1256.

TESSENSOHN, F. 1994. The Ross Sea region, Antarctica: structural interpretation in relation to the evolution of the Southern Ocean. *Terra Antarctica* 1:553–558.

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.

TURNER, N. J.; BOTTRILL, R. S. 1993. Blue amphibole in the Proterozoic to Cambrian Arthur Metamorphic Complex, northwest Tasmania. *Report Mineral Resources Tasmania* 1993/26.

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.

VEEVERS, J. J. (ed.). 1984. *Phanerozoic earth history of Australia*. Clarendon Press : Oxford.

VEEVERS, J. J. 2000. Mid-Cretaceous and mid-Eocene events in the Indo-Australian and Antarctic Plates, *in:* VEEVERS, J. J. (ed.). *Billion-year earth history of Australia and neighbours in Gondwanaland*. 102–109. GEMOC Press.

WALSHE, J. L.; SOLOMON, M. 1981. An investigation into the environment of formation of the volcanic-hosted Mt Lyell copper deposits using geology, mineralogy, stable isotopes, and a six-component chlorite solid-solution model. *Economic Geology* 76:246–284.

WHITE, M. J.; MCPHIE, J. 1996. Stratigraphy and palaeovolcanology of the Cambrian Tyndall Group, Mt Read Volcanics, western Tasmania. *Australian Journal of Earth Sciences* 43:147–159.

WILLCOX, J. B.; BAILLIE, P. W.; EXON, N. F.; LEE, C. S.; THOMAS, B. 1989. The geology of western Tasmania and its continental margin — with particular reference to petroleum potential. *Record Bureau of Mineral Resources Geology and Geophysics Australia* 1989/13.

WILLIAMS, E.; MCCLENAGHAN, M. P.; COLLINS, P. L. F. 1989. Mid-Palaeozoic deformation, granitoids and ore deposits, *in:* BURRETT, C. F.; MARTIN, E. L. (ed.). Geology and mineral resources of Tasmania. *Special Publication Geological Society of Australia* 15:238–292.

YOUNG, I. M.; TRUPP, M. A.; GIDDING, M. J. 1991. Tectonic evolution of Bass Strait — Origins of Tertiary inversion. *Exploration Geophysics* 22:465–468.

**Department *of* Infrastructure, Energy *and* Resources**
**MINERAL RESOURCES TASMANIA**

PO Box 56  Rosny Park
Tasmania  Australia  7018
30 Gordons Hill Road  Rosny Park
Tasmania  Australia
Tel: +61 3 6233 8377  Fax: +61 3 6233 8338
Email: info@mrt.tas.gov.au
Internet: www.mrt.tas.gov.au

Mineral Resources Tasmania
mrt
doing the groundwork
DEPARTMENT *of* INFRASTRUCTURE, ENERGY *and* RESOURCES