Devonian granites and associated mineralisation in northeast and northwest Tasmania

Excursion Guide — May 2005

Tasmanian Geological Survey Record 2005/03

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MINERAL RESOURCES TASMANIA

DEPARTMENT *of* INFRASTRUCTURE, ENERGY *and* RESOURCES

Mineral Resources Tasmania
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Devonian granites and associated mineralisation in northeast and northwest Tasmania

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Introduction

These notes have been produced as a guide for an excursion covering Devonian granites and mineralisation in northeast and northwest Tasmania. The aim of this four-day excursion was to provide participants with the opportunity to inspect some of the more accessible and interesting Devonian granites and their associated mineralisation in northeast Tasmania. Three geologically and economically significant granite-related deposits were also inspected in north and northwest Tasmania.

The excursion travelled from Hobart to northeast and then northwest Tasmania, before returning to Hobart. This guide includes excerpts from numerous published and unpublished studies of the granites and related mineral deposits.

Summarised itinerary

The locations of the stops are shown on the map on Page 37.

Day 1 (22 May 2005)

Depart Wrest Point Casino 9.00 am

1.1 Garnet-biotite, adamellite/granite, Bicheno

1.2 Mathinna Supergroup (time permitting)

1.3: Alkali feldspar/greisenised granite with associated tin and uranium mineralisation, Royal George mine

1.4 St Marys Porphyrite (time permitting)

Overnight: Arrive at Scamander about 5.30 pm.

Day 2 (23 May 2005)

Depart Scamander 8.30 am

2.1 Granodiorite and associated copper and molybdenite mineralisation at Akaroa quarry, St Helens

2.2 Gardens granodiorite, St Helens

2.3 Anchor mine (including old battery site, time permitting)

2.4 Derby Museum

2.5 Alkali feldspar granite, Tulendeena (time permitting)

2.6 Tonganah clay deposit (time permitting)

Overnight: Arrive Bridport about 5.00 pm.

Day 3 (24 May 2005)

Depart Bridport 8.30

3.1 Contact between granodiorite and Mathinna Supergroup, Bridport

3.2 Golconda goldfield

3.3 Winery

3.4 Shepherd and Murphy mine, Moina (time permitting)

Overnight: Arrive Cradle Mountain about 5.00 pm

Day 4 (25 May 2005)

Depart Cradle Mountain 8.30 am

4.1 Mt Bischoff, Waratah

4.2 Kara mine, Hampshire

Arrive Hobart about 6.00 pm

Cover: The Desert Face, Mt Bischoff, showing variably oxidised tin and sulphide-rich magnesian skarns (mostly red and yellow), in contact with Burnie Formation siltstone and quartzite (pale grey).

Safety issues

There will be a number of important safety issues to consider on this trip, and participants should always stay with the group and heed all warnings and directions from the trip leaders, mine personnel or other authorities. Hard hats, boots and safety glasses should be used at all mines sites and as directed elsewhere. Do not enter any underground workings or approach overhangs. The main issues are:

Hazard	*Sites*	*Procedure*
Abandoned, unstable mine workings	Royal George, Mt Bischoff, Anchor mine	Watch out for unstable faces and shafts
Active mine workings	Kara mine	Wear all safety gear, watch for mine vehicles and unstable faces
Respirable dusts containing possible quartz, uranium and asbestiform amphiboles	Royal George, Mt Bischoff, Anchor mine, Kara mine	Avoid dusty areas or generating dusts
Radioactivity in mineralised granites	Royal George	Restrict time spent in indicated hazardous areas, and minimise sample collecting
Traffic hazards at road cuts	Fingal, Tonganah	Wear safety vests
Hypothermia	Mt Bischoff, Waratah	Dress in preparation for bad weather
Coastal hazards (e.g. slippery rocks, waves, tides)	Bicheno, Bridport	Watch footing

Northeast Tasmanian geology*

Introduction

The pre-Carboniferous geology of northeast Tasmania comprises a folded and faulted package of Cambrian or Ordovician to Devonian turbidites belonging to the Mathinna Supergroup (Powell *et al.*, 1993). Several batholiths comprising both I-type and S-type granites intrude the Mathinna Supergroup. The area is of considerable economic importance, containing more than 600 known gold deposits.

Mathinna Supergroup

The stratigraphy of the Mathinna Supergroup is documented west of the Scottsdale Batholith (Powell and Baillie, 1992; Powell *et al.*, 1993). Here, it comprises (in depositional order) the Stony Head Sandstone, Turquoise Bluff Slate, Bellingham Formation, and Sidling Sandstone (fig. 1).

Throughout northeast Tasmania, Mathinna Supergroup rocks typically consist of fining-up Bouma sequences of less than two metres thickness comprising variable quantities of sand to silt-sized, quartz-dominated, detritus. Beds are commonly arranged into upwardly-fining packages. Cross-lamination is restricted to C-intervals of Bouma units. Slumping, scour, and load casts, as well as graded bedding, indicate a deep water turbiditic setting for Mathinna Supergroup sedimentation (Walker, 1957; Williams, 1959; Marshall, 1969). Conglomerate has not been observed except in one section near the contact between the Stony Head Sandstone and Turquoise Bluff Slate. Here, rounded clasts of microcrystalline chert to 10 mm form layers at the base of sandstone beds.

Time of sedimentation

The Turquoise Bluff Slate at the Australasian Slate Quarry (near Back Creek) contains Arenig to Llanvirn age graptolites (Banks and Smith, 1968) showing that deposition of the Mathinna Supergroup commenced during or just prior to the early Ordovician. Dating of detrital zircons (unpublished) from the Stony Head Sandstone indicates that deposition commenced some time after about 540 Ma. Plant fossils from the east coast near Scamander (Cookson, 1937; Banks, 1962; Rickards and Banks, 1979) and radiometric dating of granitoids (McClenaghan and Higgins, 1993) constrain the upper age of the Mathinna Supergroup to the Early Devonian, between 405 Ma to 388 Ma. The Mathinna Supergroup is therefore Cambrian or Ordovician to Devonian in age.

Thickness

The total thickness of the Mathinna Supergroup is poorly constrained, but probably between five (Keele, 1994) and seven (Powell and Baillie, 1992) kilometres.

Stratigraphy

Poor fossil control has made it difficult to determine the stratigraphy of the Mathinna Supergroup. Banks (1962) first subdivided the Mathinna Supergroup into two sedimentary associations based on gross compositional differences; an Ordovician shale-dominated succession and a younger mixed shale-sandstone succession. Turner (1980) proposed a fault separating Ordovician and Siluro-Devonian rocks, based largely on the lack of a recognised Silurian fauna. After completing a structural traverse along the George Town–Bridport road, Powell and Baillie (1992) concluded that the contact between the shale-dominated succession and the younger mixed shale-sandstone succession was conformable. The subsequent discovery of Silurian graptolites from the Golden Ridge area between Mathinna and Scamander (Rickards *et al.*, 1993) further supported the hypothesis that sedimentation was continuous. New structural data across the boundary between the shale-dominated and sandstone-dominated Turquoise Bluff Slate and Bellingham Formation shows that the older rocks have suffered an extra deformation. Consequently, the boundary between the two formations is either a fault or an unconformity.

Provenance

The provenance of the Stony Head Sandstone remains enigmatic. Strontium isotope data indicate Proterozoic siliciclastic and Cambrian to Ordovician Mt Read Volcanics rocks in western Tasmania were not a sediment source for the Mathinna Supergroup turbidites (Gray and Webb, 1995), a conclusion corroborated by recently acquired but as yet unpublished U-Pb data. These results are intriguing as palaeocurrent data from this part of the Mathinna Supergroup show derivation of sediment from the west (Powell *et al.*, 1993) where outcropping Proterozoic and early Palaeozoic rocks dominate.

Post-Mathinna Supergroup sedimentary and igneous rocks

Granitoid rocks

Three roughly NNW-oriented elongate composite I-type and S-type granitoid batholiths of Late Devonian age intrude Mathinna Supergroup rocks. These are (from west to east) the Scottsdale, Blue Tier and Eddystone batholiths. The batholiths generally comprise (in order of decreasing age) hornblende-biotite granodiorite, biotite granite/ adamellite and alkali-feldspar granite. Each is surrounded by narrow metamorphic aureoles indicative of intrusion at a high crustal level (emplacement depth of granodiorites range from about 1 km to 12 km based on hornblende geobarometry; Varne and Fulton, 1994).

* From Reed, 2001a

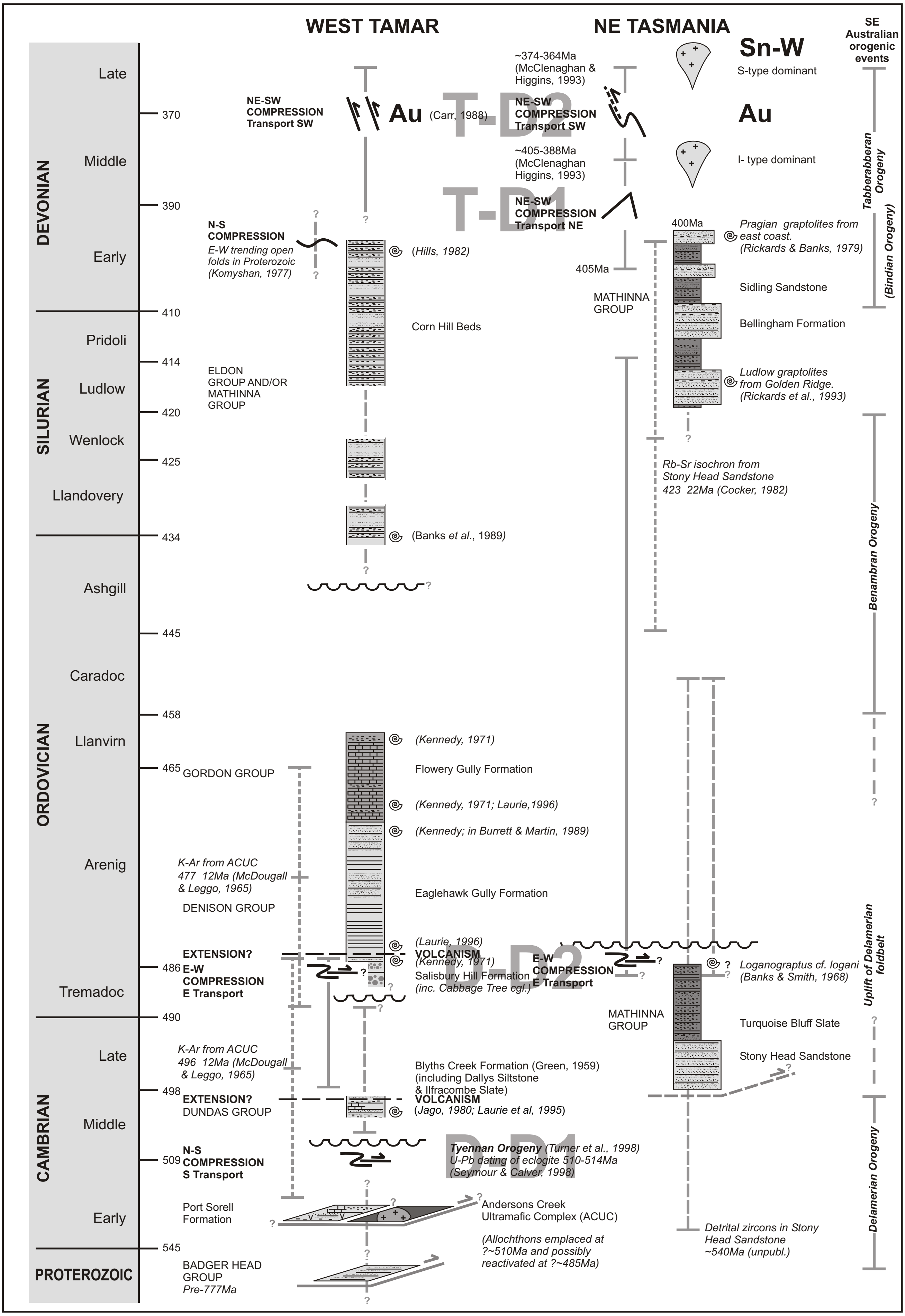


Figure 1

Time-Space diagram for pre-Carboniferous rocks east of the Port Sorell embayment (from Reed, 2001a).

Post-Carboniferous rocks

Mathinna Supergroup and granitoid rocks are unconformably overlain by flat-lying Permo-Triassic rocks of the Parmeener Supergroup which are intruded by sills of Jurassic dolerite. Exhumation and weathering during the Tertiary was accompanied by basaltic volcanism. The basalt is preserved in lowland areas along the north coast and scattered plateau remnants inland. The Parmeener Supergroup is typically unmineralised although palaeoplacer cassiterite deposits occur in the Roys Hill area and there is a small quantity of gold in the Mangana area. Pleistocene to Recent sedimentary deposits are largely restricted to valley bottoms and where proximal mineralised Mathinna Supergroup rocks may contain alluvial gold.

Regional structure and mineralisation in the Mathinna Supergroup

Introduction

The rocks of the Mathinna Supergroup in northeast Tasmania have experienced three major deformational episodes (Reed, 2001*a*, 2001*b*, 2004). The first was a NE-directed thrusting and associated recumbent folding event (D_1), only developed in the Ordovician-aged Stony Head Sandstone and Turquoise Bluff Slates of the Tippogoree Group (i.e. restricted to the region west of the Scottsdale Batholith). This deformation corresponded to the culmination of the Delamerian Orogeny in western Tasmania (D–D_2, according to the terminology adopted by Reed (fig. 1). Reed has speculated that an unconformity, which is located in the vicinity of Pipers River, separates the Tippogoree Group from the Panama Group.

A second deformation occurred during the Middle Devonian Tabberabberan Orogeny and involved the formation of upright chevron folds with limbs overturned towards the northeast (T–D_1, see fig. 1). A third deformation event caused a second generation of upright folds to form, also of Middle Devonian age, but this time associated with SW-verging thrusts (T–D_2). The effects of this deformation are strongest around the Tamar Valley (e.g. Port Sorell and Beaconsfield), where a series of east-dipping thrusts played an important role in stitching the terrains of eastern and western Tasmania together at the close of the Tabberabberan Orogeny (fig. 2).

The metamorphic grade and structural setting of pelites in the Mathinna Supergroup has been studied by the illite crystallinity method (Patison *et al.*, 2001). This work indicated that the regional metamorphic grade varied from diagenetic to lower greenschist (excluding metamorphic aureoles). The data can be interpreted to indicate the location of major structural breaks (e.g. thrusts) and the approximate depth within the stratigraphy.

The three deformation episodes will be referred to as D_1, D_2 and D_3 respectively for the purposes of this report.

D_1

Macroscopic D_1 structures, such as the Pipers River recumbent fold, occur east of the River Tamar within the Stony Head Sandstone and Turquoise Bluff Slate (Powell and Baillie, 1992). A northwest-trending zone of D_1 thrusting and folding passes through the gold mineralised lode system at Lefroy. D_1 structures have been described in detail from drill core at the Volunteer mine and at the Sir John Franklin mine at Back Creek. A shallow-dipping high-strain zone, commonly silicified and up to 300 mm thick, occurs with folded chert layers that indicate shortening of up to 90% in the hangingwall of the fault (Reed, 2001*a*). Fold vergence and quartz fibre directions indicate tectonic transport of the hangingwall rocks towards the northeast. Recumbent folds are open to tight and plunge northwest and southeast (average 13° towards 135°). Regionally, bedding strikes northwest and dips to the northeast and southwest.

D_2

D_2 folds re-fold earlier recumbent structures into upright, open and northeast-verging chevron folds. Folds have a spaced cleavage (S_2), particularly where they overprint S_1 in the zone of recumbent folding at Lefroy. D_2 folds trend NW to NNW and dip steeply southwest. Typically they are open to closed and chevron in style, but no re-folding of D_1 is observed east of the Scottsdale Batholith, where D_1 is absent (Reed, 2001*a*).

D_3

A regional southwest-directed compressional event has been recognised in the Alberton–Mathinna gold corridor (Keele, 1994). A good example of D_3 folds re-folding D_2 folds may be seen in the Golden Hinges adit at Mathinna (see below). East of the Scottsdale Batholith, D_3 faults fold D_2 veins, fractures and foliations and have preferentially developed in the steep northeast limbs of the D_2 folds (Reed, 2001*a*). In the recumbent fold zone, west of the Scottsdale Batholith, an S_3 crenulation cleavage occurs in the more pelitic units (it is rare, or absent, in the sandy layers) and has a 15–20° clockwise rotation with respect to S_1 (e.g. Lefroy goldfield). S_2 and S_3 generally share a common strike direction, hence they may not always be easy to tell apart — the major difference being top-to-NE transport for D_2 structures compared with top-to-SW transport for D_3 structures.

Timing of gold mineralisation

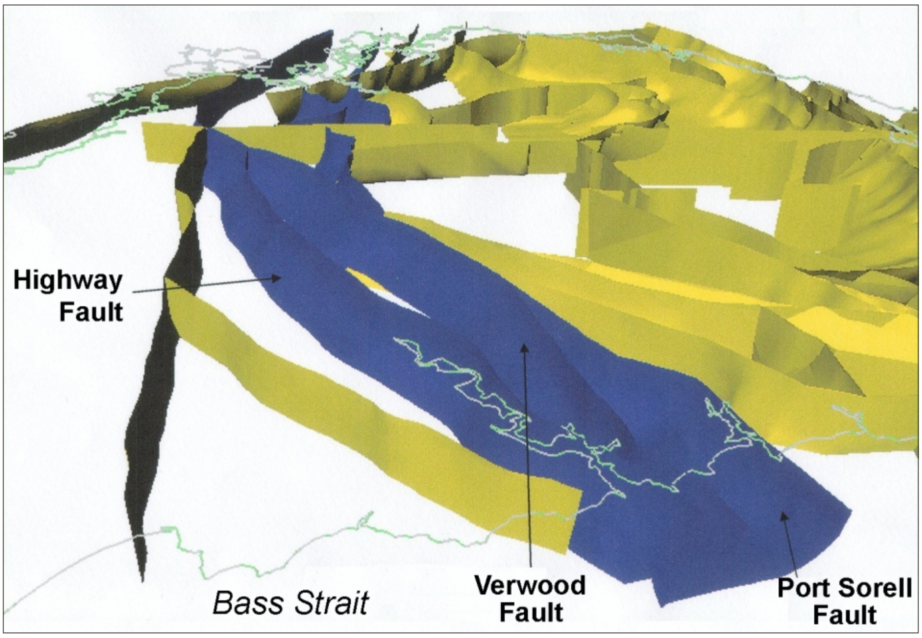
Gold mineralisation across northeast Tasmania is related to the D_3 event. D_3 quartz veins, which carry gold mineralisation at Lefroy, cut D_1/D_2 thrusts in the recumbent fold zone. The gold lodes at the Golden Gate mine (Mathinna) lie on the eastern flank of a regional D_2 anticline, whose steep northeast limb was

Faults separating eastern and western Tasmania terrains

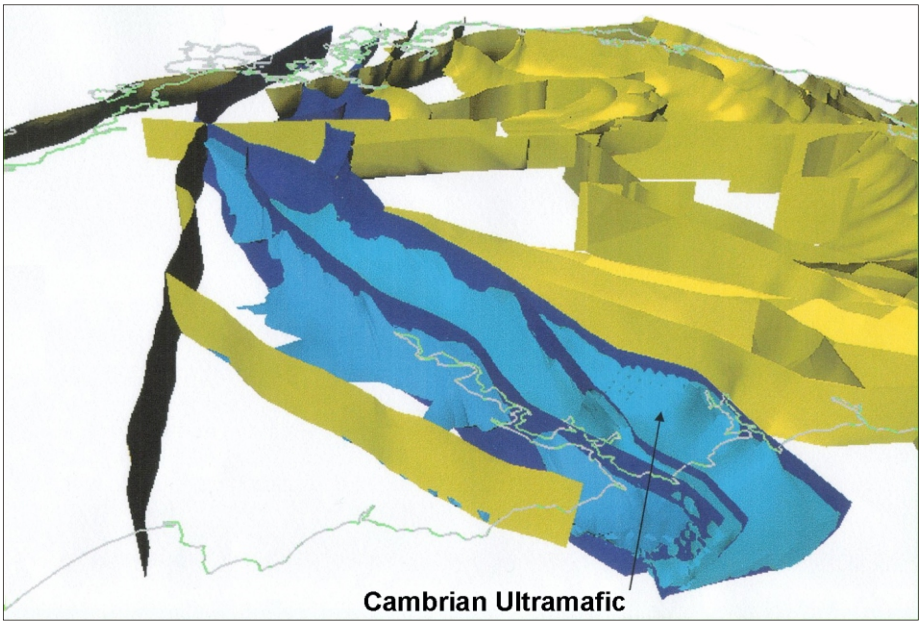
Figure 2

3D Model view from over Bass Strait looking south at the D_3 thrusts that joined eastern and western Tasmania together in the Middle Devonian.

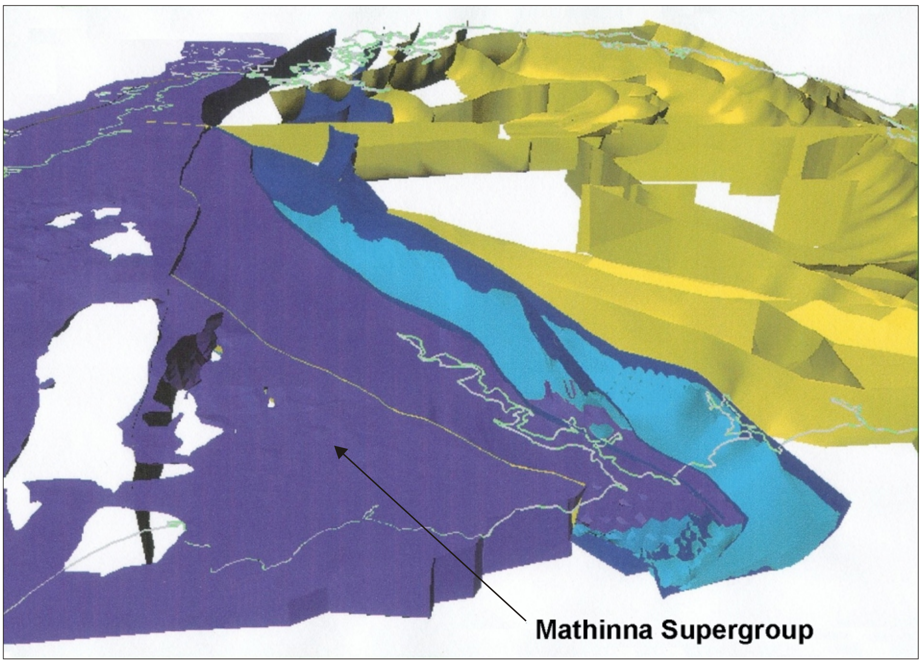
(From 3D Geological Model of Tasmania, pmd CRC, Mineral Resources Tasmania, CSIRO and Fractal Graphics).*



(a) this view shows the Verwood, Port Sorell and Highway faults.



(b) the Beaconsfield gold deposit lies on the Highway Fault, adjacent to where Cambrian ultramafic rocks come to the surface at Andersons Creek.



(c) the Mathinna Group is shown as a solid.

in a suitable orientation conducive to shear failure during D_3 (Reed, 2004). A maximum age for gold mineralisation (i.e. for the D_3 event) of 391 Ma comes from Rb-Sr and K-Ar radiometric dating of a post-D_2 granodiorite (McClenaghan and Higgins, 1993; Black, in prep.). A minimum age for mineralisation of ~389 Ma comes from Rb-Sr and K-Ar dating of an undeformed adamellite.

Lithological control on gold mineralisation

Whilst there is little evidence for a stratigraphic control on the gold deposits of northeast Tasmania, there is some evidence that many deposits tend to be hosted by mixed sandstone-shale cycles that are transitional between lower sandstone-dominated and upper shale-dominated turbiditic sequences (Reed, 2004). The deposits at Lefroy and Back Creek occur at the transition between the Stony Head Sandstone and the Turquoise Bluff Slate. These settings provide maximum ductility contrast, i.e. excellent conditions for dilation and gold mineralisation. More field work is required to establish a stratigraphic sequence east of the Scottsdale Batholith, and at Mathinna, in particular.

Tectonic evolution of the Blue Tier Batholith

The composite Devonian-aged Blue Tier Batholith was emplaced into the Mathinna Supergroup at depths ranging from 12 km to 1 km below surface. Radiometric ages, intrusive field relationships and mapped foliations within the individual granite phases all point to a protracted, episodic, intermediate to shallow crustal granite-forming event, whose unknown parent magma evolved through granodiorite, biotite-adamellite/granite to alkali-felspar granite, over a 25 Ma period (fig. 3).

A series of cartoons (fig. 4a–d) are able to demonstrate a possible sequence of events with respect to crustal extension that took place during emplacement of the Blue Tier Batholith. The broad vectors of this extension can be represented, to a first approximation, by the foliations mapped in each individual pluton by field geologists. The exception to this is the alkali-felspar (Sn-bearing) granites that may have involved a significant amount of stoping, as opposed to crustal dilation. It should be stated that such a scenario is speculative and would be unlikely to get full agreement from the many geologists who have mapped in northeast Tasmania, especially as regards the significance of the various foliations, i.e. compressional v extensional. However, the intention is to stimulate further geochemical, radiometric and geological studies that may ultimately lead to a better understanding of the processes that led to the formation of the batholith and its associated polymetallic mineral deposits.

The following order of pluton emplacement is suggested:

D_2 DEFORMATION

1. Pyengana and George River (Phase 1) pluton
2. Gardens and George River (Phase 2) pluton

D_3 DEFORMATION — 391–389 Ma (Black, in prep.)

3. Mt Pearson Pluton
4. Scamander Tier Phase, including St Marys Porphyrite, Catos Creek Dyke and Haleys New Country pluton
5. Poimena Pluton
6. Lottah and Paris Pluton

The early granodiorite phase (Pyengana/George River Pluton) was emplaced under conditions of NE–SW crustal compression (D_2), or what amounts to a NW–SE extension (E_1–A). The two next phases (pre-D_3 George River/Gardens and post-D_3 Mt Pearson Pluton) represented a change in compression direction to E–W, or approximately N–S extension (E_1–B), as D_3 peaked and began to wane. The various plutons of the Scamander Tier phase shared a weak N–S foliation trend. The final phase of extension involved the Poimena Pluton, which was emplaced broadly under conditions of NE–SW extension (E–2).

By removing the late to post-D_3 plutons from the Blue Tier Batholith and joining the remaining parts together, an interesting relationship between gold deposits and granodiorite emerges (fig. 5). This relationship suggests that the distribution of gold deposits is aureole-related (Thermal Aureole Gold deposits?), and that the gold mineralisation had commenced possibly as early as late-D_2, but certainly by D_3 (see above). Further work is required to resolve the issue of whether there is more than one gold event in the northeast, or not.

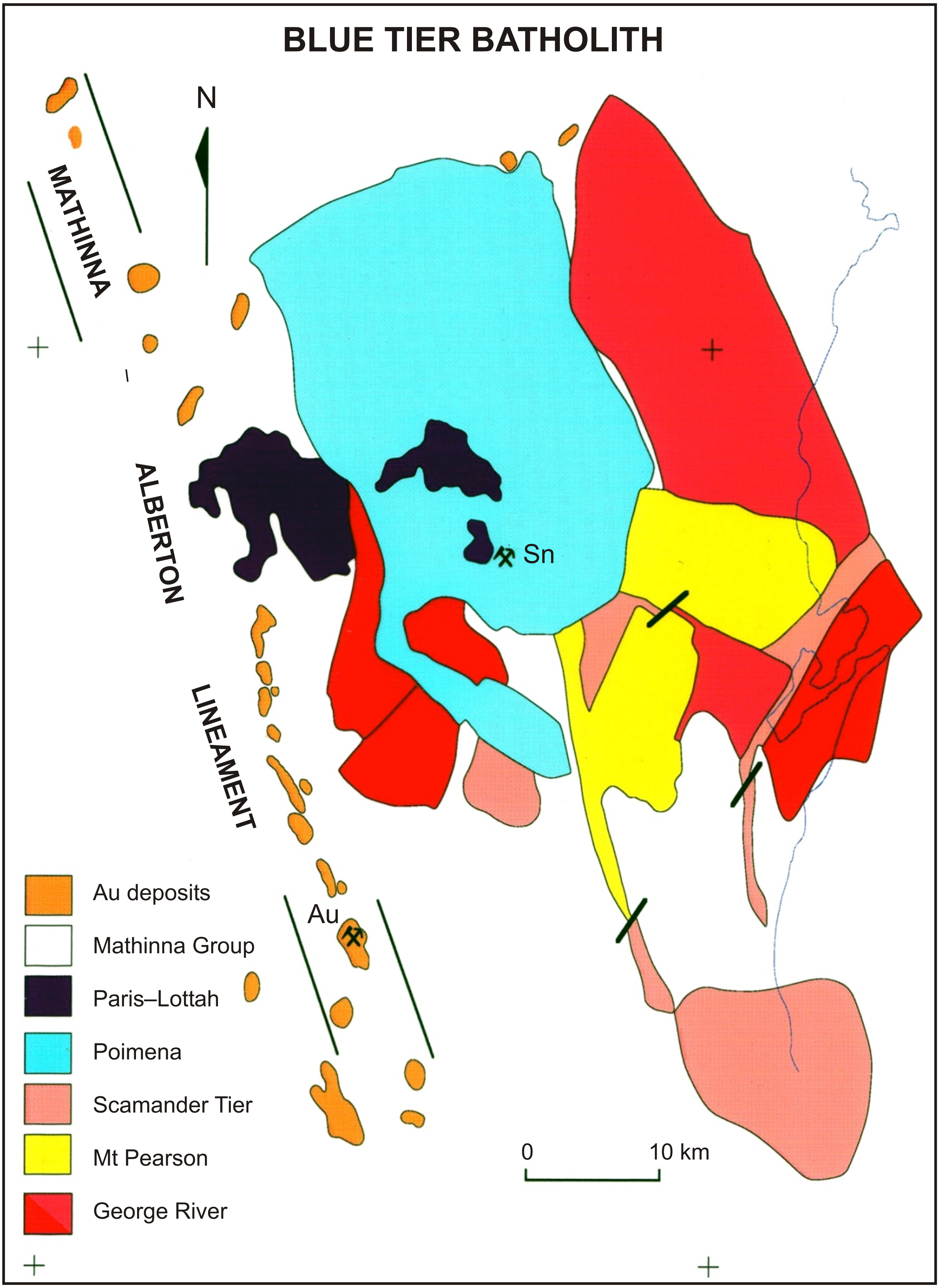


Figure 3

Phases of the Blue Tier Batholith with gold deposits on the Mathinna–Alberton Lineament.

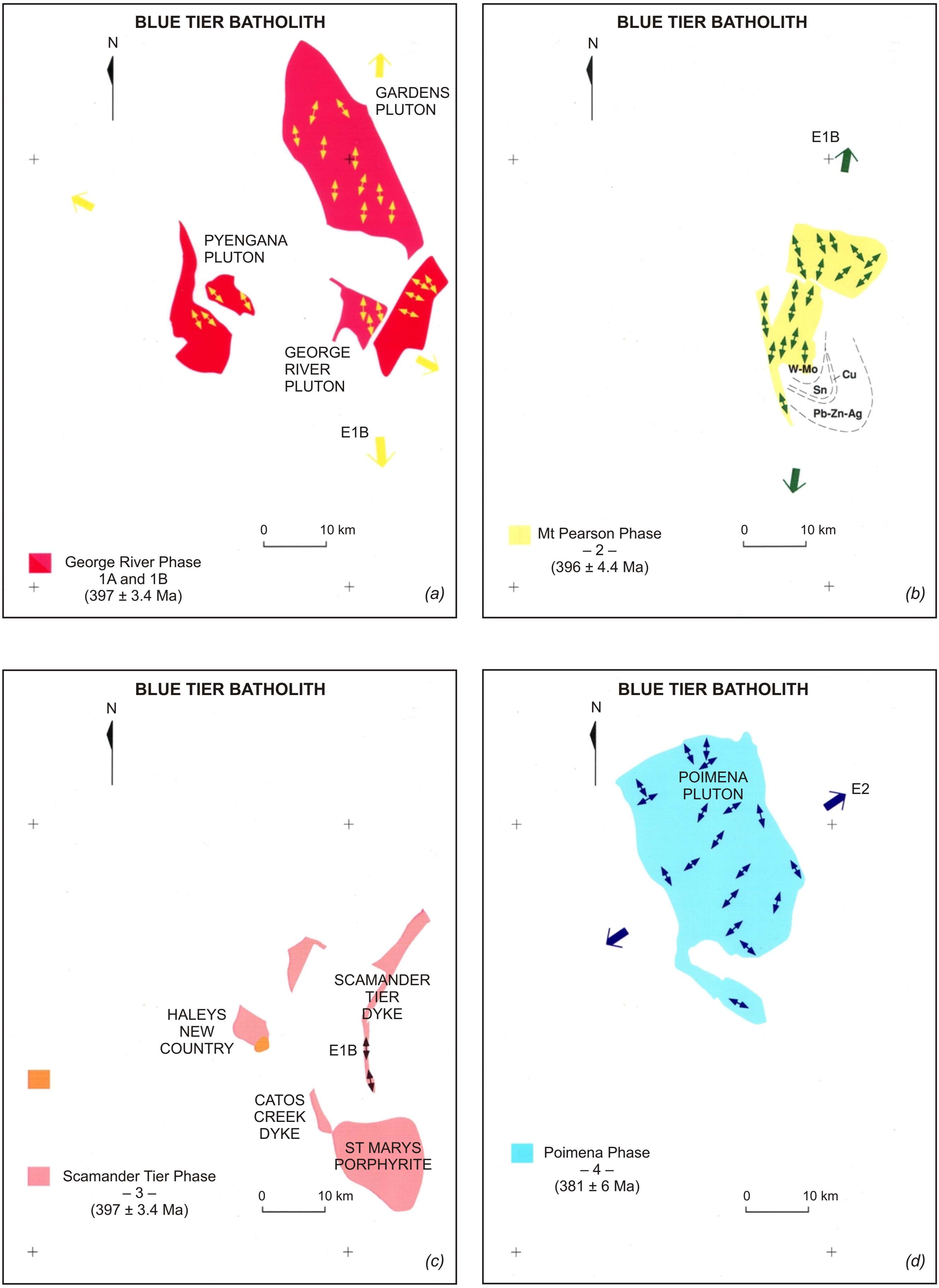


Figure 4

Foliation trends (arrows) in the phases of the Blue Tier Batholith. (a) Pre-D_3 Pyengana, George River and Gardens plutons; (b) Mt Pearson Pluton; (c) Post-D_3 Scamander Tier Dyke and related plutons; (d) Poimena Pluton. No distinction is made as to whether the foliation is due to deformation or flow, or both.

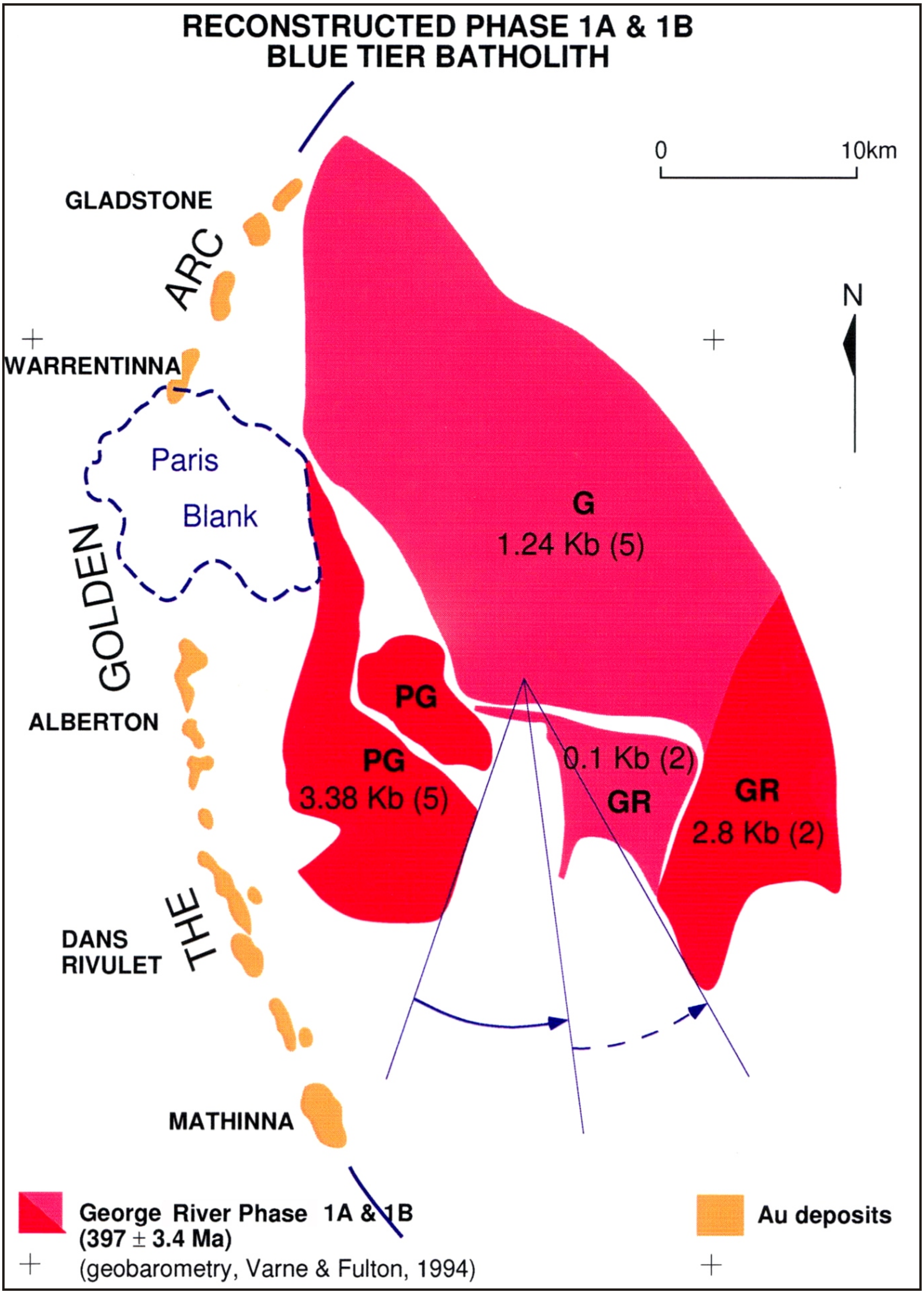


Figure 5

Reconstructed pre-D$_3$ phases of the Blue Tier Batholith. The gold deposits form an arc on the western flanks of the proto-batholith.

Western Tasmanian geology*

Western Tasmania consists of two main zones of Proterozoic rocks separated by Early Cambrian allochthonous ultramafic, mafic and sedimentary rocks and Middle to Late Cambrian volcanic and sedimentary sequences. These are overlain by Ordovician to Early Devonian shelf sequences and are intruded by pre-tectonic Late Devonian and Early Carboniferous granites, prior to a major deformational episode in the Middle Devonian. Flat-lying Late Carboniferous to Permian mainly glaciomarine sequences, Jurassic dolerite, Tertiary sedimentary rocks and basalt, and Pleistocene glacigene sediments locally drape the mineralised rocks (fig. 6).

Precambrian rocks

The oldest rocks are a Mesoproterozoic shelf quartz sandstone, pelite and minor carbonate succession that is unmetamorphosed and generally weakly deformed by open upright folds in far northwest Tasmania (Rocky Cape Group) and variably metamorphosed and strongly polydeformed, including two phases of isoclinal folding, in central Tasmania (Tyennan region). There is also a turbidite unit in central western Tasmania (Burnie and Oonah formations) that may be of partly equivalent age. The upper part of these formations contains minor dolomite and basalt, possibly reflecting a shallower marine depositional environment. The lower age of these rocks is constrained by U-Pb dating on detrital zircon, which show similar age spectra, with the lowest age peak at 1200 Ma (Black *et al*., 1997). A date of 725 35 Ma on a syn-sedimentary dolerite sill in the Burnie Formation is the only other age constraint.

The quartzose rocks are overlain with low angle unconformity in northwest Tasmania by the Togari Group, a sequence of thin basal quartz conglomerate, dolomite, mudstone, diamictite, volcaniclastic sandstone and conglomerate, tholeiitic and alkalic volcanic and intrusive rocks, and an upper dolomite unit (Everard *et al*., 2001). The contact has long been considered to mark a major phase of compressive deformation (Penguin Orogeny), but more probably represents block rotation during extensional deformation accompanying Rocky Cape Group sedimentation (Everard *et al*., 2001). The Togari Group is Neoproterozoic in age based on C and Sr isotope chemostratigraphic ages of c. 750–650 Ma for the lower dolomite and c. 580–545 Ma for the upper dolomite (Calver, 1998). These rocks were also deposited in an extensional setting, as shown by thickness variations in the rock units and geochemical characteristics of the igneous rocks consistent with magmatism and lithospheric thinning induced by a rising mantle plume (Everard *et al*., 2001) during a major phase of rifting (Crawford and Berry, 1992). Near Renison Bell, the lower part of the Togari Group equivalent is the Success Creek Group, comprising siltstone, dolomite horizons and conglomerate, which has been considered to overlie the Oonah Formation unconformably (Brown, 1986). Significant magnesite and magnetite deposits and potential for Cu-Au deposits are associated with the Bowry Formation (see below).

Middle Cambrian Tyennan Orogeny and allochthonous rocks

A collisional event emplaced ultramafic and mafic intrusive rocks, boninitic and low-Ti andesitic lavas and allochthonous sedimentary rocks from the north and east during the Tyennan Orogeny in the Middle Cambrian (Berry and Crawford, 1988; Turner *et al*., 1998), although the collisional event was more complex and probably involved at least two subduction zones (Reed *et al*., 2002). The event is correlated with the Ross Orogeny of Antarctica and the Delamerian Orogeny in South Australia. The collision was also responsible for the main folding and thrusting events in the Tyennan region and emplacement of high-grade metamorphic rocks, which include eclogite dated at 502 8 Ma (Turner *et al*., 1998) and a garnet amphibolite at Forth in northern Tasmania dated at 514.1 4.6 Ma (Black *et al*., 1997). The lavas were probably emplaced in an oceanic forearc setting (Berry and Crawford, 1988; Brown and Jenner, 1989). A tonalite, intruded apparently late in the history of one of the ultramafic complexes, has been dated at 513.6 5.0 Ma (Black *et al*., 1997). The ultramafic rocks have shed sediments that were major world producers of Os-Ir-Ru alloys in the 1930s.

The Arthur Lineament is a NNE-trending belt of strongly deformed fault-bound slices of Tyennan age rocks in northwest Tasmania, transitional into a correlate of the Togari Group on the western side and the Burnie and Oonah formations to the east. The belt consists of basaltic amphibolite and greenschist, and siliciclastic and dolomitic metasedimentary rocks. The eastern zone of amphibolite and surrounding rocks (Bowry Formation) is distinctive in having a higher grade of metamorphism, including relict blueschist with glaucophane developed some distance from its present location, and rift tholeiite geochemistry slightly different to that of the Togari Group (Holm *et al*., 2003). The rocks have subsequently been mostly regressed to amphibolite and greenschist facies assemblages. The Bowry Formation is intruded by granite dated at 777 7 Ma (Turner *et al*., 1998) and regarded as a fractionate of the tholeiites with some crustal contamination. It is considered to be allochthonous and may represent a correlate of the 780 Ma Willouran Hood basalts of South Australia that may represent the breakup of the Rodinian supercontinent (Holm *et al*., 2003). The Bowry Formation is unique in hosting significant stratiform

* From Green in Green and Taheri (2004)

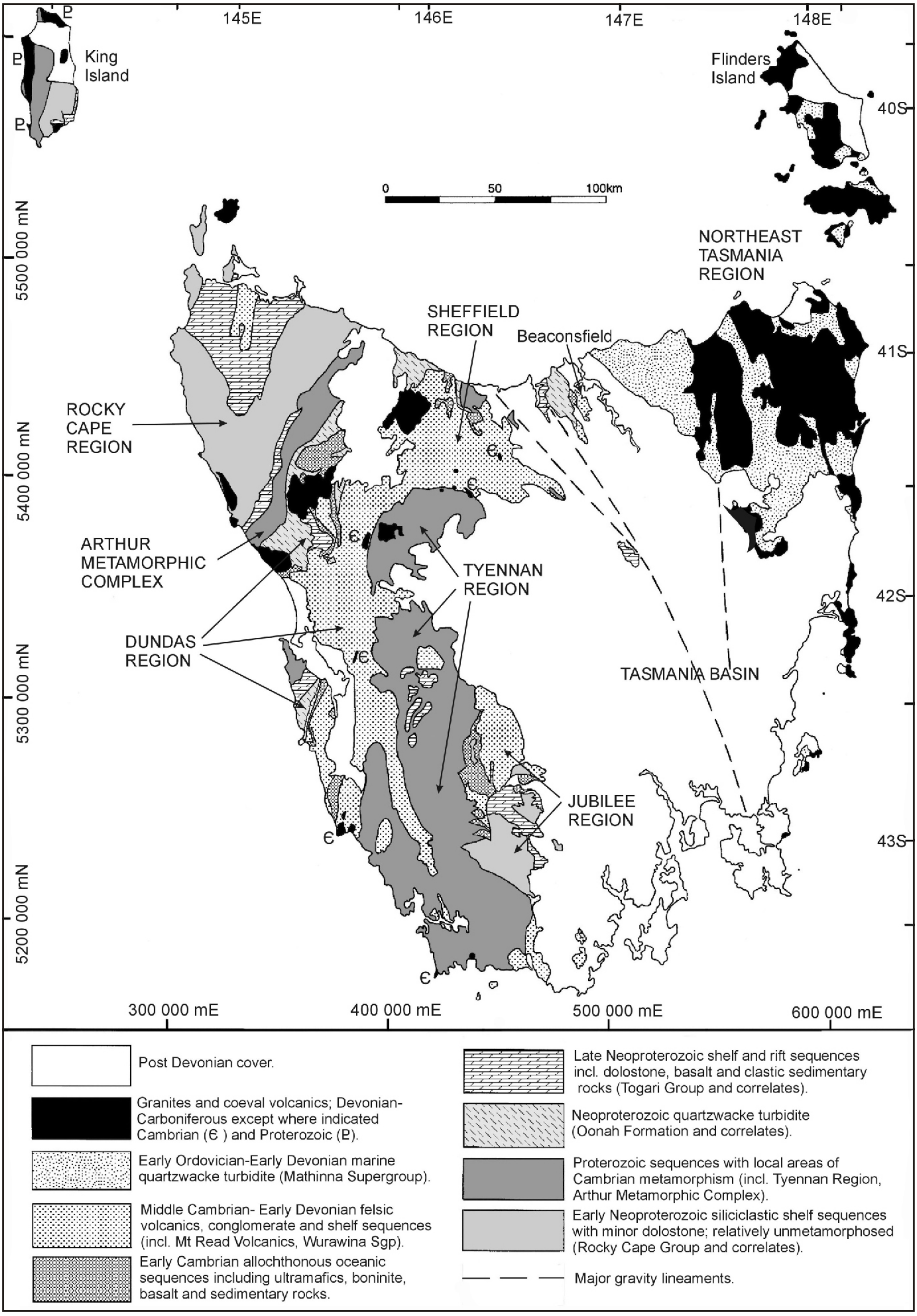


Figure 6. *Simplified geology of Tasmania, with tectonic regions shown.*

magnetite deposits, including Savage River, and the major magnesite deposits of Arthur River, Lyons River and Main Creek.

The restriction of the significant deposits to the Bowry Formation, and general acceptance by most workers that the major deposits of magnetite and magnesite are pre-tectonic, is consistent with the regional geological interpretation.

Mount Read Volcanics and coeval Cambrian sedimentary rocks

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 late Middle Cambrian to 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.

The main mineralised belt of the MRV between Mount Darwin and Hellyer is the Central Volcanic Complex (CVC), 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; 2003; Gifkins, 2003). This belt is flanked to the west by the coeval Western Volcano-Sedimentary Sequence (WVSS; Corbett, 2003) of lithicwacke, 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, 2003; 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. 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 (2003) considers that it is a time equivalent of the CVC, while Murphy *et al*. (2004) consider that it could possibly be part of the Tyndall Group.

Tectonism was mostly extensional during Mount Read Volcanics deposition, with accompanying growth faulting and cauldron subsidence associated with the formation of the thick pumice-rich breccia underlying the Rosebery deposit (Green *et al*., 1981; Gifkins, 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 restricted to the proximal CVC volcanic facies (Gifkins, 2003; Herrmann and Kimber, 2003).

Ore fluids have been considered to have been derived by convective circulation of seawater and interaction with volcanic and basement rocks (see Solomon and Groves, 2000), but there is evidence for a magmatic contribution to the ore fluids at Mount Lyell (Large *et al*., 1996) and at Hellyer (Solomon and Groves, 2000).

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.

Late Cambrian siliciclastic molasse: Owen Conglomerate and correlates

The Tyndall Group is overlain by a fault-controlled unit of Precambrian and mixed Precambrian-volcanic derived conglomerate, sandstone and minor mudstone. In the region abutting the Tyennan Precambrian region, deposition of the Owen Conglomerate took place in an active half graben, resulting in a westward thickening sequence of up to 2000 m thick ponded against the Great Lyell Fault. The Owen Conglomerate was deposited in a variety of settings from alluvial slopes to marine shelf to submarine fans (Noll and Hall, 2003). Further west equivalent units are turbidite sandstone, conglomerate and mudstone (Seymour and Calver, 1995).

Shelf sequences

The Owen Conglomerate is overlain unconformably by Ordovician peritidal limestone and dolomite with a thin quartz sandstone unit at the base. The limestone is succeeded by Silurian to Early Devonian shelf sandstone, mudstone and minor limestone. The Ordovician limestone hosts stratabound and cross-cutting Zn-Pb-Ag mineralisation in a number of localities. Mineralisation at the largest deposit, the Oceana mine near Zeehan, is regarded as part epigenetic, part exhalative (Taylor, 1989).

Devonian deformation

A major period of faulting and folding, the Tabberabberan Orogeny, affected Tasmania in the Middle Devonian. In central western Tasmania, the earliest phases of deformation produced broad north-trending folds and faults, some of which reactivated earlier structures (e.g. Great Lyell Fault). The second phase involved NW to WNW-trending faulting and minor folding that was concentrated along discrete zones that may extend for over 100 kilometres. The Mount Lyell area coincides with one of these, the Linda Fault Zone. Some of the faults are reactivated structures dating back at least to the Late Cambrian.

Orogenic gold mineralisation accompanied deformation and is most important in northeast Tasmania, where it accompanied the last major phase of deformation, southwest-directed thrusting (Reed, 2002). Small orogenic gold deposits occur in western Tasmania in rocks as young as Silurian.

Devonian granite emplacement and mineralisation

Western Tasmania is notable for the wide variety of significant mineral deposits associated with Late Devonian to Early Carboniferous granite emplacement, a feature related to the frequency of strongly fractionated granites, syn-intrusive structures to focus fluids, high level of intrusion and variety of host rocks, particularly carbonate units. Granites were emplaced from about 370 to 330 Ma, generally younger than, but overlapping with those in eastern Tasmania, where there is a higher proportion of granodiorites (McClenaghan, 1989). Both western and eastern Tasmania are richly endowed in Sn and W deposits, but the apparently lower degree of unroofing and greater variety of host rocks in western Tasmania has resulted in a greater diversity and economic importance of deposits there.

World-class scheelite mineralisation is associated with weakly fractionated magnetite series biotite-hornblende monzogranite and granodiorite on King Island and replaces dolomite in Togari Group correlates (Wesolowski *et al*., 1988) and a smaller magnetite-scheelite skarn with a magnetite series granite at Kara. This is mined mainly for magnetite for use as a heavy medium in coal washing. There may be insignificant endogranitic tin mineralisation in some associated granites and minor Pb-Ag-Zn in country rock haloes, but these deposits have been of no economic significance.

Tin deposits are invariably associated with fractionated (high Li, Rb, U, Sn, Be, K etc.) ilmenite series granites with high volatile contents (F, B etc.), depleted in Ca, Sr, Mg, Fe and Ba. High Fe/Mg, Rb/K and Rb/Ba ratios and tourmalinisation of the associated granites and immediate country rocks are characteristic.

Economic granite-related tin mineralisation is also typically surrounded by recognised haloes of base metal vein deposits, mostly Ag-Pb-Zn but also Cu, with the argentiferous deposits most proximal to tin deposits (Solomon and Groves, 2000). The historically important Ag-Pb-Zn deposits of the Zeehan and Dundas districts provide type examples. The vein deposits are typically high grade, but are mostly small. Mining of the largest deposit, at Magnet, produced 630 000 tonnes of ore grading 7.3% Zn, 7.3% Pb and 427 g/t Ag.

Skarn deposits, hybrid skarn-vein and/or greisen deposits, and possibly other metasomatic deposits form the most important granite-related resources. The economically important tin deposits were mainly sulphide skarns where tin is present as cassiterite (Renison Bell, Mount Bischoff, Cleveland, Queen Hill–Montana, Razorback). There are also significant tin silicate skarns (St Dizier), but these are not effectively treatable.

Skarns have the potential to host economically significant base metal (Pb-Zn-Ag) deposits in western Tasmania. An example of the latter is the Sylvester base metal skarn, where massive sulphide mineralisation at Sylvester has been intersected over a one kilometre strike length and remains open at depth over its entire length. The current inferred resource is 5.1 million tonnes at 4% zinc, 2.3% lead and 30 g/t silver. The potential for larger deposits of possibly higher grade exists at the eastern end of this deposit where the mineralisation has replaced a dolomite unit.

Throughout the West Coast there is potential for the discovery of more base metal skarn deposits where carbonate sequences are intruded by, or overlie the upper contact of, the subsurface continuation of the associated granite.

Skarn deposits also contain other commodities of economic significance, such as magnetite, gold, bismuth, boron, fluorite and copper. The Moina skarn, with 26 million tonnes of 18% CaF_2, is Australia's largest fluorite resource. Industrial-grade magnetite is the major commodity mined at Kara, while at Moina, secondary alteration of the magnetite-fluorite-vesuvianite skarn adjacent to a fault cutting the deposit has produced an amphibole-hematite-quartz-sulphide-fluorite assemblage with high Zn and Au and significant In, Cd and Cu. The Stormont skarn deposit is characterised by Au-Bi-Pb mineralisation which is related to late-stage chlorite-fluorite-muscovite-calcite-quartz alteration of a distal garnet-pyroxene-vesuvianite skarn which has been extensively retrogressed to actinolite prior to the mineralising event.

Skarn mineral assemblages are developed at depth at Cleveland where they are centred on a quartz-feldspar porphyry plug, grading upwards and outwards through different hydrothermal mineral assemblages to unaltered limestone. Mineralisation is spatially zoned, with cassiterite-chalcopyrite forming the central zone, within or immediately above the intrusive rock, and passes through wolframite to an outer shell of molybdenite, bismuthinite and native bismuth. Other minerals associated with the mineralisation include quartz, fluorite, sericite, topaz, tourmaline, siderite, arsenopyrite, pyrite and sphalerite.

In the 1980s a drill hole intersected two major zones of copper-tungsten-tin bearing skarns west of Renison Bell. The upper skarn zone was typically very low in sulphide content but the lower skarn zone was considerably higher, with pyrrhotite and chalcopyrite occurring as veins, disseminations, stringers, and semi-massive bands up to 50% pyrrhotite and 5% chalcopyrite. The drill hole intersected a sulphide-rich, tin-bearing granite at depth.

Devonian epigenetic nickel mineralisation occurs in the southern aureole of the Heemskirk Granite at the Avebury, Burbank and Nickel Reward deposits, where ultramafic rocks overlie concealed Devonian mineralised granite. There are many other areas in western Tasmania that satisfy this condition, particularly adjacent to the Meredith Granite. These may also be prospective for hydrothermal nickel mineralisation.

There is also potential for low-grade intrusion-related disseminated gold deposits in Tasmania. The Fire Tower prospect in central northern Tasmania occurs in quartz stockwork veinlets in intensively sericite-carbonate altered volcaniclastic rocks of the Tyndall Group and is associated with minor arsenopyrite, chalcopyrite, sphalerite and galena (Callaghan, 2002). The fine quartz stockwork (most veinlets are about 1 mm thick) is unstrained and scheelite is disseminated in carbonate haloes around the randomly orientated veinlets. Geophysical modelling suggests that Devonian granite underlies the area at a depth of about two kilometres at this locality (Leaman and Richardson, 2003).

Devonian Granites*

Introduction

The Tasmanian mid-Palaeozoic granites (fig. 7, 7*a*, 8) were emplaced at high crustal level with narrow contact aureoles mostly after the main Tabberabberan deformation. In eastern Tasmania, the granites intruded an Ordovician to Early Devonian (Banks and Smith, 1968; Rickards and Banks, 1979; Rickards *et al*., 1993) quartzwacke turbidite sequence (Mathinna Supergroup) whereas in western Tasmania they intruded more variable successions ranging in age from Precambrian to Early Devonian. In eastern Tasmania, intrusion took place in the Devonian whereas in western Tasmania intrusion was generally later and took place in the Devonian and Early Carboniferous. Intrusion appears to have occurred mainly by upward displacement or crustal rifting, although there is local, marginal evidence for 'shouldering-aside' of the country rock (Gee and Groves, 1971). The granites generally range from granodiorite to alkali-feldspar granite, with granodiorite being much more abundant in eastern Tasmania. The classification used is that of Streckeisen (1973) with the modification that the term adamellite is used for part of the 'granite' field with over 35% modal plagioclase. The term granite will be used to refer to granitic rocks in the wide sense except where specifically discussing composition. Rock types used in body names will be used rather broadly to remain consistent with previous usage.

Granites of eastern Tasmania

Introduction

The following notes on the granites of northeast Tasmania been drawn from McClenaghan (in prep.).

The granites of mainland eastern Tasmania occupy an area of ~3120 km^2 and form the Scottsdale, Ben Lomond, Blue Tier and Eddystone batholiths (fig. 7*a*, 8). A number of disconnected lesser bodies extend south as far as Deep Glen Bay on the Forestier Peninsula and the Hippolyte Rocks a short distance to the east of the peninsula (fig. 7). Granites also occur in the Furneaux Group to the northeast of mainland Tasmania where they form about 70% (~1090 km^2) of the Palaeozoic basement. All the granites of eastern Tasmania and the Furneaux Group, together with the granites of Wilsons Promontory in Victoria, have been depicted as constituting the Bassian Batholith by Chappell *et al*. (1991) in order to stress the essential unity of the granites across Bass Strait. The granites may generally be divided into hornblende-biotite granodiorites, biotite adamellites, garnet-cordierite-biotite adamellites, and alkali-feldspar granites. Description of the granites will be largely confined to those on mainland eastern Tasmania, although geochemical plots include data from the Furneaux Group.

Hornblende-biotite granodiorite

Hornblende-biotite granodiorites occur throughout eastern Tasmania with the most southerly body on the Freycinet Peninsula (fig. 7). Isotopic dating and field relationships show that they are generally the earliest intrusive phase. Most are distinguished from later phases by having a tectonic foliation parallel to a foliation in the Mathinna Supergroup country rocks.

The granodiorites show little variation. They are massive, medium to coarse-grained, dark grey rocks, commonly with abundant mafic fine-grained dioritic enclaves. Major minerals are euhedral to anhedral amphibole and biotite, plagioclase and intergranular K-feldspar and quartz. Amphibole ranges in composition from actinolite in the core regions of some crystals to hornblende in the rims and in euhedral crystals. Amphibole and biotite are commonly present in intermingled clusters of crystals. Trace amounts of clinopyroxene are associated with actinolite in the cores of hornblende crystals in some of the granodiorite bodies. Plagioclase commonly contains sericitised unzoned calcium-rich core regions (An_{80-70}) which have sharp boundaries with the clear, zoned rims (An_{60-25}). Accessory minerals include apatite, zircon, sphene, allanite and ilmenite. Variable amounts of magnetite and ilmenite occur in the Pyengana and Lisle plutons.

Early, foliated, biotite-hornblende granodiorite plutons in the Scottsdale Batholith are the Diddleum, Tulendeena and Porcupine Creek bodies. The Upper Blessington body in the south of the Scottsdale Batholith is slightly younger, non-foliated, contains less hornblende and is coarser-grained than the other granodiorite bodies. The Lisle body, a short distance to the west of the Scottsdale Batholith, ranges in composition from granodiorite to tonalite (Roach, 1994; Bottrill, 1996).

The main early, foliated, biotite-hornblende granodiorite plutons in the Blue Tier Batholith are the Pyengana, Gardens, George River and Long Point bodies. The Akaroa and Grants Point bodies, in the St Helens area, may have a transitional boundary with the George River pluton but have very little or no hornblende, and range from granodiorite to adamellite.

In the Blue Tier Batholith, later, non-foliated granodiorite dykes in the Scamander Tier and Catos Creek areas are considered to be the intrusive equivalents of the dacitic, welded, ash-flow tuff that occurs in the St Marys area (Turner *et al*., 1986). This tuff consists of plagioclase, quartz, biotite, augite, hypersthene and sanidine phenocrysts in an aphanitic groundmass. The pyroxene is pseudomorphed by amphibole at higher levels of the body and in the hypabyssal equivalent micro-granodiorite body; both

* From McClenaghan (in prep.)

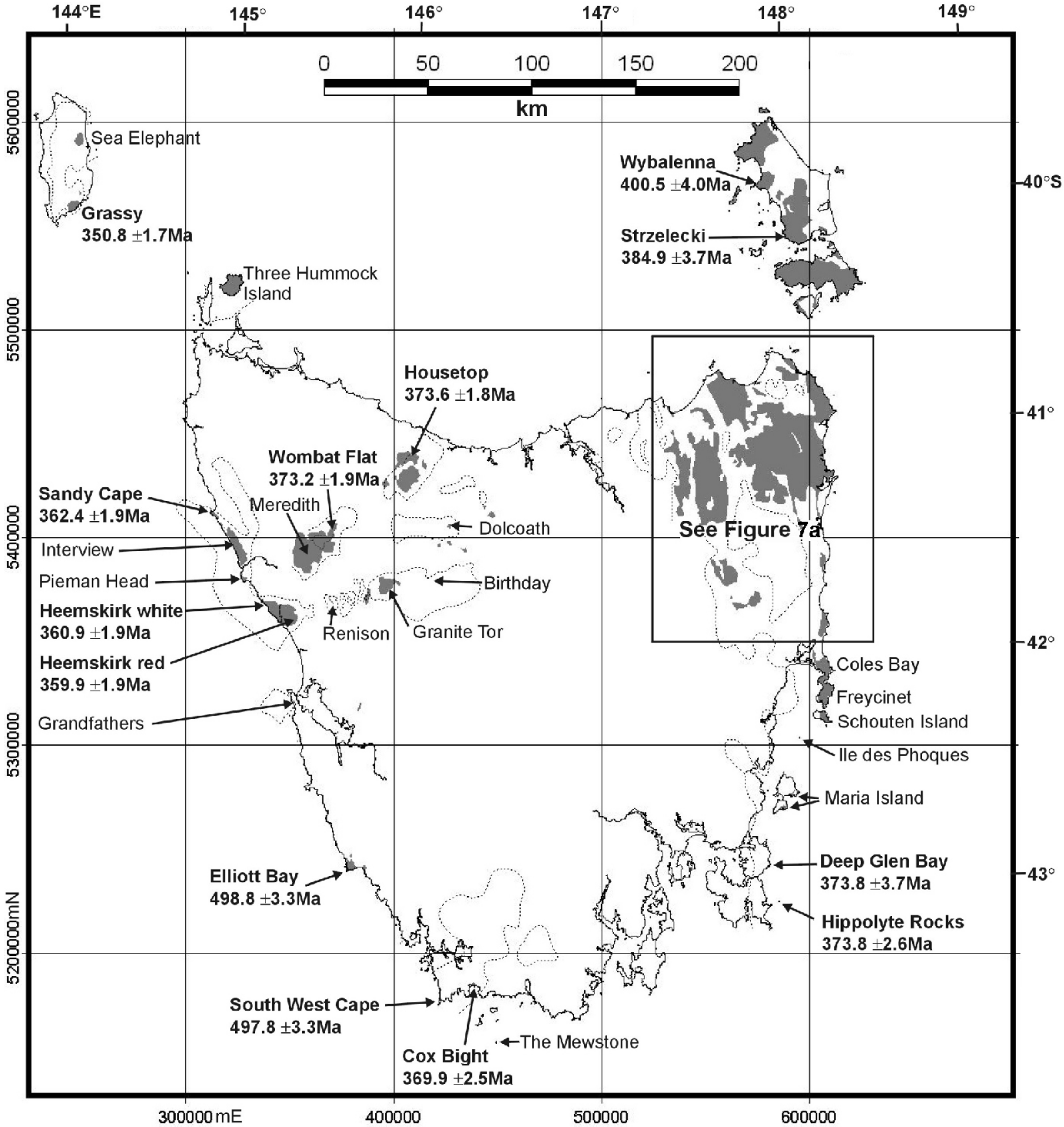


Figure 7

Location map for Tasmanian granites. Details of granites in northeast Tasmania are shown in Figure 7a.

orthopyroxene and clinopyroxene remnants are found in amphibole cores.

The proportion of granites containing biotite and hornblende decreases from west to east in eastern Tasmania; it is nearly half for the Scottsdale Batholith, less than a third for the Blue Tier Batholith, and still less in the Furneaux Group and Freycinet Peninsula.

Biotite adamellite/granite

Biotite adamellite/granites are the most common granite type in eastern Tasmania. In the Scottsdale Batholith, they locally grade into biotite-hornblende granodiorite and in some cases into alkali-feldspar granite. Elsewhere in eastern Tasmania, gradations to alkali-feldspar granite are more common than gradations to granodiorite. The adamellites/granites are younger than the foliated granodiorites.

Although of varied texture, all biotite adamellite/granites consist of plagioclase, biotite, quartz and K-feldspar, with accessory zircon, monazite, ilmenite and apatite. The plagioclase is zoned, and has relatively Ca-rich cores and thin albite rims. K-feldspar is coarsely perthitic. Megacrysts (20–60 mm) of this mineral generally contain up to five concentric zones of plagioclase inclusions (McClenaghan and Williams, 1982).

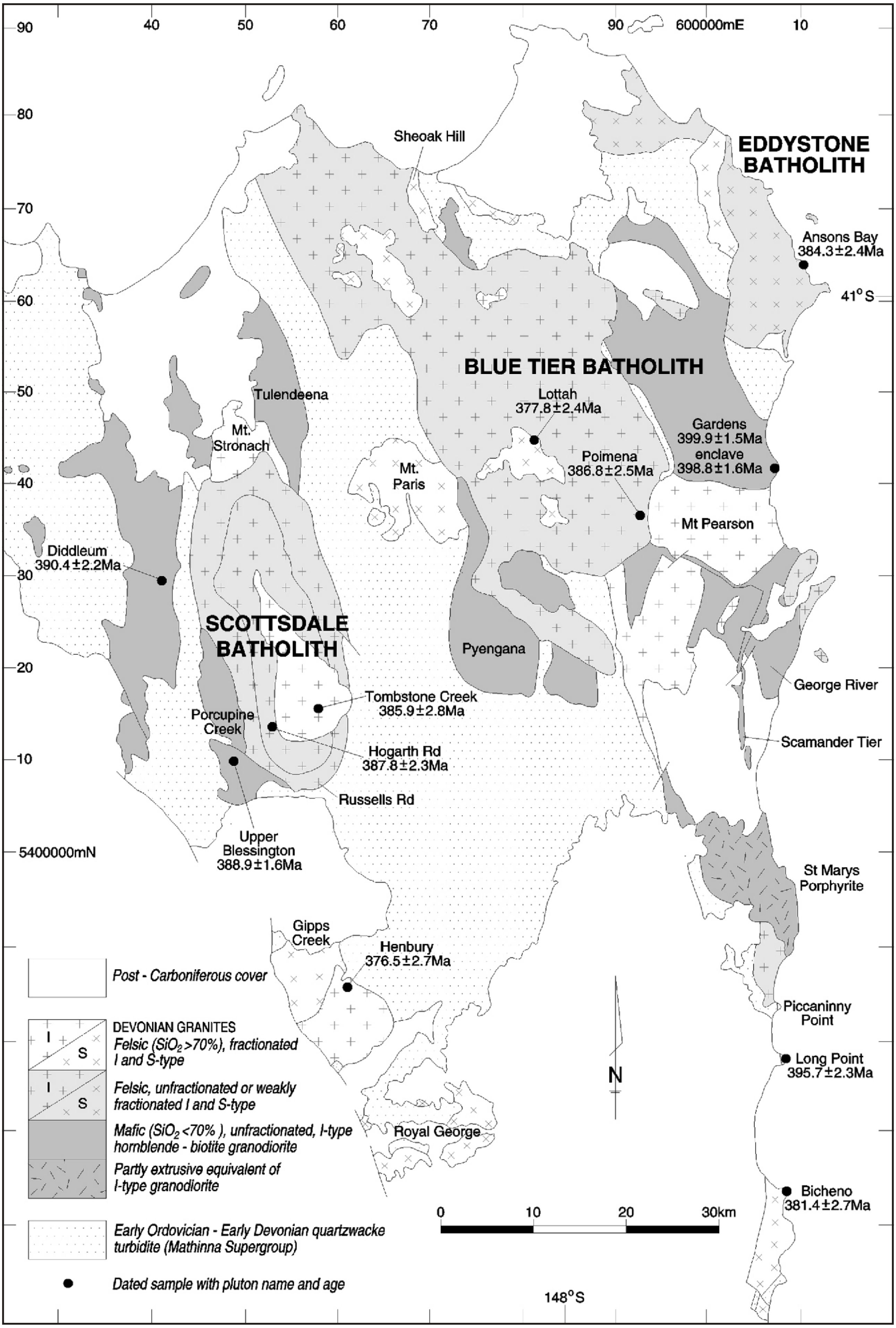


Figure 7a. *Detail of northeast Tasmanian Devonian granites.*

The Hogarth Road and Russells Road bodies of the Scottsdale Batholith are coarse to very coarse-grained equigranular to sparsely porphyritic, largely adamellite with megacrysts of K-feldspar (20–50 mm). In the southern part of the bodies they grade to granodiorite with minor amounts of hornblende in addition to biotite.

The main biotite adamellite-granite bodies in the Blue Tier Batholith are the Poimena and Mt Pearson plutons. The Poimena Pluton is a very large medium to coarse-grained body with abundant K-feldspar megacrysts (20–50 mm). Several small sheet and dome-shaped intrusions of similar composition but different texture within the Poimena Pluton, in the central Blue Tier area near Lottah, are later phases (McClenaghan and Williams, 1982). A small difference in chemical composition between granite in the outer and the central parts of the body suggests that the intrusion may be composite. The Mt Pearson Pluton is compositionally similar to the Poimena body but is very coarse grained with abundant K-feldspar megacrysts (20–60 mm). At the southern margins of the Mt Pearson body, a short distance west of St Helens, the granite is pink, and ranges to alkali-feldspar granite in composition. Here it is intruded by irregular-shaped bodies of fine to medium-grained granite.

The Haleys New Country Pluton, in the southern part of the Blue Tier Batholith, is similar in texture to the Poimena body, but is slightly darker and contains minor amounts of hornblende in addition to biotite.

Garnet-cordierite-biotite adamellite/granite

Adamellite/granite plutons containing biotite and minor amounts of garnet and/or cordierite occur in mainland eastern Tasmania in the Boobyalla, Musselroe Point, Ansons Bay, Bicheno and Maria Island bodies (fig. 7*a*, 8). The garnet-cordierite-biotite adamellite/granite plutons consist of quartz, K-feldspar, plagioclase and red-brown biotite. Garnet and cordierite rarely exceed 0.5% and are commonly less than 0.01% (Cocker, 1977). Other accessory minerals are zircon, apatite, ilmenite, monazite and xenotime. K-feldspar is perthitic, and the plagioclase has generally uniform cores ($An_{40–45}$), with normal to oscillatory zoned rims ($An_{20–10}$). Subsolidus phases are tourmaline, white-mica, pale green biotite, chlorite, topaz and andalusite (Cocker, 1977). Replacement of the garnet and cordierite by pale green biotite is widespread. Mafic inclusions are generally rare and occur near the granite/country rock contacts. The inclusions mostly consist of biotite-bearing hornfelsed rock similar to the contact metamorphosed country rocks.

Coarse-grained equigranular to porphyritic granite variants are present. Many of the bodies have complex textural variations with gradational and sharp boundaries between textural types suggesting auto-intrusive relationships. In several plutons the principal textural variation is the proportion of K-feldspar megacrysts.

Alkali-feldspar granite

Throughout eastern Tasmania there are granite bodies that range in composition from adamellite to alkali-feldspar granite but are dominantly of the latter composition, and will be described as such (fig. 8). These bodies generally show evidence of hydrothermal alteration.

There are two alkali-feldspar granite plutons in the Scottsdale Batholith, Mt Stronach to the east of Scottsdale and Tombstone Creek further south near the eastern margin of the batholith. These bodies are light coloured, pink, equigranular, coarse to fine-grained, and consist of quartz, perthitic K-feldspar, plagioclase and iron-rich biotite. In the Mt Stronach Pluton the plagioclase is close to pure albite and the biotite is annite. The Tombstone Creek Pluton has a stronger pink colouration at the margins where the plagioclase is close to pure albite. The interior of the body is adamellite and the plagioclase ranges to oligoclase composition.

The Ben Lomond Batholith lies to the south of the Scottsdale Batholith, and is separated from it by younger rocks (fig. 8). There is considerable variation in texture but the most abundant rock type is a coarse-grained, pink, porphyritic granite with K-feldspar megacrysts (Blissett, 1959). The granite consists of quartz, K-feldspar and albite, with minor biotite, muscovite and accessory tourmaline and zircon. The main rock type is intruded by irregular dykes of pale grey microgranite, which is commonly porphyritic and greisenised with accompanying tin, tungsten and sulphide minerals (Blissett, 1959). The Royal George Granite lies a short distance to the south of the Ben Lomond Batholith (fig. 8) and consists of a generally equigranular, coarse-grained granite with minor bodies of microgranite and granite porphyry (Beattie, 1967). The main granite type is very similar to that in the nearby Ben Lomond Batholith, with topaz and fluorite as additional accessory minerals (Beattie, 1967).

The two main alkali-feldspar bodies in the central part of the Blue Tier Batholith are the Lottah and Mt Paris plutons (fig. 7*a*, 8). These are composed of pale pink to cream-coloured, equigranular to porphryritic granite with K-feldspar phenocrysts. This rock consists of quartz, K-feldspar, albite and biotite with accessory apatite, zircon and monazite, and secondary muscovite. Topaz, fluorite, cassiterite and tourmaline are rare (McClenaghan and Williams, 1982; MacKenzie *et al*., 1988). Other smaller bodies of similar character occur at Little Mt Horror and Mt Cameron. These bodies, together with the Lottah Pluton, have been described as having a sheet-like form (Gee and Groves, 1971) but in the case of the Lottah Pluton a steep-sided dome is more likely (McClenaghan and Williams, 1982). The evidence for a sheet-like form for

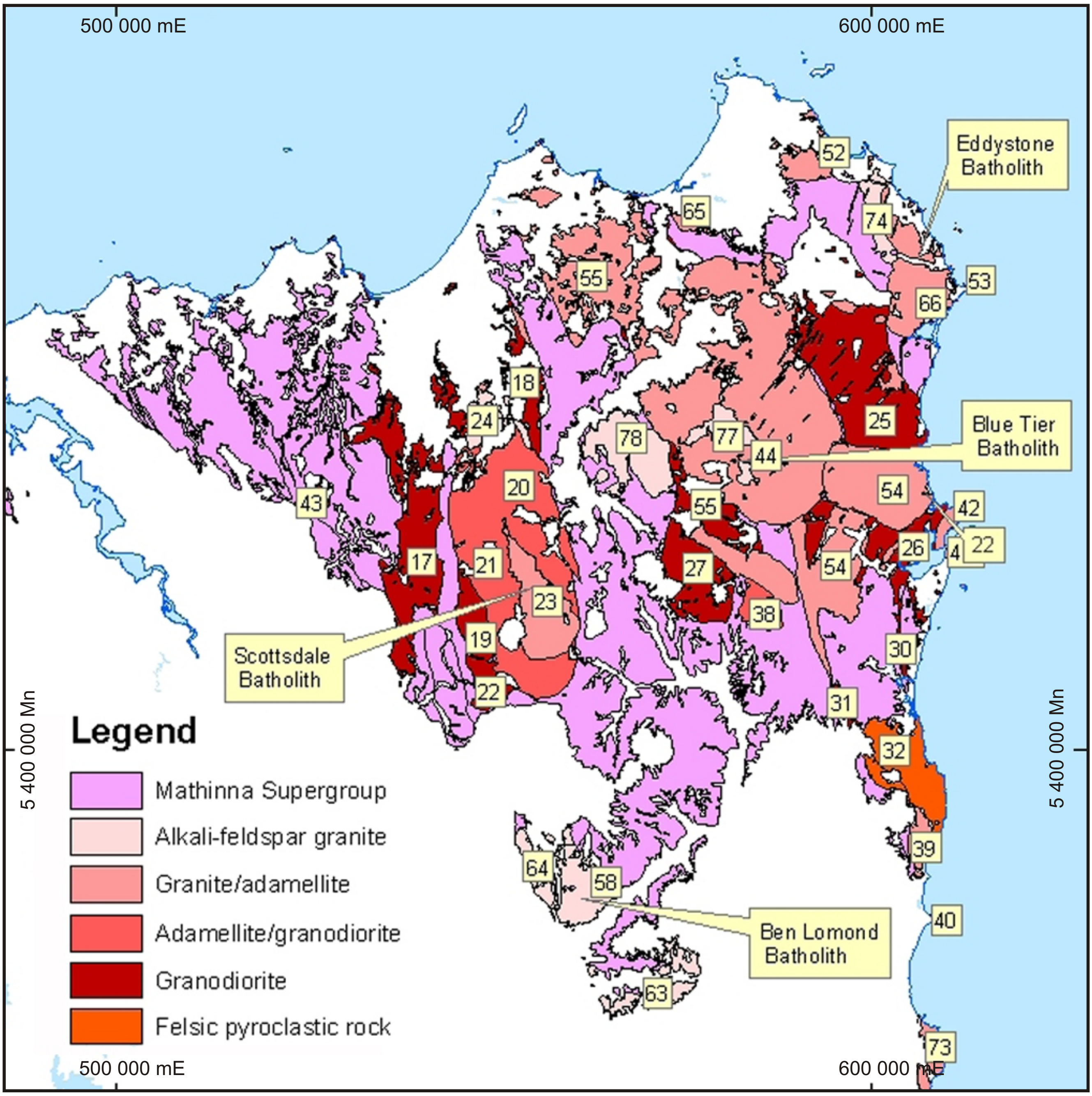


Figure 8

Northeast Tasmanian Devonian granites. White areas are post-Devonian cover sequences. Numbers on map refer to granite bodies in Table 1.

the body at Little Mt Horror is also equivocal (McClenaghan *et al.*, 1982).

The Mt William alkali-feldspar body occurs a short distance to the north of Ansons Bay in the Eddystone Batholith (fig. 8). It is an equigranular pink, medium-grained biotite-muscovite granite that contains quartz, microperthitic K-feldspar, albite, biotite and muscovite, together with accessory apatite. The muscovite occurs as discrete flakes and as smaller flakes replacing feldspar and biotite (Groves *et al.*, 1977).

Further south on the Freycinet Peninsula (fig. 7) the dominant rock type is medium to coarse-grained pink alkali-feldspar granite with K-feldspar megacrysts (Dunderdale, 1989; Everard, 2001). The granite consists of quartz, cryptoperthitic microcline, albite and iron-rich mica (transitional from siderophyllite to zinnwaldite). Accessory minerals are apatite, zircon, fluorite, tourmaline and rare cassiterite, together with secondary muscovite. On the basis of fluid inclusion studies, Dunderdale (1989) interpreted the granite as having been hydrothermally altered.

Classification and geochemistry of the granite

The northeastern Tasmanian granite plutons can be grouped into suites with distinctive chemical, isotopic and petrographic character (Table 1). The suites have been characterised as having been derived from partial melting of sedimentary (S-type) or igneous (I-type) source rocks using the criteria of the restite model (White and Chappell, 1977; Chappell *et al.*, 1987; Chappell and White, 1992; Chappell, 1999). Some of the suites consist of very felsic granite and have been distinguished as crystal-fractionated on the basis of their major and trace element variation. Chappell (1999) has suggested that crystal-fractionated granites can be distinguished from unfractionated granites by having less than 50 ppm Sr (also greater than 200 ppm Ba, and less than 250 ppm Rb). In fractionated granites, decrease in Sr values with increasing fractionation (using total iron as FeO (FeO*) as a fractionation index) is coupled with rises in Rb and falls in Ba. Other indications of crystal fractionation are marked rises in Nb, Ga and U.

In eastern Tasmania, granites of the Diddleum, Russells Road, Gardens and Scamander Tier suites

Table 1

Suite assignments for eastern Tasmanian mid-Palaeozoic granite bodies. Numbers in brackets after body name correspond with approximate locations shown in Figure 8 (not all locations are shown).

Suite	*Granite body or volcanic unit associated with granite body*
Eastern Tasmanian I-type granites	
Diddleum	Diddleum (17), Tulendeena (18), Porcupine Creek (19)
Russells Road	Russells Road (20), Hogarth Road (21), Upper Blessington (22), Tombstone Creek (23), Mt Stronach (24)
Gardens	Gardens (25), George River (26), Pyengana (27), Bluestone Bay (28), Chappell Islands (29)
Scamander Tier	Scamander Tier (30), Catos Creek (31), St Marys (32)
Long Point	Long Point (40)
Unassigned or other single body suites	Haleys New Country (38), Piccaninny Creek (39), Akaroa (41), Grants Point (42), Lisle (43)
Eastern Tasmanian S-type granites	
Poimena	Poimena (central parts) (44). Granite bodies in the Furneaux Group: Corner (45), Modder River (46), Craggy Island (47), Prime Seal Island (48), Cape Frankland (49), Martins Rise (50), Killiecrankie (51)
Musselroe	Musselroe (52), Eddystone Point (53), Mt Pearson (54), Poimena (marginal parts) (55)
Eastern Tasmanian crystal fractionated I-type granites	
Henbury	Henbury (58)
Freycinet	Freycinet (59), The Hazards (60), Coles Bay (61), Schouten Island (62)
Eastern Tasmanian crystal fractionated S-type granites	
Royal George	Royal George (63), Gipps Creek (64)
Boobyalla	Boobyalla (65), Ansons Bay (66), Bicheno (73), Mt William (74), Maria Island (75), Deep Glen Bay (76), Hippolyte Rocks (80). Granite bodies in the Furneaux Group: Key Bay (67), Strzelecki (68), Patriarchs (69), Mt Kerford (70), Hogans Hill (71), Babel Island (72)
Lottah	Lottah (77), Mt Paris (78)

contain hornblende, a diagnostic mineral for I-type granites (Chappell and White, 1992). Granites from the Diddleum, Gardens and Scamander Tier suites also contain very minor amounts of clinopyroxene, sphene and allanite which are also minerals characteristic of I-type granite. Magnetite, which is characteristic of I-type granite, occurs only in parts of the Pyengana Pluton (part of the Gardens suite) and in the Lisle body.

These I-type suites have a wide range of chemical composition and generally show well defined linear trends on two element plots (see fig. 9). The lowest Sr values in the suites are generally well in excess of 50 ppm, which suggests they have not undergone significant crystal fractionation. Plagioclase in these suites contains calcic core regions of comparatively uniform composition, which may represent restite material (Chappell *et al.*, 1987). This feature, together with the linear nature of the chemical variation, is consistent with the suites having undergone restite fractionation, although it does not rule out other possibilities.

The pink granites of the Mt Stronach and Tombstone Creek plutons, which contrast with the grey granites for most of the Scottsdale Batholith, have higher Rb values and depart from the linear trend of the other granites from the Russells Road suite. This suggests that they had shed their restite component and undergone some feldspar fractionation.

The Poimena and Musselroe suites also show linear trends on two element plots (fig. 10). That variation may also have been produced by restite fractionation as they contain plagioclase core regions. Granites from these suites are peraluminous, generally with higher aluminium saturation indices than the I-type Diddleum, Russells Road, Gardens and Scamander Tier suites. Some of the granite bodies forming these suites contain cordierite and garnet, Al-rich minerals diagnostic of S-type granite, possibly representing restite. Bodies such as the Poimena and Mt Pearson plutons do not contain these minerals but have similar aluminium saturation indices to those that do. The Poimena and Musselroe suites have been designated as S-type but their CaO content is higher than usual for S-type granite, possibly indicating that they were derived from a sedimentary source rock containing a high proportion of unweathered igneous rock detritus. The Poimena and Musselroe suites include a small proportion of granite with Sr values less than 50 ppm and Rb greater than 250 ppm, suggesting that feldspar crystal fractionation has produced the more felsic part of the trend.

High Rb and low Sr values indicate that the Henbury, Royal George, Boobyalla, Lottah and Freycinet suites are strongly crystal-fractionated. Of these the Lottah suite is the most strongly fractionated, and the Boobyalla suite the least fractionated (fig. 11). The Henbury and Freycinet suites range from metaluminous to peraluminous, whereas the Boobyalla suite is peraluminous and the Lottah and Royal George suites are strongly peraluminous (fig. 11). This is consistent with the characterisation of the Henbury and Freycinet suites as I-type and the Boobyalla, Lottah and Royal George suites as S-type. The Freycinet suite has previously been described as a strongly fractionated I-type granite by Chappell (1999). Several of the plutons making up the Boobyalla suite contain the Al-rich minerals cordierite and garnet, which are diagnostic of S-type granites and may represent restite. The absence of potential restite minerals in the Henbury, Royal George, Lottah and Freycinet suites is consistent with their greater degree of crystal fractionation which would be expected to clear the magma of restite.

Age of the granites

Black (in prep.) has pointed out that the U-Pb zircon isotopic clock has been more resistant to resetting than those for Rb-Sr and K-Ar. Dates for northeast Tasmanian granites using the Rb-Sr and K-Ar systems (McDougall and Leggo, 1965; Cocker, 1982; Turner *et al.*, 1986; MacKenzie *et al.*, 1988) are generally about 10 Ma to 15 Ma younger than U-Pb zircon dates for the same body. The discussion that follows refers to U-Pb zircon dates (Black, in prep.).

In eastern Tasmania the oldest igneous activity is I-type granite represented by the foliated Diddleum (391.5   2.5 Ma), Gardens (400.9   2.0 Ma) and Long Point (395.6   2.6 Ma) bodies. The start of intrusion was about 10 Ma earlier in the Blue Tier Batholith than in the Scottsdale Batholith. In the Blue Tier Batholith igneous activity extended over about 23 Ma and ended with the intrusion of the highly crystal-fractionated S-type Lottah (377.0   2.9 Ma) body. The Lottah body is distinctly younger than the Poimena (383.8   2.9 Ma) body, which it intrudes.

Ages of restite zircon inherited from the granite source rocks have been measured for granite bodies from eastern Tasmania (Black, in prep.). No significant difference was detected between zircon age components for I and S-type granites, although I types contained less inheritance.

Isotopic variation of the granites

In eastern Tasmania the initial $^{87}Sr/^{86}Sr$ ratios of the foliated I-type granites of the Scottsdale and Blue Tier batholiths are very similar and fall within a small range (0.706–0.707) (McDougall and Leggo, 1965; Turner *et al.*, 1986). The initial $^{87}Sr/^{86}Sr$ ratios for the other I-type granite bodies, including the Freycinet body (Cocker, 1982), which is part of a strongly crystal-fractionated suite, have similar values and range. The similarity of these values suggests that these I-type granites may have originated from a moderately homogeneous source rock.

Initial $^{87}Sr/^{86}Sr$ ratios for the Mt Pearson (0.7082   0.0019) and Poimena (0.7105   0.0036) bodies (Cocker, 1982), which are part of the Musselroe and Poimena S-type granite suites, are only slightly higher. This small difference in values may reflect the marginal S-type chemical and mineralogical character of these

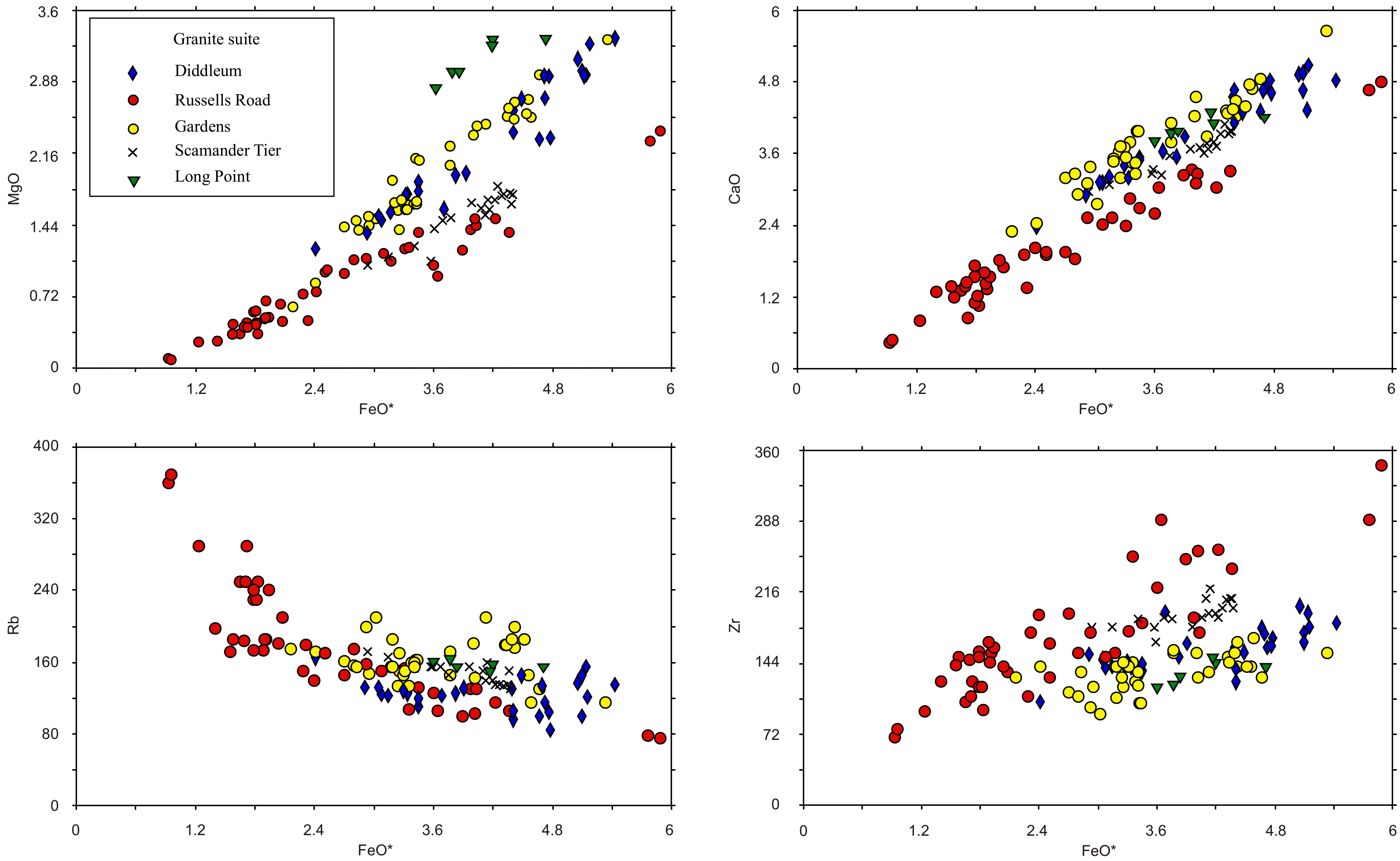


Figure 9

Variation diagrams for less strongly crystal-fractionated eastern Tasmanian mid-Palaeozoic I-type granites. FeO is total iron as FeO. Sources of data in descending order of contribution: Chappell (pers. comm., 2001); McClenaghan (unpublished data); McClenaghan* et al.*, (1992); Black (pers. comm., 2001); Everard (pers. comm., 2001); Baillie (1986).*

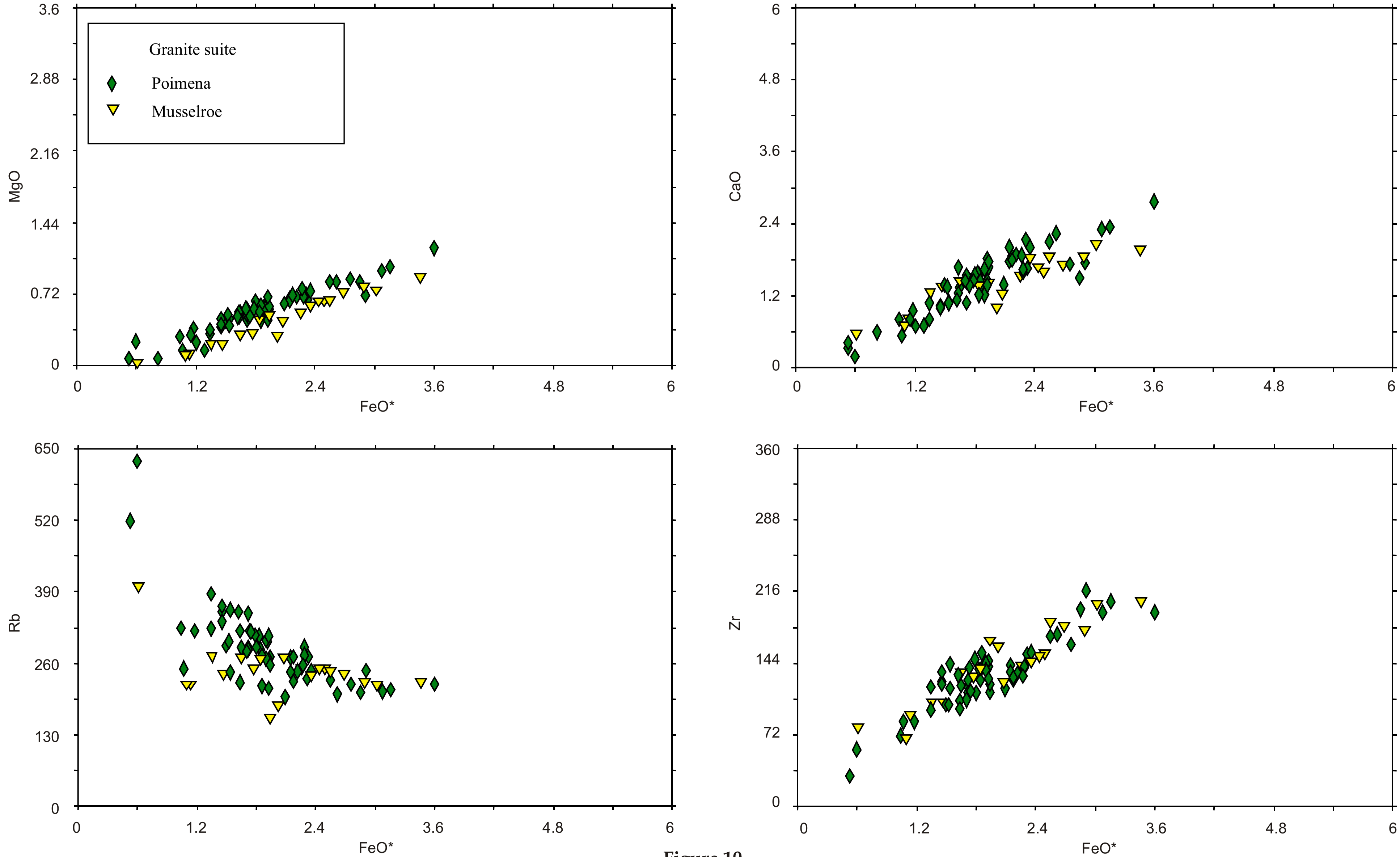


Figure 10

Variation diagrams for less strongly crystal-fractionated eastern Tasmanian mid-Palaeozoic S-type granites. FeO is total iron as FeO. Sources of data in descending order of contribution: Chappell (pers. comm., 2001); McClenaghan (unpublished data); MacKenzie* et al. *(1988); McClenaghan (1984); Turner* et al. *(1986); Vicary (1987); Black (pers. comm., 2001); Baillie (1986); Everard (pers. comm., 2001); McClenaghan* et al. *(1982).*

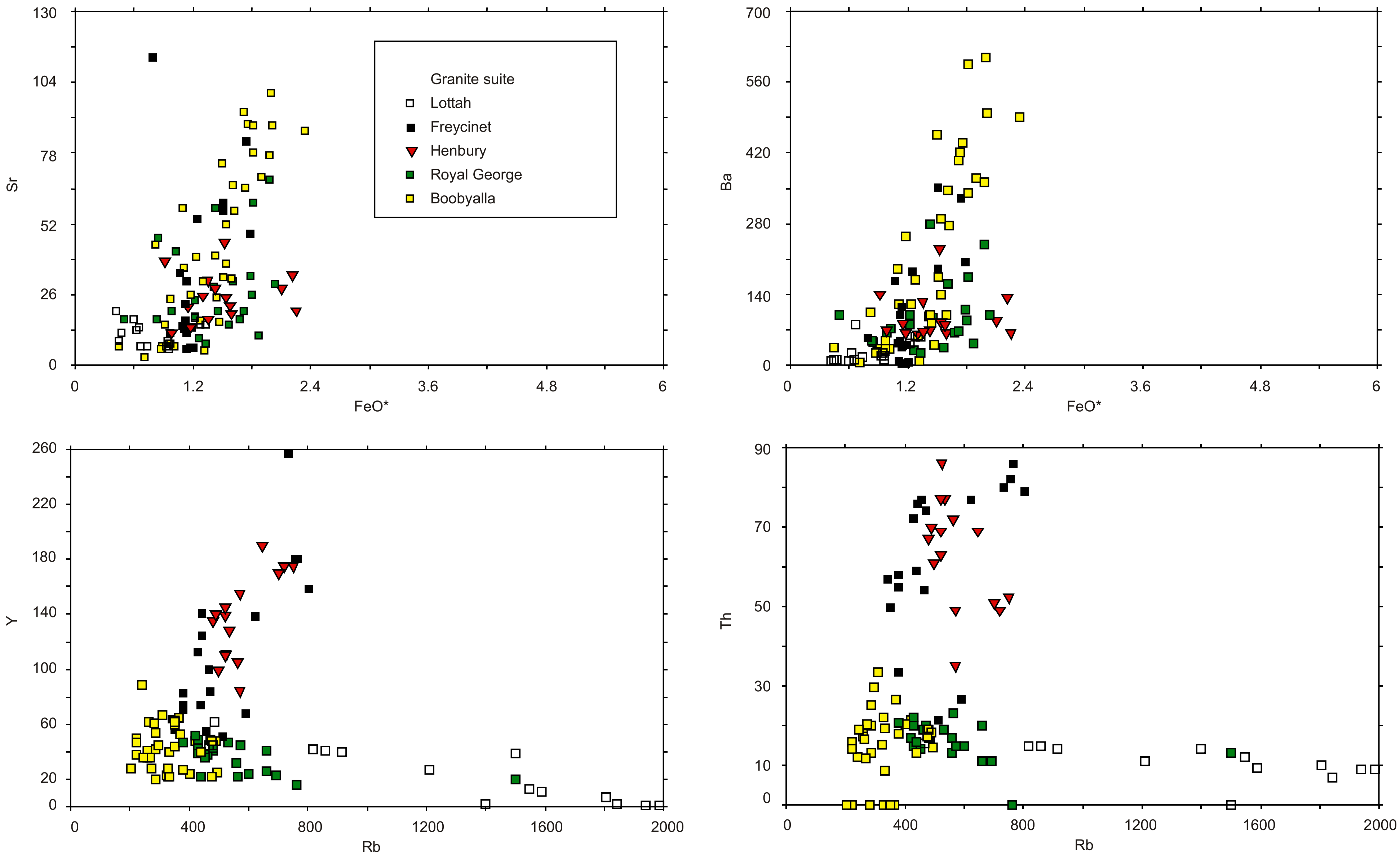


Figure 11

Variation diagrams for strongly crystal-fractionated eastern Tasmanian mid-Palaeozoic granites. FeO is total iron as FeO, ASI is the aluminium saturation index (mol. $Al_2O_3/(CaO+Na_2O+K_2O)$) and CORUNDUM is normative corundum.*

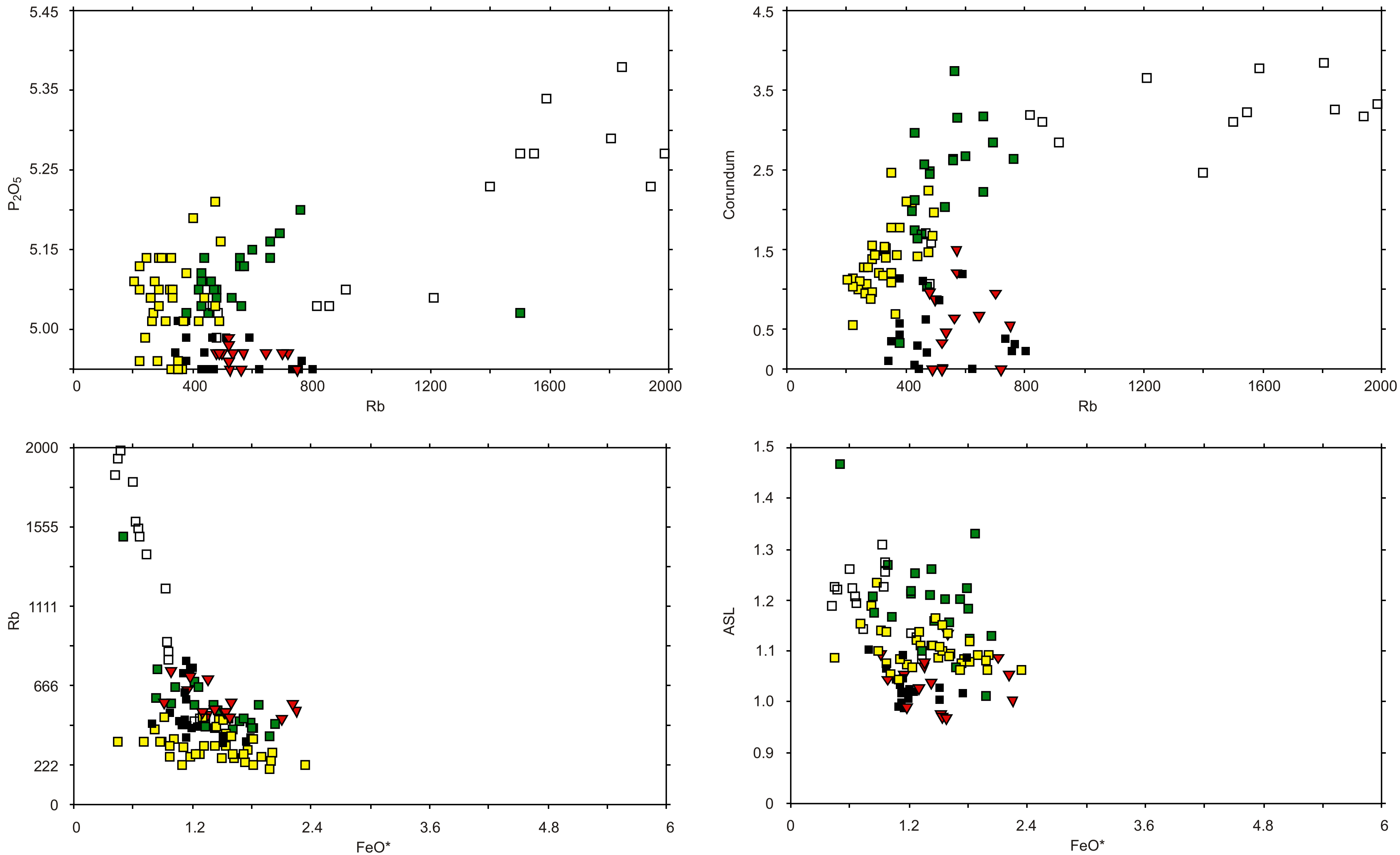


Figure 11 *(continued)*

Variation diagrams for strongly crystal-fractionated eastern Tasmanian mid-Palaeozoic granites. FeO is total iron as FeO, ASI is the aluminium saturation index (mol. $Al_2O_3/(CaO+Na_2O+K_2O)$) and CORUNDUM is normative corundum.*

suites. The values for the Ansons Bay north and south (0.7122 0.0033, 0.7136 0.0005) and Boobyalla (0.7119 0.0009) bodies (Cocker, 1982), which are part of the Boobyalla S-type granite suite, are distinctly higher than those of the I-type granite suites and than those of the Musselroe and Poimena S-type granite suites.

Granites from the I and S-type granite suites, excluding the Lottah suite, of the Blue Tier and Eddystone batholiths have similar $^{143}Nd/^{144}Nd$ initial ratios (Nd = -5.0 to -6.5) at 370 Ma (Sun and Higgins, 1996). The source-rock model age for these granites, based on a depleted mantle evolution model (TDM), is about 1.6 Ga (Sun and Higgins, 1996). The Sm-Nd model age represents the time when the major REE fractionation in the components of a rock occurred, that is during their derivation from a mantle reservoir (McCulloch and Chappell, 1982, p.60). The Sm-Nd model age approximates the average age of the components of the source rock, not the age of the source rock itself and so may be consistent with the presence of a small proportion of inherited zircons of about ~500–600 Ma in eastern Tasmanian granites (Black, in prep.).

Metallogenic potential of the granites

The type of fractionation mechanism that produced the compositional variation in a granite suite is an important factor in concentrating trace elements. Restite fractionation does not cause marked enrichment or depletion of trace elements relative to the original source concentration (Blevin and Chappell, 1992). The ore elements may be contained within restite minerals and not in the melt. They are therefore not available for partitioning into an exsolving aqueous phase. In addition, restite-rich magma would be cooler than the melt-rich equivalent magma and so there would be less crystal fractionation if the magma rid itself of the entrained restite, and less heat would be provided to the wall rocks or would be available to drive hydrothermal systems. Extended crystal fractionation can produce considerable enrichment in incompatible trace elements, and significant feldspar fractionation is indicated by Rb concentrations above 250 ppm (Blevin and Chappell, 1992). The identification of the Diddleum, Russells Road, Scamander Tier, Poimena and Musselroe suites as likely to have been substantially restite-fractionated rather than crystal-fractionated downgrades their mineralisation potential. The higher Rb values for the Mt Stronach and Tombstone Creek bodies indicates that they would have the most mineralisation potential of the Russells Road suite. In eastern Tasmania the Henbury, Royal George, Lottah and Freycinet suites are strongly crystal-fractionated, and have high mineralisation potential. The Boobyalla suite is less strongly crystal-fractionated and has a lesser potential.

The type of ore-elements associated with a granite suite can be related to its oxidation state (Blevin and Chappell, 1992). Sn can be in the Sn^{4+} or Sn^{2+} state. In the octahedrally co-ordinated Sn^{4+} state it has an ionic radius similar to that of Ti^{4+} and Fe^{3+}, enabling it to substitute for these ions in biotite, hornblende, sphene, ilmenite and magnetite. If the melt is reduced, Sn will not be in the Sn^{4+} state, and will not be removed from the melt by the crystallisation of these minerals. Concentration of Sn in the melt by fractional crystallisation would be favoured for less oxidised granite suites. In extreme cases of crystal fractionation, where the fractionation process had become dominated by feldspar, Sn would be concentrated even in more oxidised granite suites.

S-type granites are generally less oxidised than I-type granites, which has been ascribed to the presence of graphite within their S-type source-rocks (Flood and Shaw, 1975). The S-type granites in eastern Tasmania are all ilmenite bearing rather than magnetite bearing and generally have lower Fe_2O_3/FeO ratios than the magnetite-bearing granites. The I-type granites in eastern Tasmania differ from most I-type granites elsewhere in the Lachlan Fold Belt by being ilmenite bearing, apart from the Pyengana and Lisle plutons which include some magnetite-bearing granite. Sn mineralisation is most strongly associated with the crystal-fractionated S and I-type granites of eastern Tasmania, as would be predicted from the above discussion. In relatively reduced granitic melts Mo partitions more into ilmenite and so is depleted in the residual melt by crystal fractionation. In more oxidised conditions Mo is partitioned more into the melt (Blevin and Chappell, 1992). The Russells Road suite shows low-grade Mo mineralisation in the Mt Stronach pluton (Langsford and Westhoff, 1982), which is the most fractionated body in the suite.

W mineralisation shows little dependence on the oxidation state of the granite magma (Blevin and Chappell, 1992), but is enhanced by crystal fractionation. The strongly crystal-fractionated granite suites, both I and S type, would be expected to have the highest potential for W mineralisation. Economic deposits of W have been associated with the Ben Lomond granite in northeastern Tasmania. (Brown, *in* Burrett and Martin, 1989).

Petrogenesis of the granites

A cumulate fractional crystallisation model has been proposed for the Blue Tier Batholith in eastern Tasmania (McCarthy and Groves, 1979). This model envisaged that the batholith formed by fractional crystallisation of a single magma, of adamellite composition, which underwent crystallisation *in situ* by progressive nucleation and solidification from the margins inwards. Progressive changes in liquids and the cumulate mineralogy during crystallisation led to the observed sequence of early granodiorites followed by biotite adamellites and the alkali-feldspar granites. This model has been criticised by Cocker (1982) on the grounds that the apparently unique mineral, chemical and isotopic composition of each pluton points to a number of separate magmas rather than an origin from a single magma. The division into suites suggested here, based on chemical and petrographic character, although pointing to a connection between some plutons, also supports the conclusion that the granites have been derived from a number of distinct magmas.

Crystal fractionation models have been proposed to explain the chemical variation in the Pyengana and Gardens granodiorite plutons. McClenaghan (1984) suggested that fractionation in both plutons involved hornblende, plagioclase, biotite and minor amounts of apatite and sphene. Higgins *et al*. (1985) modelled the variation in the Pyengana granodiorite pluton using clinopyroxene, orthopyroxene, biotite and plagioclase as the fractionating phases. Work on the St Marys volcanic body and associated intrusive rocks (Higgins *et al*., 1986) supports a crystal fractionation model for the variation in bodies of granodiorite composition.

McClenaghan and Williams (1982) suggested that chemical variation in the Poimena Pluton adamellite could be explained by restite-unmixing or fractional crystallisation. Higgins *et al*. (1985) favoured a fractional crystallisation model involving pyroxene, biotite, plagioclase and K-feldspar. Mackenzie *et al*. (1988) considered that the Poimena Pluton adamellite was generated by partial melting of a source of basaltic andesite composition, which underwent limited restite unmixing.

McClenaghan and Williams (1982) considered that the Lottah Pluton alkali-feldspar granite could have been derived from Poimena Pluton adamellite by crystal fractionation, or by a combination of restite unmixing and crystal fractionation. Higgins *et al*. (1985) suggested that the derivation process was one of fractional crystallisation combined with variable metasomatism following aqueous fluid saturation of the magma. They advocated this process also for the derivation of the Mt William alkali-feldspar granite from the Ansons Bay biotite-cordierite-garnet adamellite. Mackenzie *et al*. (1988) concluded that the Lottah alkali-feldspar granite formed from a different magma to the Poimena adamellite. They based this conclusion on the lack of continuity of trends on variation diagrams, the ~10 million year difference in emplacement age (more recently measured as ~7 Ma by Black (in prep.), the difference in initial $^{87}Sr/^{86}Sr$ ratios, and the difference in Nd values between the two granite types. Mackenzie *et al*. (1988) considered that the compositional variation in the Lottah alkali-feldspar granite is consistent with fractional crystallisation of a felsic peraluminous melt rich in F, Li and B, to produce even more peraluminous residual melts, progressively further enriched in these components. They did not consider that there was a significant metasomatic component to the chemical variation.

On the basis of P, Th and Y variation, Chappell (1999) postulated that the granites of the Freycinet suite were produced by strong crystal fractionation of I-type granitic magma.

The similarity of inherited zircon age components for I and S-type granites (Black, in prep.) suggests a common component or components to their source rocks. The variation in chemistry between the I and S-type granites in Tasmania may reflect varying proportions of igneous and sedimentary material in a mixed source rock sequence.

Granites of Western Tasmania

Grassy, Bold Head and Sea Elephant adamellites

The Grassy and Sea Elephant adamellites are small bodies in the eastern part of King Island. The Grassy body has sharp discordant contacts with the Cambrian (?) country rock. Another small intrusion at Bold Head, near Grassy, may be a faulted-off sliver of the Grassy intrusion. Work on the granites has concentrated on the Grassy bodies, which are associated with tungsten mineralisation (Haynes, 1973; Tan, 1979; Calcraft, 1980; Wesolowski *et al*., 1988).

The Grassy Granite is porphyritic with megacrysts of pink K-feldspar. The groundmass consists of quartz, K-feldspar, plagioclase ($<An_{32}$), biotite and amphibole, with accessory apatite, allanite, sphene, magnetite and zircon. The Sea Elephant body is more felsic and is texturally similar to the other bodies but differs in that it contains lesser amounts of amphibole and sphene. Although dominantly of adamellite composition granodiorite is also present.

Three Hummock Island Granite

Three Hummock Island, lying 24 kilometres off the northwest coast of Tasmania, is composed entirely of granite, apart from an extensive thin cover of Tertiary and Quaternary deposits (McDougall and Leggo, 1965; Jennings, 1976; Everard *et al*., 1997). Aeromagnetic data indicate that the granite is part of a much larger oval-shaped pluton lying mostly to the north of the island. Gravity data indicate that the granite also extends southwest to underlie much of Hunter Island at shallow depth (<1 km) (fig. 7). Biotite-muscovite granite exposed on Penguin Islet, a short distance east of Hunter Island, may represent a small apophysis of the Three Hummock Island body (Everard *et al*., 1997).

The granite is medium to coarse-grained, consisting of K-feldspar, plagioclase, quartz, biotite and muscovite with accessory tourmaline and apatite. K-feldspar megacrysts are variably abundant, and cordierite may be recognised in outcrop as scattered euhedral phenocrysts up to 20 10 mm concentrated within zones of K-feldspar accumulation (Wyborn and Chappell, 1998). In the south of the island, the granite is slightly more mafic and locally contains biotite-rich schlieren. At one place a swarm of melanocratic enclaves comprise 30–40% of the overall outcrop (Everard *et al*., 1997). Composition ranges from adamellite to alkali-feldspar granite.

Housetop Batholith

The Housetop Batholith is a large 157 km^2 red granite body 20 km south of Burnie. Previous field investigations on the body include work by

McDougall and Leggo (1965), Calcraft (1980), Camacho (1987) and Baillie *et al*. (1986). The batholith has sharp discordant contacts against folded Precambrian–Devonian rocks (McDougall and Leggo, 1965). Despite compositional and textural variations, Baillie and Lennox (*in* Seymour, 1989) considered that the granite was not composite, whereas Camacho (1987) distinguished individual granite plutons within the body. It is undeformed, and schlieren are the only form of layering observed.

The following description is drawn from Baillie and Lennox (*in* Seymour, 1989) and Camacho (1987). The rocks of the batholith are generally reddish pink, and have a range of textures with fine to medium-grained groundmass, moderately to strong porphyritic texture, miarolitic cavities and thin (100–150 mm) vuggy pegmatites. All textural types are cut by porphyritic, aplitic and microgranite dykes. The strongly porphyritic rocks contain many large crystals of K-feldspar and quartz, and to a lesser extent plagioclase in the range 10–20 mm, and have a fine-grained granophyric groundmass dominantly of quartz, K-feldspar and plagioclase. The K-feldspar is perthitic and locally shows rapakavi textures. The plagioclase is zoned. Distinct calcic-rich cores are not present, with the most calcium-rich composition being An_{26}. Biotite is always present, with the addition of hornblende in the less felsic granites, mostly in the western part of the batholith. Accessory minerals are allanite, zircon, magnetite, ilmenite, sphene, apatite and fluorite. Tourmaline is present in the southern part of the batholith. Mafic hornblende-rich and biotite-rich xenoliths are confined to the least felsic rocks in the western and northern margins of the batholith. Composition ranges from adamellite to alkali-feldspar granite.

Dolcoath Granite

The Dolcoath Granite crops out as a small (<10 km^2) roughly circular body in the Forth Valley near Lake Cethana, where it intrudes folded Cambrian and Ordovician rocks (Jennings, 1963; Gee, 1965; Webb, 1974). Granite isobaths (fig. 7) indicate that the exposure is the tip of a much larger subsurface granite body that extends to the west.

The granite is medium to coarse-grained and consists of quartz, perthitic microcline, plagioclase and biotite. Accessory minerals include zircon, apatite, fluorite, topaz, cassiterite and disseminated sulphides (molybdenite and pyrite). Contact metamorphic assemblages are characteristic of the amphibolite hornfels facies (Webb, 1974).

Meredith Batholith

The Meredith Batholith is a large composite granite body that underlies an area of 285 km^2 southwest of Waratah. Contacts with the folded Precambrian–Devonian country rocks are irregular and discordant. There are two main intrusive units in the batholith, a smaller finer-grained and less felsic body in the northeast (Wombat Flat Adamellite) and a much larger, coarser-grained and more felsic body (Meredith Granite) forming the remainder of the batholith (Reid, 1923; Groves, 1968; Stockley, 1972; Groves *et al*., 1972; Collins, 1984; Camacho, 1987; Turner *et al*., 1991).

The Wombat Flat body consists of an equigranular fine to medium-grained core comprising quartz, K-feldspar, plagioclase, biotite and hornblende with accessory allanite, zircon, apatite, sphene, ilmenite and magnetite, and a porphyritic rim with phenocrysts of quartz, K-feldspar, plagioclase (An_{20-30}) and hornblende, with the same groundmass mineralogy (Camacho, 1987). The K-feldspar is perthitic; sparse miarolitic cavities occur in the fine-grained variety. Ilmenite is the dominant opaque mineral; fluorite and monazite are rare accessories.

The Meredith body is composed of adamellite to alkali-feldspar granite. It consists of a very coarse grained, pale grey to very pale pink, equigranular phase and abundant fine to coarse-grained, porphyritic intrusions containing phenocrysts of quartz, K-feldspar and plagioclase (Turner *et al*., 1991). Aligned K-feldspar megacrysts are common in some places. Groundmass mineralogy consists of quartz, K-feldspar, plagioclase and biotite, with accessory apatite, zircon, monazite, ilmenite and tourmaline. Alteration minerals are fluorite, topaz and sericite. Quartz/tourmaline nodules are abundant, and veins of aplite, quartz, quartz/tourmaline and quartz/muscovite are common. Miarolitic cavities are locally present. This body does not contain allanite or amphibole, has more iron-rich biotite, and contains more monazite than the Wombat Flat body.

Mt Bischoff porphyry dykes

At Mt Bischoff, northeast of the Meredith Batholith, radiating quartz porphyry dykes intrude Precambrian and Cambrian rocks. It has been suggested that these dykes emanate from the cupola of an underlying granite body (Groves and Solomon, 1964). Greisenisation of the porphyries is generally extreme with the formation of topaz, tourmaline, muscovite, cassiterite and pyrite pseudomorphing primary feldspar.

Birthday Granite

Two small bodies of muscovite-biotite granite, presumed to be of Devonian age (Macleod *et al*., 1961; Jennings, 1963), intrude Precambrian quartzite and quartz mica schist in the upper part of the Forth Valley.

The granite consists of biotite and muscovite, pinkish white feldspar and coarse quartz. The biotite and muscovite are locally predominant. Near the contact the granite commonly contains large phenocrysts of feldspar and abundant biotite. Some associated quartz veins contain wolframite, pyrite, cassiterite and rare molybdenite.

The granite bodies may connect to the same larger subsurface granite (fig. 7) that extends west to the Granite Tor Granite and the Heemskirk Batholith.

Granite Tor Granite

Little information is available on the Granite Tor Granite, which lies to the northeast of Rosebery, intrudes Precambrian rocks, and crops out over an area of 55 km^2 (McDougall and Leggo, 1965). The granite isobaths (fig. 7) suggest it is part of a much larger subsurface body that extends further to the east and also exists as a subsurface ridge westward to the Heemskirk Batholith.

The body is medium to coarse-grained biotite-muscovite granite with megacrysts of K-feldspar. Accessory minerals include zircon, apatite and garnet (Speijers, 1979). Tourmaline is present, as an accessory or essential phase, at some localities. The uniformity of texture and composition suggests that the granite was intruded in a single event. Small areas of country rocks overlying the granite in several areas suggest that the current outcrops are close to the roof of the intrusion (McClenaghan, 2003).

Renison Complex

Granite of this complex has been intersected below mineralised rocks in the Renison Bell mine. Patterson *et al.* (1981) related it to the strongly greisenised Pine Hill stock, which crops out two kilometres south of the mine. The granite intrudes both the Success Creek Group and Crimson Creek Formation of Neoproterozoic age. Granite isobaths (fig. 7) indicate that the body is the tip of a subsurface granite ridge connecting the Heemskirk Batholith with the Granite Tor Granite.

The granite, described by Patterson (1980), Ward (1981) and Camacho (1987), consists of medium to coarse-grained equigranular and porphyritic (K-feldspar, quartz) varieties with a groundmass of quartz, K-feldspar, plagioclase and biotite. Accessory minerals include apatite, zircon, magnetite, ilmenite, allanite, sphene, monazite, tourmaline, fluorite and topaz. Fluid-rock interaction produced zones of tourmalinisation, sericitisation and albitisation (Bajwah *et al.*, 1995). The granite ranges in composition from adamellite to alkali-feldspar granite.

Heemskirk Batholith

The Heemskirk Batholith is a large elongate body occupying 117 km^2 on the west coast of Tasmania near Zeehan. It has steep, sharp intrusive contacts with Precambrian metasedimentary rocks. The contact aureole on the southern side includes Neoproterozoic, Cambrian and Silurian to Devonian sedimentary rocks (Brooks and Compston, 1965).

The intrusion is composite, consisting of red and white types (Heier and Brooks, 1966; Klominsky, 1972; Hajitaheri, 1985; McClenaghan, 1994*b*). The red granite occupies the upper parts of the intrusion and has been intruded by the white granite, which forms the western and major part of the body. A tourmaline nodular facies in the white granite occurs at the contact between the red and white granite, suggesting the trapping of a fluid-rich phase (Klominsky, 1972). The body has grown by intrusion of granite sheets into space created by subsidence within a semi-circular cauldron-type structure (Hajitaheri, 1985). Isotopic dating shows that the two intrusions are almost contemporaneous.

The southern part of the batholith is dominantly very coarse to coarse grained, with the remaining parts of the red granite having a complex distribution of fine to coarse-grained and very coarse-grained granite. Finer granite varieties generally have small phenocrysts of feldspar and quartz. The white type consists of fine and coarse-grained varieties with the fine-grained granite generally near the contact with the red type or the country rock. Prominent quartz-tourmaline nodules and patches occur in both the red and white type but are more abundant in the white, whereas quartz-tourmaline veins are more abundant in the red granite.

The major mineralogy of both rock types is quartz, K-feldspar and plagioclase with varying amounts of biotite and tourmaline. The K-feldspar in the red type is pink; in contrast the K-feldspar of the white type is white. Hornblende ranges up to 3% in the red granite. Accessory minerals are apatite, zircon and fluorite in both types with magnetite, sphene and allanite confined to the red granite, and monazite, cassiterite and muscovite only occurring in the white granite. Biotite from the red granite is higher in Ti and lower in Al than biotite from the white granite (Hajitaheri, 1985). The white granite ranges in composition from adamellite to alkali-feldspar granite and the red granite includes adamellite and granite.

The granite contact has produced a variety of albite-epidote, hornblende and pyroxene hornfels facies mineral assemblages in the mafic and ultramafic country rocks (Green, 1966).

Pieman Granite

The Pieman Granite is a small body (8 km^2) intruded into Precambrian rocks on the southern side of the mouth of the Pieman River 15 km north of the Heemskirk Batholith (Spry and Ford, 1957; Brooks, 1966).

The granite is coarse grained, consisting of K-feldspar, plagioclase, quartz, biotite and muscovite with accessory tourmaline, zircon and apatite. Texture varies from equigranular to distinctly porphyritic, with phenocrysts of altered perthitic K-feldspar. Tourmaline nodules are common. The Pieman Granite, while petrologically similar to the white granite variety of the Heemskirk Batholith, ranges in composition from adamellite to alkali-feldspar granite.

Interview and Sandy Cape granites

The Interview Granite is an elongate body occupying an area of 87 km^2 from the Pieman River to near Sandy Cape, where there is a small granite body of similar character. It intruded a faulted anticline in Precambrian rocks (Spry and Ford, 1957).

The granite is generally equigranular and medium to coarse-grained, consisting of K-feldspar, plagioclase, quartz, biotite and muscovite with accessory tourmaline, zircon, apatite and ilmenite. Small amounts (<1%) of subhedral cordierite less than 1 mm across, mostly altered to chlorite, together with rare resorbed almandine-rich garnet are also present (Wyborn and Chappell, 1998). Aligned K-feldspar megacrysts are locally common. A number of intrusive phases are present at Sandy Cape, with a coarse-grained, equigranular, muscovite leucogranite being dominant (Wyborn and Chappell, 1998). The Interview and Sandy Cape granites range in composition from adamellite to alkali-feldspar granite.

Grandfathers Granite

The Grandfathers Granite crops out over a small area on the western coast of the northern part of the Cape Sorell peninsula (Baillie and Corbett, 1985). The granite isobaths (fig. 7) suggest that this is part of a larger subsurface body that occurs adjacent to the coast.

The granite has irregular contacts with Precambrian country rocks. Xenoliths of quartzite are common near the margins of the body. The granite is an unfoliated, leucocratic, equigranular medium to coarse-grained rock consisting of quartz, K-feldspar, plagioclase, biotite, muscovite and tourmaline. Accessory minerals are zircon and apatite. The biotite shows red/brown pleochroic colours where fresh. Secondary muscovite is present as irregular-shaped crystals in cracks, and as small flakes enclosed by plagioclase. Miarolytic cavities and tourmaline segregations are common.

Cox Bight Granite

A small area of contact aureole granite (Hall, 1965) occurs at Cox Bight. The granite isobaths (fig. 7) suggest that this body is part of a much larger volume of granite that underlies a large part of southwest Tasmania. A small, entirely granite island (Mewstone), about 22 km SSE of Cox Bight, may be part the same granite mass.

The Cox Bight Granite is a medium to fine-grained, leucocratic, biotite granite with rare muscovite. It consists of quartz, K-feldspar, biotite, and rare albite-oligoclase and muscovite. It has a clearly defined intrusive margin with country rock of thick-bedded quartzite and quartz schist. A boulder of greisen was found near the granite margin.

The granite of the Mewstone is a medium light grey fine-grained muscovite granite, with sparse quartz and feldspar phenocrysts and miarolytic cavities. Tourmaline occurs in some samples (Banks, 1993).

Tin-tungsten deposits of Tasmania*

Primary tin and tungsten deposits in Tasmania occur in rocks which range in age from Late Proterozoic to Late Devonian. They are genetically and spatially related to the intrusion of Middle to Late Devonian granite bodies.

Tin-tungsten deposits mainly occur in two metallogenic provinces, northeastern and western Tasmania. The latter is characterised by more complex host rocks which has resulted in the formation of a much wider variety of deposit types.

Tin-tungsten of the northeastern province has yielded about 90 000 tonnes of Sn, of which some 65 tonnes was from alluvial deposits. Tungsten production amounts to about 14 000 metric tonnes WO_3. Primary deposits include exogranitoid vein systems (Aberfoyle, Storys Creek, Scamander; fig. 13) and endogranitoid greisens in lenses and pipes (Rex Hill, Anchor; fig. 13, 14). There are some economic kaolin deposits which may represent hydrothermally-altered granites.

There are also many alluvial tin workings in Tertiary to Recent gravel along the Ringarooma River, which drains a large part of the central Blue Tier Batholith. Alluvial tin was also recovered from the Scamander and St Helens area on the east coast. The cassiterite was largely derived from the alkali feldspar granite plutons lying within and alongside the biotite granite bodies of the Blue Tier Batholith (fig. 14).

The western province is, in contrast, characterised by tin-tungsten skarn and replacement deposits reflecting the abundance of carbonate rocks in Precambrian to Early to mid-Palaeozoic sedimentary sequences (e.g. Renison, Mt Bischoff, Cleveland, Zeehan, King Island, Kara). There are also other types of tin-tungsten deposits which are schematically summarised in Figure 12.

Renison, with a total resource of 400 000 tonnes of tin, is one of the world's largest tin deposits. Mt Bischoff, discovered in 1871, was one the world's richest tin mines, producing some 5.6 Mt of ore at 1.6% tin, and is targeted for re-opening. King Island, with a pre-mining resource of about 17 Mt @ 0.85% WO_3, is amongst the major tungsten deposits in the world (Turner *et al.*, 2003).

* Adapted from Solomon and Groves (1994) and Collins (1986)

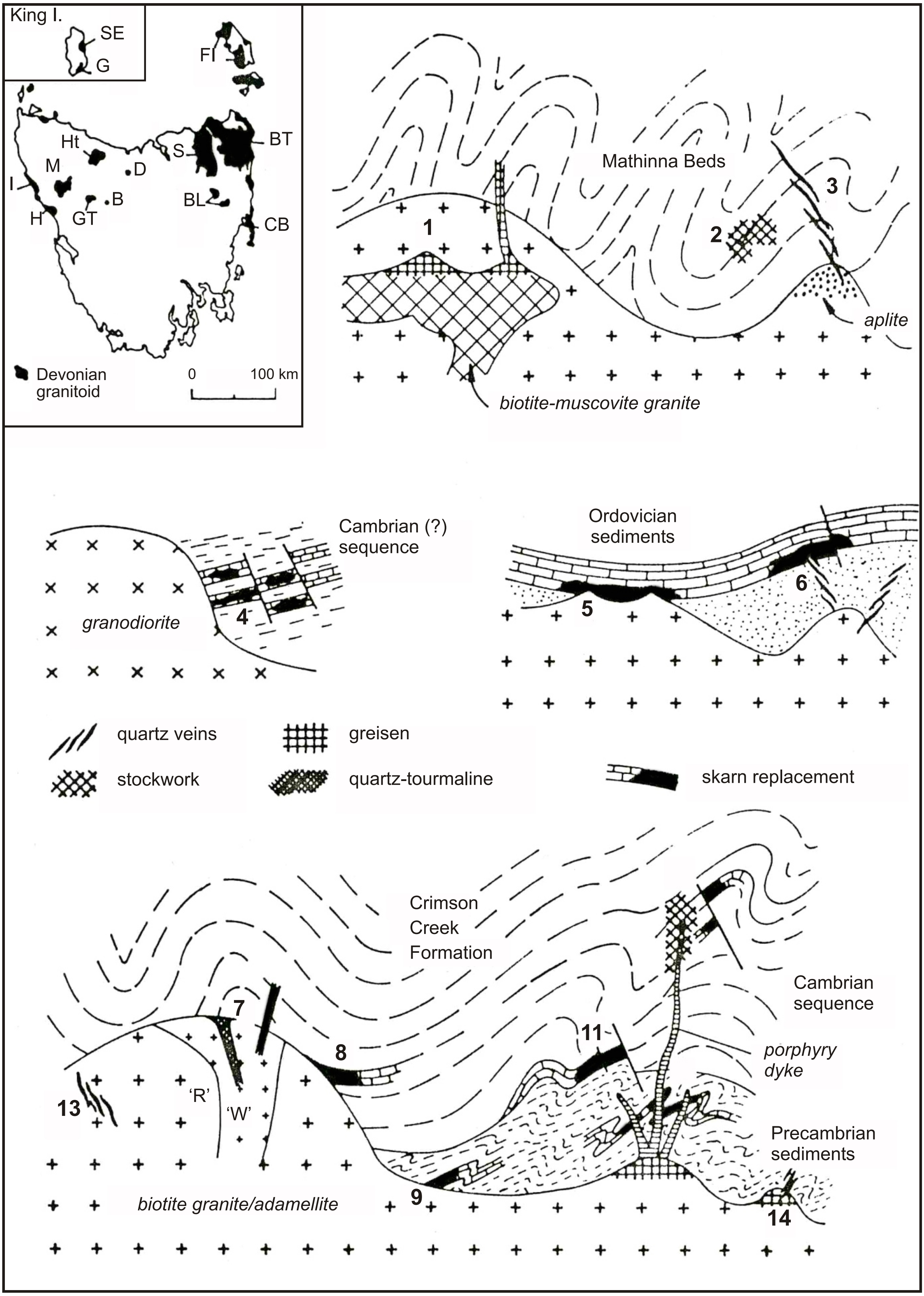


Figure 12

Sketches illustrating genetic types of tin and tungsten deposits in Tasmania.
The map also shows the locations of major granites. See page 35 for key to granites and deposits.

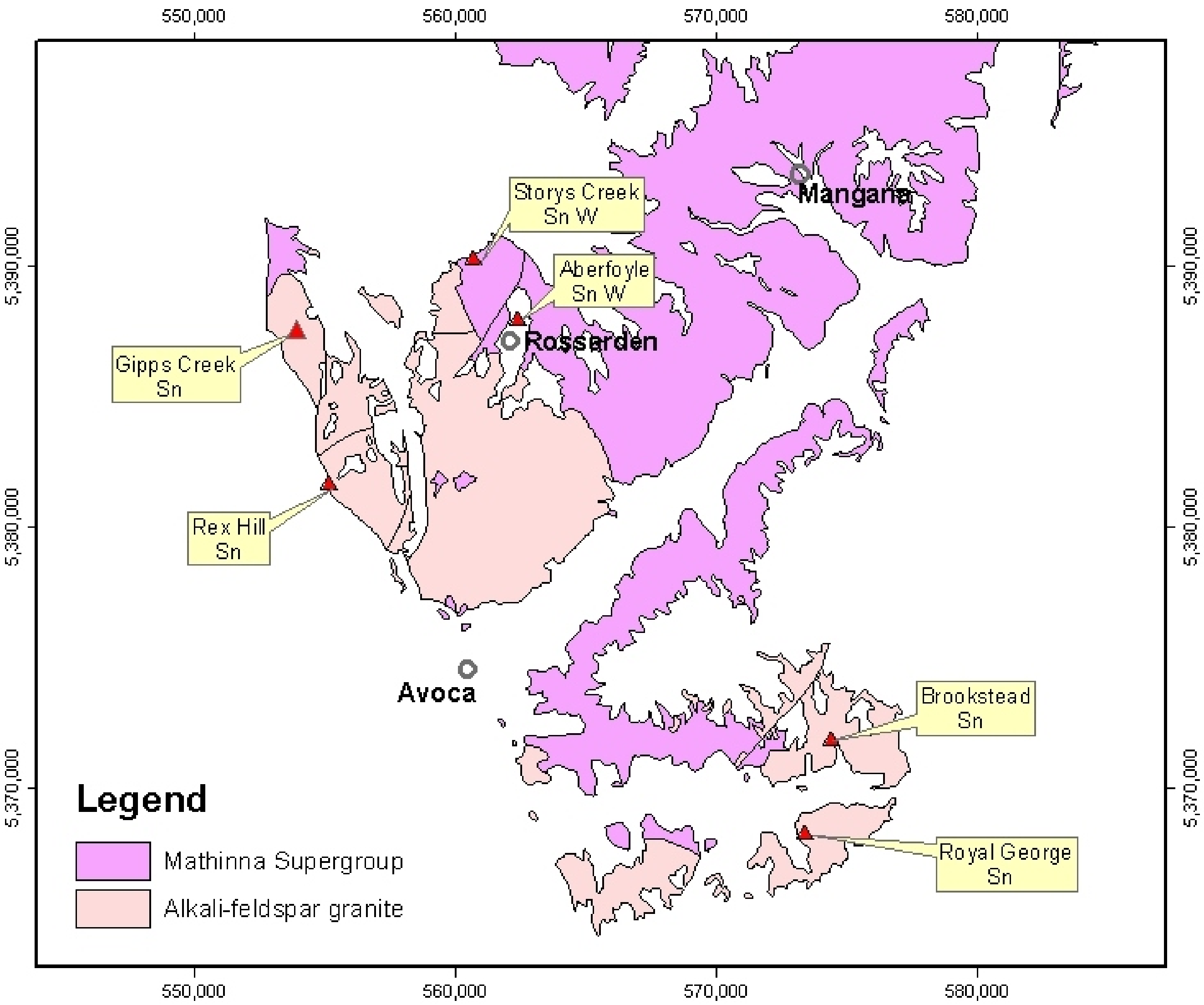


Figure 13

Simplified geological map of Rossarden area, showing the main tin-tungsten deposits.

Key to deposits and granites, Figure 12

Mine/deposit:

1. *Anchor mine: cassiterite in greisen*
2. *Great Pyramid mine: cassiterite in sheeted quartz vein system*
3. *Aberfoyle, Lutwyche and Storeys creek mines: cassiterite and wolframite in quartz veins*
4. *King Island Scheelite mine: scheelite in skarn*
5. *Kara mine: scheelite in skarn*
6. *Shepherd and Murphy mine: cassiterite, wolframite, molybdenite and bismuthinite in quartz veins*
7. *Federation mine, Maynes mine, South Heemskirk: cassiterite in greisen, quartz-tourmaline-topaz alteration, veins pipes and breccias (r = red granite, w= white granite)*
8. *Mt Lindsay prospect: cassiterite and Sn-bearing silicates in skarns*
9. *St Dizier prospect: tin in skarn*
10. *Cleveland mine: cassiterite, stannite and chalcopyrite in replacement lenses and wolframite, cassiterite and fluorite in quartz vein stockwork*
11. *Renison mine: cassiterite in carbonate replacement bodies and fracture system*
12. *Mt Bischoff: cassiterite in replacement bodies and altered porphyry dykes*
13. *Interview River mine: wolframite and scheelite in quartz greisen veins*
14. *Oakleigh Creek mine: wolframite in quartz veins.*

Major Devonian granites:

B = Birthday Granite	*BL = Ben Lomond Granite*	*BT = Blue Tier Batholith*	*CB = Coles Bay Granite*
D = Dolcoath Granite	*FI = Flinders Island Granite*	*G = Grassy Granodiorite*	*GT = Granite Tor Granite*
H = Heemskirk Granite	*Ht = Housetop Granite*	*I = Interview Granite*	*M = Meredith Granite*
S = Scottsdale Granite	*SE = Sea Elephant Granite*		

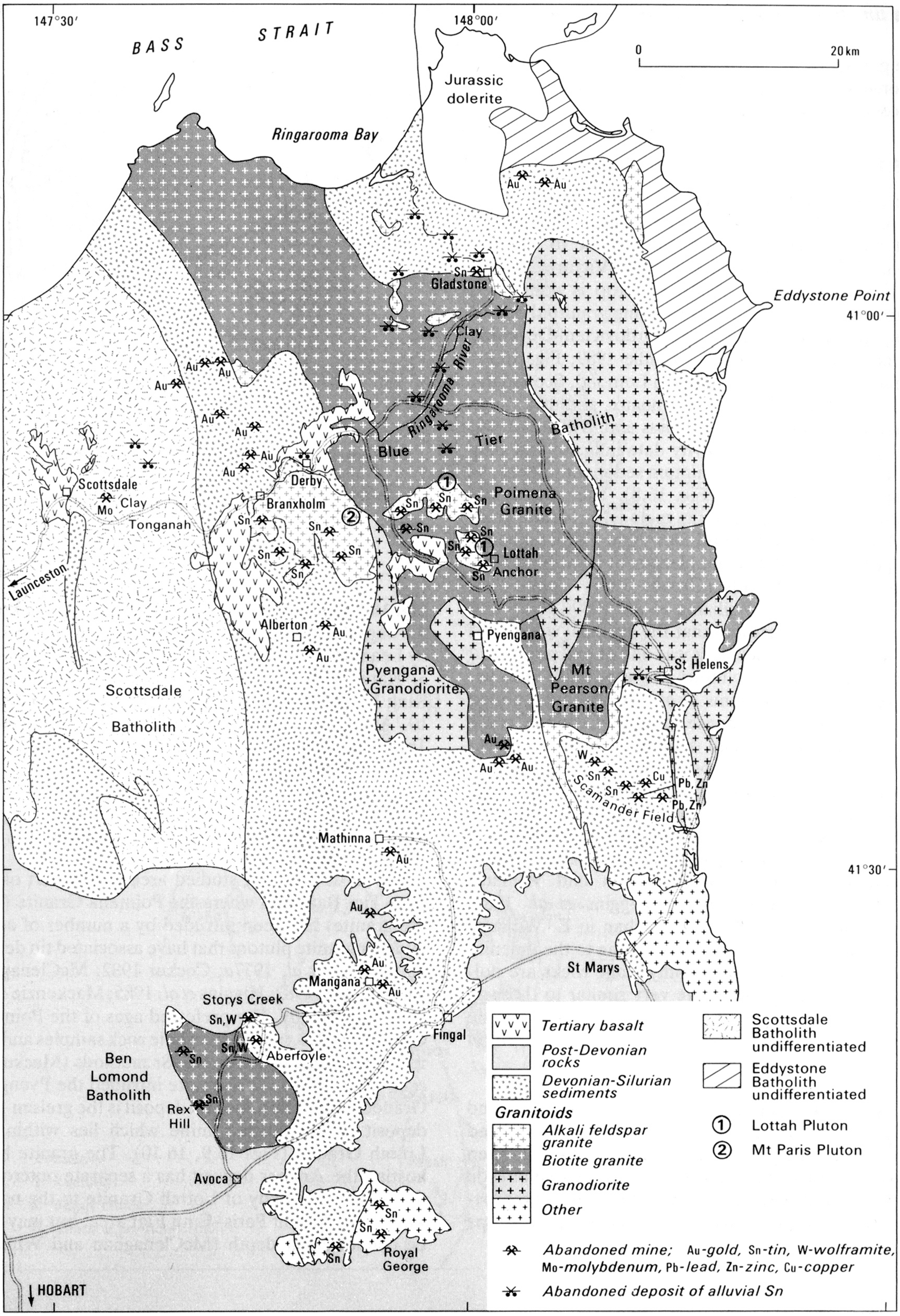
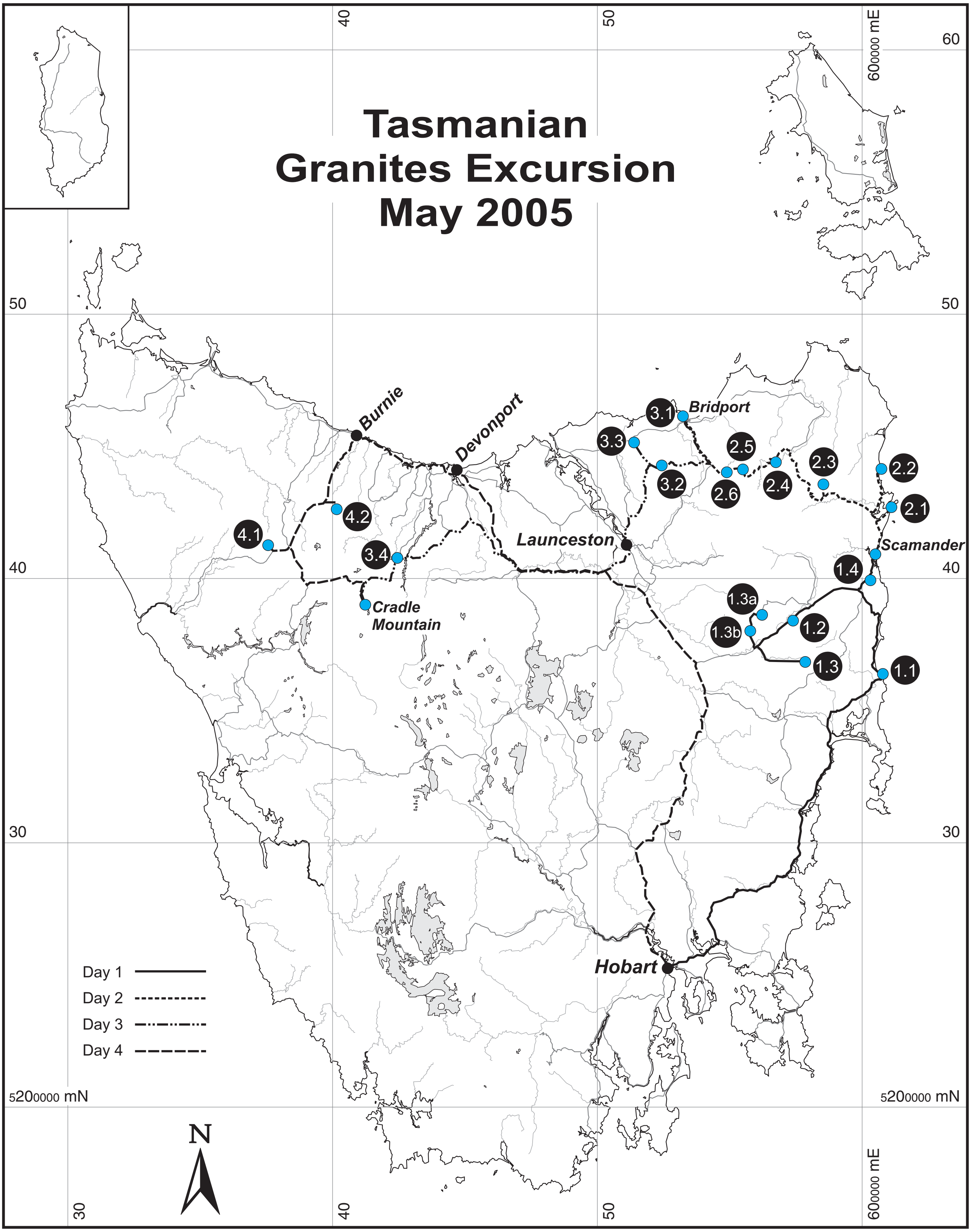


Figure 14

Simplified geological map of northeast Tasmania, showing the major batholiths, tin-tungsten and clay deposits (from Solomon and Groves, 1994).

Detailed Itinerary



Day 1

Stop 1.1

Garnet-biotite adamellite/granite, Bicheno, (608 450 mE, 5 362 420 mN)

The adamellite/granite at Bicheno contains two main phases: a coarse-grained, porphyritic phase, and a finer grained, equigranular phase. The former appears to intrude the latter. There are also some small aplite/pegmatite dykes, quartz veins, irregular mafic clots and some rounded mafic xenoliths. Quartz-tourmaline nodules of different sizes are locally abundant.

The porphyritic granite contains abundant orthoclase megacrysts to ~100 mm in size. It consists of quartz, K-feldspar, plagioclase and red-brown biotite, tourmaline, white-mica, pale green biotite, chlorite, topaz and andalusite. Accessory minerals are zircon, apatite, ilmenite, monazite and xenotime. The K-feldspar megacrysts are commonly tabular and may include all other phases. K-feldspar is perthitic, and the plagioclase has generally uniform cores (An_{40-45}), with normal to oscillatory-zoned rims (An_{20-10}).

Mafic clots, usually irregular to tabular bodies to about 0.1 1 m, are generally rare and occur near the granite/country rock contacts. They contain abundant biotite with minor amounts of garnet and cordierite. The garnet (dark pink, probably almandine) has been extensively replaced in places by pale-coloured biotite. Cordierite is locally prominent, occurring as euhedral phenocrysts, in places being replaced by andalusite (Cocker, 1977). The clots are probably hornfelsed metasedimentary rocks similar to the contact metamorphosed country rocks in other parts of the northeast.

Medium to coarse-grained, equigranular to porphyritic granite variants are present. The granite is characterised by complex textural variations with gradational to sharp boundaries between textural types, suggesting auto-intrusive relationships. In particular, it is notable that garnet-bearing mafic clots occur near the roof of the coarser body, near its contact with the finer grained, equigranular phase. Dolerite dykes (100–150 mm wide) of possible Devonian age can also be observed.

Stop 1.2 (time permitting)

Tullochgorum fold locality, Mathinna Supergroup (road cutting) (577 581 mE, 5 388 641 mN)

This locality is a good introduction to the Mathinna Supergroup turbidite sequences and the type of folding that typically develops within them. It was initially mapped by Chris Powell for his Ph.D. thesis some 30 years ago.

The road section shows interbedded sandstone and argillite folded into a series of D_2 anticlines and synclines (fig. 15). Folding is close to tight in style with steep or overturned eastern limbs. Cleavage dips to the southwest and the sense of transport is towards the northeast. In the northeast part of the section there is a folded zone, 10–15 m thick, which has a series of three anticline and syncline pairs. In the southwest part of the section there are some distinctly open flexures (Powell, 1967).

Warning: Beware of fast-moving vehicles at this stop.

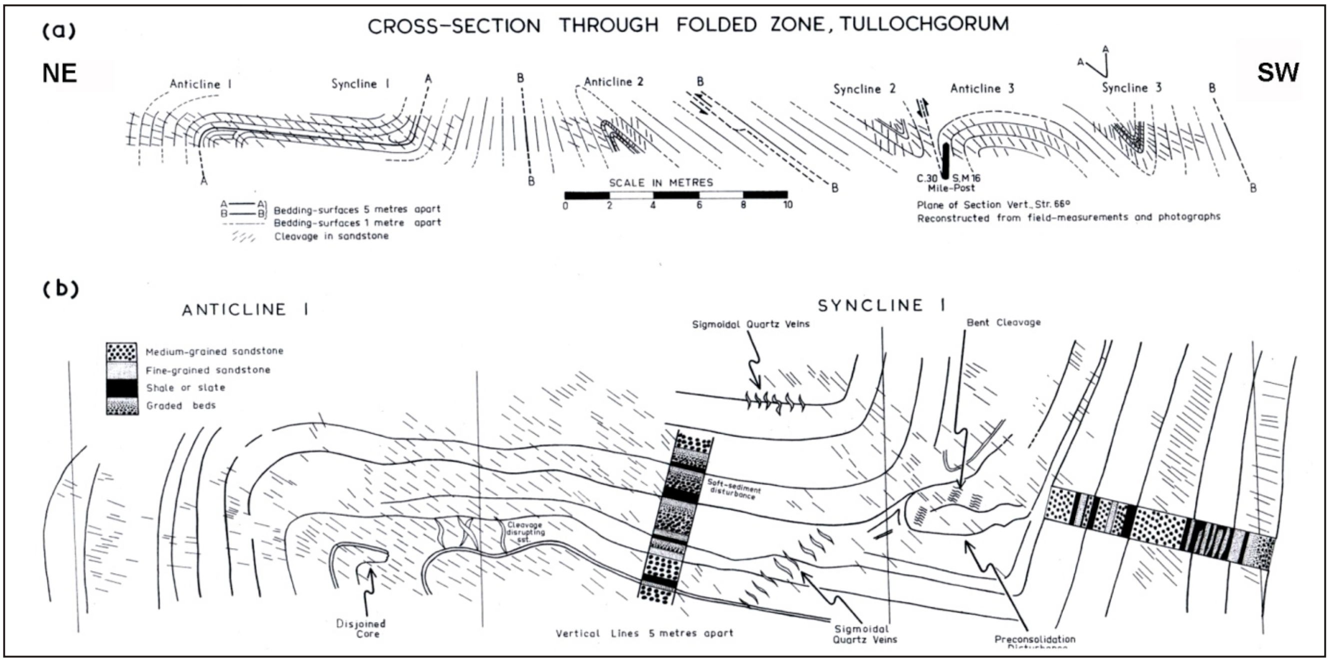


Figure 15. *Sketch of the geology and structure along the Tullochgorum road section (from Powell, 1967).*

Stop 1.3

Royal George tin mine (573 520 mE, 5 368 320 mN)

History and exploration

The Royal George tin mine is the largest deposit in the St Pauls tin field, some 18 km southeast of Avoca in northeast Tasmania. It was found in the 1880s but mined mostly between 1911 and 1922, producing 1100 tonnes of tin from 170 000 tonnes of ore (@ 0.65% Sn). This was produced from open stopes and two underground levels (Ruxton, 1984).

The mine was drill-tested by BHP in 1957, and by the Cornwall Coal Company in 1965; the latter company also deepened the shaft and added two levels (Urquhart, 1968). Reserves, calculated from 18 drill holes and channel samples by CRAE, were estimated as about 1.2 Mt @ 0.34% Sn at 0.25% cut-off, with the potential for 1.7 Mt (Purvis, 1981). The last recorded exploration was by Billiton in 1984, which reviewed the mining potential and calculated estimated tailings reserves as 0.17 Mt @ 0.25% Sn, with Zn and Ag credits (Ruxton, 1984).

Geology

Tin deposits in the St Pauls tin field comprise some east-west and NW-SE trending lodes in the Devonian Avoca Granite, part of the Ben Lomond Batholith, and some associated Cainozoic alluvial deposits (Ruxton, 1984).

The Royal George deposit is about 250 m long by 200 m deep and 20 m wide, with patchy mineralised extensions one kilometre to the south (Ruxton, 1984). The deposit contains a series of greisen veins hosted by medium-grained biotite granite, with dykes of fine-grained quartz feldspar porphyry. There are also some aplite dykes and some pegmatitic segregations (fig. 16), plus locally abundant tourmaline nodules (<200 mm) mostly in the porphyry. There are also some minor tourmalinised breccias, with fine-grained green tourmaline.

Figure 16

Sub-vertical and sub-parallel porphyritic aplite dykes, alkali feldspar granite with some pegmatitic segregations, and cassiterite-sulphide-uranium bearing, quartz-rich greisen, Royal George mine.

The greisen lodes are a series of *en echelon* quartz-topaz-sulphide-tourmaline-fluorite-mica veins which strike 310–315°T and dip 75–82° SW, and are 0.3–1.5 m wide (Ruxton, 1984). They cut the host granite structures obliquely and become more sericitic with depth and near their peripheries. The lodes contain up to 20% sulphide minerals, including pyrite, arsenopyrite, sphalerite and chalcopyrite. Cassiterite is mostly disseminated but also occurs as coarse crystals in central lode fractures, with quartz and colourless topaz crystals. Tourmaline is disseminated as dark brown to black crystals, resembling cassiterite. Uranium is enriched in the periphery of the tin mineralisation, and mostly occurs as coarse-grained metatorbernite in joints and fractures. The primary uranium minerals are unknown. The granite contains <40 ppm U (Ruxton, 1984) and the ore has minimal radioactivity.

Safety issues

There are some dangerous underground workings and open stopes in the mine workings. The underground workings should not be entered because of the high potential for radioactive radon build-ups and the unstable wall rocks. The open stopes are also highly radioactive in some areas and have dangerous overhangs that should be avoided. The rim of the open cut is largely unstable and should be approached with care. Visitors should only enter the mine with great caution and also be wary of collecting too much material, as the uranium-rich samples can be dangerous to store in some situations. These samples should be handled carefully, and hands washed thoroughly before eating or smoking. Avoid ingesting the potentially radioactive dust.

Stop 1.3a (optional)

Aberfoyle tin-tungsten mine (562 500 mE, 5 388 000 mN)

The Aberfoyle, Storys Creek and Lutwyche vein systems

The following has mainly been drawn from Solomon and Groves (1994). These mines are within a few kilometres of each other and are located about 20 km from the township of Avoca (fig. 13). The veins at Aberfoyle were discovered in 1916 and production began in the same year, while those at Storys Creek were worked for tin from 1891. The highest recorded production of hard-rock tin in northeast Tasmania was from the Rossarden area, including Aberfoyle (15 490 t), Storys Creek (1980 t), Royal George (1141 t) and Rex Hill (651 t). All mines in the area closed in early 1980s (De Graaf, 1983).

These three sheeted vein systems occur in sedimentary rocks of the Mathinna Supergroup overlying the surface of the Ben Lomond Granite in the Rossarden area. The granite is mostly a coarse-grained, porphyritic biotite granite with local textural variations, and contains muscovite, tourmaline and topaz (Blissett, 1959; Clayton, 1981). The locations of the sheeted vein systems at Aberfoyle and Storys Creek are clearly related to cupolas of greisenised aplite; muscovite from the aplite has been dated by K-Ar methods as 367 Ma.

The Aberfoyle and Lutwyche systems have been studied in some detail (Lyon, 1957; Edwards and Lyon, 1957; Kingsbury, 1965; Hellsten, 1979; Wilkins and Ewald, 1981; Halley, 1982; Collins *et al*., 1984). There are eight main and many subsidiary veins in the Aberfoyle deposit. The veins dip westerly at 60–65° in the upper levels and 45–50° in the lower levels. The veins extend along strike for about 700 m, are up to 1.5 m thick, and have been mined over a vertical depth of about 320 metres. Only minor veins intersect the cupola, the main veins being in tightly folded, interbedded mudstone and quartzwacke of the Mathinna Supergroup. The Lutwyche system lies 800 m northeast of Aberfoyle where the granite is >600 m below the topographic surface (Collins and others in Williams *et al*., 1989). Mapping on the 13 level at Lutwyche (Hellsten, 1979) has shown that the three main vein systems of the Lutwyche deposit are independent of fold axes and other structural features, and that there is no displacement on vein walls; they thus appear to be tensile fractures formed under conditions of low differential stress. The veins show clear evidence of repeated opening.

The veins lie within a metamorphic aureole that is detectable as much as 600 m from the granite contact, indicated by graphitic spotting in the slate. The mineral assemblage at the contact includes K-feldspar, andalusite and quartz. The veins mainly comprise quartz with wolframite, cassiterite, chalcopyrite, sphalerite, pyrite, pyrrhotite, arsenopyrite, F-rich muscovite (as selvedge), topaz, fluorite, siderite, and late calcite. Except in vugs in the central parts of veins, the quartz is milky and poorly crystalline, either because it precipitated rapidly or has been recrystallised. There are numerous hair-line fractures parallel to the vein walls. A simplified paragenetic sequence, established at Aberfoyle by Edwards and Lyon (1957), is as follows:

Early Cassiterite, wolframite, muscovite, topaz, apatite, fluorite, quartz; arsenopyrite, pyrite, quartz; chalcopyrite, sphalerite, stannite, quartz.

Late Galena, tetrahedrite, native bismuth, scheelite, fluorite, siderite, quartz(?), calcite.

The Battery Vein at Lutwyche, the earliest of the vein sequence, is characterised by wolframite, while the later footwall veins contain dominantly cassiterite. At Aberfoyle, the Sn:WO_3 ratio increases upward, from 3:1 below 4 level to 15:1 in the upper levels (Kingsbury, 1965). The Aberfoyle cupola is capped by quartz known as the Contact Vein in similar fashion to the cupola at Panasqueira, Portugal (Kelly and Rye, 1979). Very irregular and local quartz veins are seen throughout the country rock and are thought to be of metamorphic/deformational origin, formed prior to granite intrusion.

Fluid inclusion data

Wilkins and Ewald (1981) and Halley (1982) recognised three fluids in quartz, cassiterite and other early phases, namely CO_2-rich, a high salinity fluid, and a low salinity fluid. T_h values ranged from about 300 to 490°C, the values above 450°C being obtained only from inclusions with the highly saline fluids. T_h values from 200 to 100°C were recorded in the late scheelite and fluorite assemblage. Details of the three fluid types are as follows:

1. CO_2 (v) + CO_2 (l) + aqueous solution, with variable CO_2 content, from a few mole per cent to 70 mole per cent. Densities bimodal at about 0.25 (vapour phase homogenisation) and 0.7 g/cm^{-3} (liquid phase homogenisation); salinities <5 wt% NaCl equivalent (from clathrate melting); T_h from 250 to 420°C. Burlinson *et al*. (1982) determined by laser Raman microprobe that the gas contents of all but one of their samples were >93 mole% CO_2 with

minor N_2, CH_4 and H_2S, although the minor constituents could not be identified by Hoffmann *et al*. (1988) using mass spectrometry.

2. Aqueous solution + vapour, probably with CO_2, although no clathrates formed on cooling. Low salinity. Some are vapour-rich; TV from 240 to 390°C.

3. Aqueous solution + vapour + daughter salts (mainly NaCl and one other), with $T_{h.\,L} > T_m$ for NaCl crystals and other solids insoluble, T_h from 280 to 490°C; salinities 35 to 40 wt% NaCl equivalent.

4. Vapour-dominant, vapour-homogenising inclusions co-existing with type (3) and having the same T_h range.

The three fluid types, which are assumed to have coexisted, may have been derived from the original magmatic fluid by phase separation during transport upward from the melt. Inclusions in cassiterite, many lying along crystal growth zones, consist only of types (1) and (2); cassiterite was not deposited from fluid (3). The CO_2-rich inclusions roughly define immiscibility surfaces in the CO_2–H_2O–NaCl system, and pressure estimates using the data of Takenouchi and Kennedy (1964) and Gehrig *et al*. (1979) range from 150 to 500 at temperatures of 320 to 360°C. Similar low pressures are indicated for the immiscible, higher temperature, salt-rich fluids, using the data of Sourirajan and Kennedy (1962). These inclusions occur only in fractures in quartz.

Stable isotope data

^{18}O values for vein quartz range from 10.5 to 14.0‰, giving a range of equilibrium water compositions from 6.7 to 8.1‰ for eight samples at known T_h, typical of magmatic values elsewhere (Taylor, 1979). However the Mathinna Supergroup near to, and at some distance from, vein systems has values like those of the vein quartz, so that a groundwater contribution cannot be eliminated. Quartz in supposed metamorphic veins in the country rock has δ^{18}O values from 15.0 to 15.3‰ but there are no thermometric data with which to calculate fluid compositions.

δ^{34}S values for pyrite, chalcopyrite, and sphalerite range from 3.6‰ to -3.3‰ and, assuming H_2S is the dominant sulphur species, indicate solution sulphur compositions of about 2.5 to -5.0‰ at 350°C. These are similar to values in greisens and other magma-related deposits in northeast Tasmania.

Field observation

There are no good surface exposures at the Aberfoyle mine, but good ore samples can be found in the large tailings and rock dumps. These show the general nature of the mineralisation and the veins. The host Mathinna Supergroup rocks, comprising mainly siltstone and slate, can also be seen. The old workings are not accessible.

Stop 1.3b (optional)

Ben Lomond granite
(557 932 mE, 5 380 423 mN)

This is a large body which crops out well in the hilly country north of Avoca and south of the dolerite-capped Ben Lomond plateau. It is crudely elongate in a NNW direction (15 km 9 km), parallel to the regional fold trend of the Mathinna Supergroup.

The granite displays a wide range of mesoscopic textures, from fine to coarse-grained, and from aphyric to markedly porphyritic in quartz and alkali feldspar. Although there is a tendency for the fine-grained variants to occur near the margins or roof, the spatial distribution of these textural types may be extremely intricate even at outcrop scale, and no consistent age relationship between them can be established. It seems likely that they simply represent variable crystal abundances in the same highly viscous granitic melt, caused by mechanical processes during intrusion and solidification.

Mineralogically the granites consist principally of quartz, alkali feldspar, plagioclase (mainly albite but with some cores as calcic as An_{27}) and minor biotite, often altered to chlorite. Tourmaline, zircon, fluorite, topaz and secondary muscovite and sphene occur as accessory minerals.

The Ben Lomond granite is associated with some major tin-tungsten vein deposits (e.g. Rex Hill, Aberfoyle and Storys Creek) in northeast Tasmania.

Stop 1.4 (time permitting)

St Marys Porphyrite
(602 594 mE, 5 399 865 mN)

(from Turner et al., 1986)

The St Marys Porphyrite and several large, related granodiorite dykes are inferred to be co-genetic with the Blue Tier Batholith. The St Marys Porphyrite consists of two petrographically similar units: zone A, interpreted as a thick (800 m) pile of compacted tuff; and zone B, which is considered to be a vesiculated intrusive rock in the upper part of the feeder system of the porphyrite (fig. 17). Zone A rests conformably on a laterally discontinuous, thin breccia zone that conformably overlies the Mathinna Supergroup.

In general, grain size increases progressively with increasing height above the base of the porphyrite body. Accompanying the changes in grain size are other changes in matrix texture. The mainly incipiently recrystallised matrix in the basal zone changes rapidly upwards to become a fine-grained, equigranular mosaic of evenly distributed, equant grains of potash feldspar, quartz and plagioclase. With increasing height, the potash feldspar coalesces to form finely granular aggregates which surround and isolate other matrix grains. The boundaries of mineral grains in the coarse fraction become progressively more pitted and irregular as the matrix changes and pyroxene grains are largely replaced by hornblende. The basal zone locally shows well preserved vitroclastic textures but minor groundmass recrystallisation including 'snowflake' texture is widespread.

In general, the St Marys Porphyrite is crystal-rich with phenocrysts comprising up to 58% of the volume. Main phenocrysts in order of abundance are plagioclase (28%), quartz (17%), biotite (8%), augite and hypersthene (4%), and alkali feldspar (2.5%). Phenocrysts are typically fragmented, but less fragmentation is evident in the high level feeder zone B.

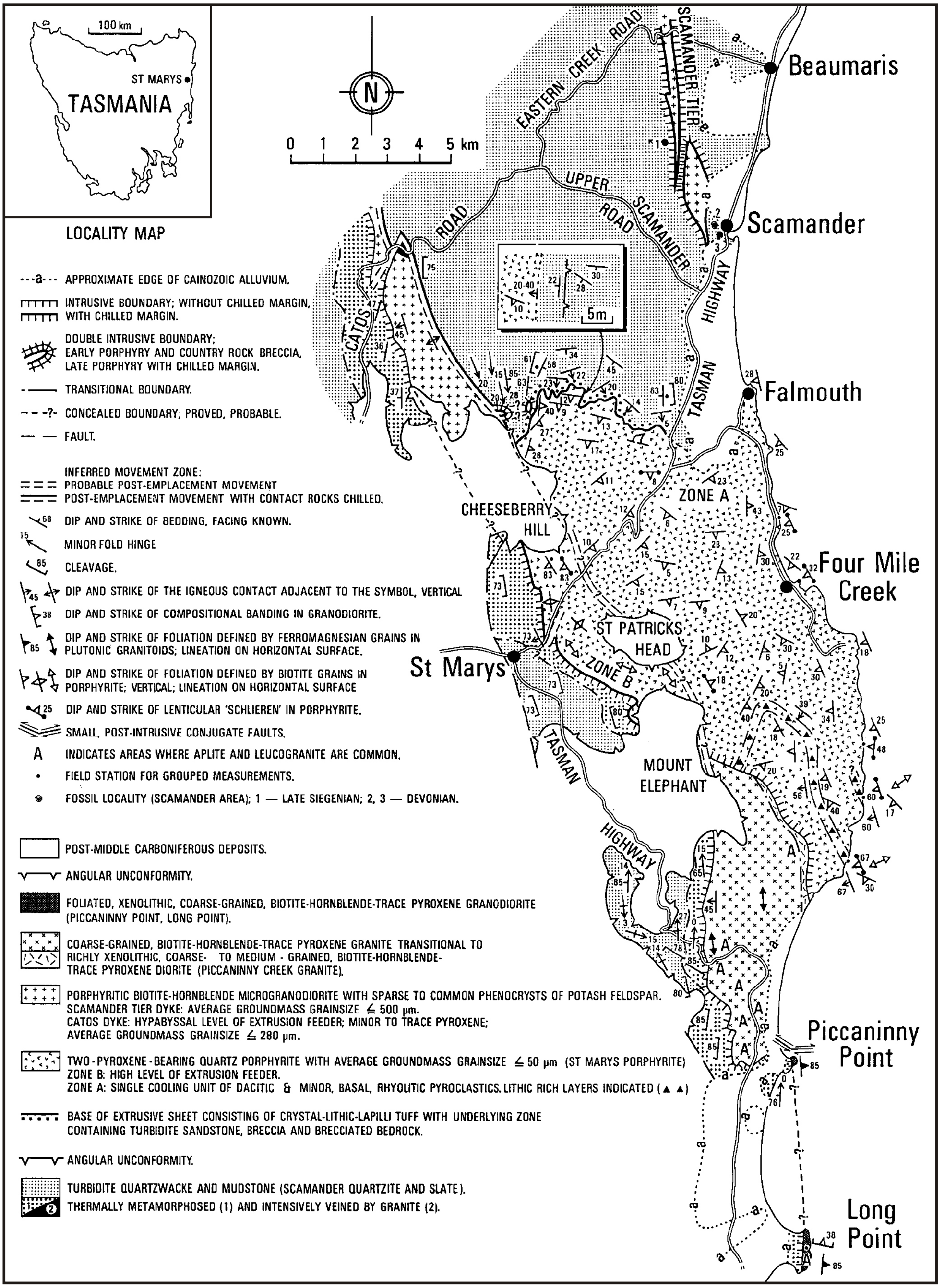


Figure 17. *Geology of the St Marys district (after Turner* et al.*, 1986*).

DAY 2

Stop 2.1

Akaroa Granodiorite
(610 723 mE, 5 426 619 mN)

This rock type corresponds to the Akaroa Granodiorite of Cocker (1977) and occurs in the areas bordering the mouth of Georges Bay. Boundaries are exposed with the biotite granite where they can be seen to be gradational over several metres.

The granodiorite is a medium to coarse-grained medium dark grey rock with sparse small K-feldspar phenocrysts (<25 mm) and rare larger ones (up to 70 mm). The rock generally has a speckled appearance due to the dispersed biotite flakes. Biotite-rich layers are common and widely dispersed biotite clots are present. Rare finer grained dioritic xenoliths also occur. In some areas the biotite-rich layers have sharp boundaries on one side and pass into zones with abundant small K-feldspar phenocrysts resembling the abundantly porphyritic biotite granite on the other side.

In thin section the rock consists of quartz, plagioclase, K-feldspar and biotite with accessory muscovite, apatite and zircon. The plagioclase has sericite and sometimes epidote alteration in the cores with clear zoned rims. The K-feldspar is microcline and has inclusions of plagioclase, biotite and quartz. The biotite is extensively altered to chlorite. Modal analyses presented by Cocker (1977) show that the composition ranges from granodiorite to adamellite.

The granodiorite at the quarry is mineralised with mainly pyrite, pyrrhotite?-chalcopyrite and molybdenite occurring as disseminations in fractures. This is the only known location where a granodiorite body is mineralised in the region. The gold content of the granodiorite is not known.

Stop 2.2

Gardens Granodiorite
(607 250 mE, 5 442 550 mN)

The Gardens Granodiorite consists of euhedral to anhedral amphibole and biotite, plagioclase and intergranular K-feldspar and quartz. The amphibole ranges in composition from actinolite to hornblende with core regions of crystals being actinolite while marginal zones and the euhedral crystals are hornblende. Among the more basic specimens, clinopyroxene is present as small anhedral grains and patches associated with the actinolite. The biotite and amphibole occur intermingled in patches as well as in isolated crystals and the amphibole contains inclusions of biotite and small quartz grains. The plagioclase frequently contains sericite-altered core regions that have sharp boundaries with clear rim zones. The core regions have a compositional range of $An_{81.2-68.6}$ which does not overlap with that of the rims ($An_{52.4-38.5}$). Accessory minerals are apatite, sphene, ilmenite and zircon. Prehnite is developed as thin strips along the cleavage of the biotites. Minor chloritic alteration of the biotite is also present. The Gardens Granodiorite contains hornblende, sphene, allanite and very minor amounts of clinopyroxene, all of which are characteristic minerals of I-type granite (Chappell and White, 1992). Magnetite, which is a characteristic of I-type granite, occurs only in parts of the Pyengana Pluton (part of the Gardens suite) and in the Lisle body.

Stop 2.3

Anchor mine
(584 800 mE, 5 435 300 mN)

Setting

The Anchor tin mine is situated on the southern slopes of the Blue Tier, 25 km northwest of St Helens in northeast Tasmania. It is located in the greisenised top of an alkali feldspar granite (Lottah Pluton/Anchor Granite) which intrudes an extensive body of biotite adamellite/granite (Poimena Pluton). Both granitoid units are parts of the Blue Tier Batholith (fig. 18).

History and production
(after Newnham in Turner and Taheri, 1990)

Production first commenced at the Anchor mine in 1875, and several intermittent phases of mining resulted in the production of 3500 tonnes of cassiterite from open-cut operations up until mining ceased in 1942.

The main period of operations was from 1898 to 1914, when a substantial open cut mine was developed in low-grade ore, at times feeding as much as 100 000 tpa of ore to a very large gravity mill. Head grades during that period are thought to have been approximately 0.25% Sn with 80% tin recoveries.

In total, these earlier operations probably treated around two million tonnes of ore at an average head grade of 0.2–0.3%. A similar tonnage of soft overburden was probably removed by hydraulic sluicing.

The main operations ceased in 1914, mainly because of low tin prices, lack of water, and increasing overburden ratios.

Aberfoyle Tin NL completed several major core drilling programs during the mid-1960s, which indicated the presence of a substantial deposit of low-grade tin mineralisation.

In 1966, Aberfoyle estimated the remaining resource as approximately 1.3 Mt of 0.35% Sn. In 1981, following drilling programs, Renison Limited estimated a total resource of 0.63 Mt @ 0.49% Sn.

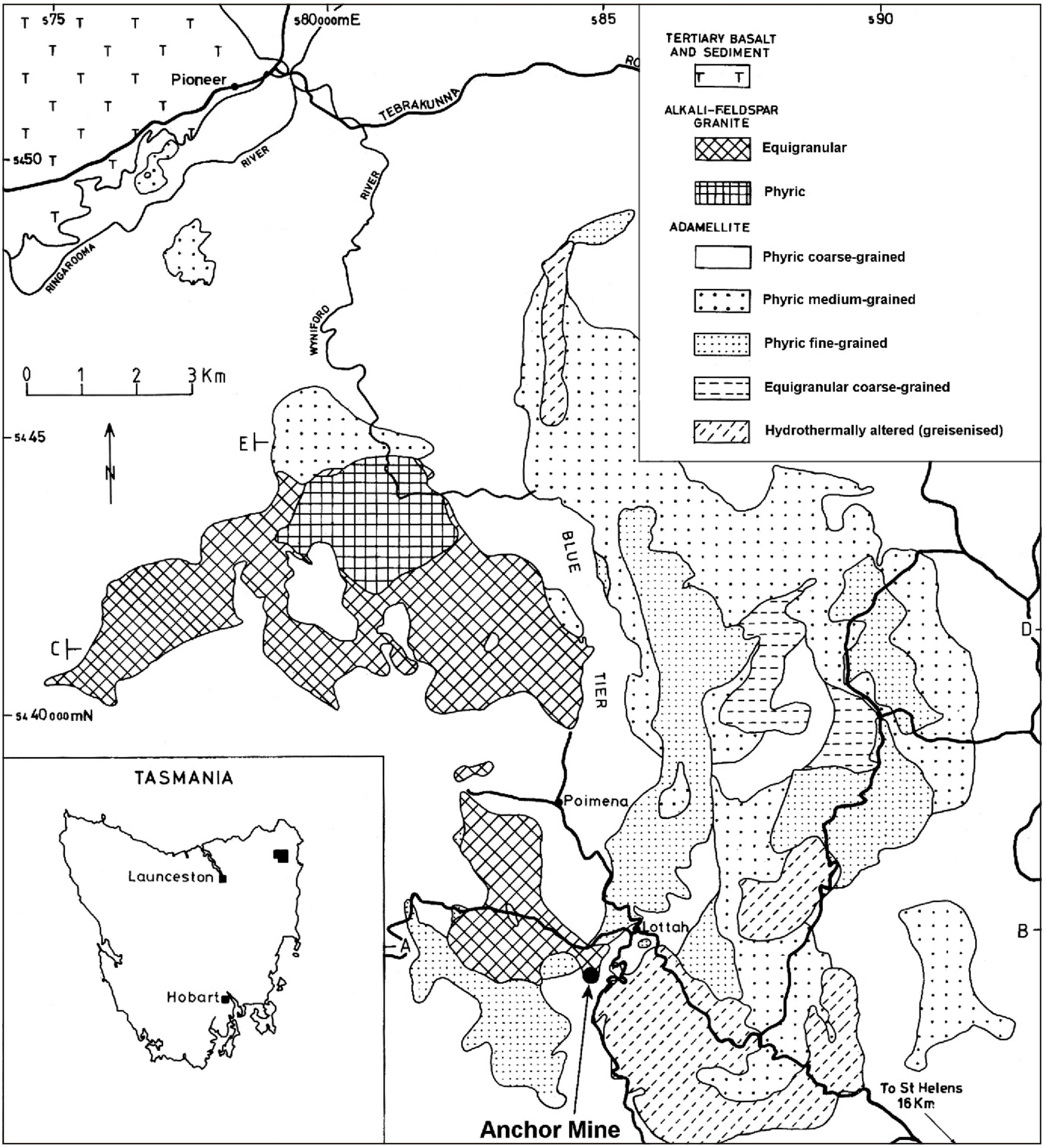


Figure 18

Geological map of the central Blue Tier area, showing the position of the Anchor tin mine, northeast Tasmania.

In a feasibility study undertaken by Spectrum Resources Australia P/L in 1988, a resource of 438 000 tonnes at 0.62% Sn was estimated. Spectrum Resources developed an underground mine and gravity concentrator in 1988/1989, but the operation closed in 1991 following a dramatic fall in the price of tin. Some 722 tonnes of cassiterite concentrates were recovered between 1989 and 1991.

Lottah Alkali-Feldspar Granite

The two main alkali-feldspar bodies in the Blue Tier Batholith are the Lottah and Mt Paris alkali-feldspar granites lying near the centre of the batholith. These bodies, together with the Lottah Alkali-Feldspar Granite, have been described as having a sheet-like form (Gee and Groves, 1971), but in the case of the Lottah Pluton detailed mapping (McClenaghan and Williams, 1982) showed that a steep-sided dome is more likely.

The Lottah Alkali-Feldspar Granite (McClenaghan and Williams, 1982) includes equigranular quartz and K-feldspar porphyritic varieties which have gradational relationships with each other. The granites consist dominantly of K-feldspar, albite and quartz with the mica content being less than 7%. The K-feldspar is present in perthitic and non-perthitic form, with the non-perthitic feldspar commonly overgrowing anhedral K-feldspar cores. The perthitic K-feldspar is invariably more sodic than the non-perthitic K-feldspar (McClenaghan and Williams, 1982). In the porphyritic granites the dark mica is annite or siderophyllite containing inclusions of zircon and apatite. This mica is replaced to varying degrees by zinnwaldite in the equigranular granites. Muscovite is minor and occurs as an alteration of K-feldspar or derived from the breakdown of annite. Additional accessory minerals are fluorite, cassiterite, topaz and tourmaline.

The low temperatures calculated from equilibrium feldspar compositions in this granite suggests sub-solidus mineral-fluid reactions (McClenaghan and Williams, 1982) and zones of greisenisation in the Lottah area (Groves and Taylor, 1973) indicate local metasomatic alteration of the granite.

Petrogenesis of the Lottah granite

McClenaghan and Williams (1982) considered that the Lottah Alkali-Feldspar Granite could have been derived from the Poimena Adamellite by crystal fractionation or by a combination of restite-unmixing and crystal fractionation. Higgins *et al*. (1985) suggested that the derivation process was one of fractional crystallisation combined with variable metasomatism following aqueous fluid saturation of the magma. Mackenzie *et al*. (1988) concluded that the Lottah Alkali-Feldspar Granite formed from a different magma to the Poimena Adamellite. They based this conclusion on the lack of continuity of trends on variation diagrams, the 10 Ma difference in emplacement age, the difference in initial ^{87}Sr ratios, and the difference in Nd values between the two granite types. Mackenzie *et al*. (1988) considered that the compositional variation in the Lottah Alkali-Feldspar Granite is consistent with fractional crystallisation of a felsic peraluminous melt rich in F, Li and B, to produce even more peraluminous residual melts, progressively further enriched in these components. They did not consider that there was a significant metasomatic component to the chemical variation.

Geology

The following summary description of the geology of the Anchor mine area is taken from Ross (1983) and McKeown (1993).

The Anchor tin deposit occurs within the polyphase Blue Tier Batholith (Late Devonian), which is subdivided into early porphyritic adamellite, greisenised adamellite and late alkali granite intrusions. The alkali granite is an equigranular leucocratic assemblage of quartz, albite, orthoclase and biotite. Soil geochemistry demonstrates the stanniferous character of the alkali granite.

Tin mineralisation (as cassiterite) is hosted in altered alkali Lottah Granite beneath an undulating, generally horizontal roof contact with overlying Poimena Adamellite. The ore consists of two lenses (A and B). A lens covers an area of approximately 40 000 m^2 and is up to 20 m thick. B lens lies at the lower level to the south of, and partly beneath A lens and covers an area of about 10 000 m^2 and is up to 15 m thick. The lenses show antiform and synform structures developed parallel to the strike. In general, the intensity of alteration decreases with depth beneath the roof contact and two alteration types (i.e. greisenised granite and granular greisen) are recognised. Greisenised granites retain original feldspar and granitic textures, while granular greisens consist of granular aggregates of quartz-topaz-mica.

The intensity of alteration is reflected in the colour of the rocks. The unaltered Lottah Granite is a cream-white rock which, with alteration, progressively changes in colour through grey and light green in greisenised granite, to dark grey-green and black in the true greisen. Darker greisens tend to contain more cassiterite than the lighter coloured greisens and greisenised granites. Cassiterite commonly occurs as disseminated grains normally greater than 40 μm, but individual crystals up to 15 mm have been observed in the underground workings. The greisens also contain traces of sulphide minerals and wolframite. Some of the sulphides include arsenopyrite, galena, sphalerite, pyrrhotite, pyrite, tetrahedrite, molybdenite, bismuthinite and chalcopyrite.

Alteration occurred in at least two phases. The first stage was probably late magmatic and involved the replacement of feldspars and primary micas by topaz and yellow-dark green siderophyllite, accompanied by cassiterite, purple fluorite and apatite. The second stage can be regarded as a low temperature post-magmatic hydrothermal phase in which some of the topaz and siderophyllite was replaced by fine sericite, siderite and trace sulphide minerals.

The alteration assemblages reflect the interaction of a late-stage metal and volatile-rich phase with a related crystallising granitic magma, i.e. second boiling. Alteration of the alkali granite is accompanied by significant increases in H_2O, CO_2, Fe, Mn, F, P, Ga, Rb, Nb, Sn, W, Bi, Mo, Cu, Zn, Ag and decreases in Li, Na, K, Y, Zr and total REE.

Alteration of the overlying adamellites is not as extensive. The porphyritic adamellite was solidified at the time of alkali granite emplacement and displays a narrow zone of alteration ranging from one to ten metres beyond the contact.

The greisenised adamellite is variably sericitised, topazised and carries secondary albite, yet its contact relationships with the alkali granite are not well understood.

Tin grades exceeding 0.2% are restricted to two lenses of granular greisen and minor greisenised granite within the stanniferous zone. The lenses occur immediately beneath the roof contact and are associated with a cupola-like structure approximately 200 m in diameter. Cassiterite occurs as erratic disseminations and aggregates.

Other workers have shown that the alkali granite is probably derived from a peraluminous magma (porphyritic adamellite) by fractional crystallisation. Field relationships and other data support this view.

Many features of the Anchor tin deposit are consistent with magmatic and post-magmatic processes of rare-element (Sn, W, Nb, Ta, Be, Mo, U, REE) formation described elsewhere in the literature. Typically, these deposits are formed in the apical zones of the last intrusives from polyphase intrusive complexes.

Field observations

The adits have been closed and the area partly rehabilitated, but good exposures of granite and mineralisation are apparent. At the mouth of some of the adits the contact between the mineralised Lottah Granite and overlying Poimena Pluton can be observed. Mineralised greisens near the adit mouths exhibit cassiterite, chalcopyrite, bornite, topaz, fluorite, siderite, molybdenite and other ore and alteration minerals.

Stop 2.4 (time permitting)

Derby Museum

The old Derby school, which was built in the late 1800s, has been converted into a museum. The museum exhibits historic records, relics and photographs of the well known tin mines which operated around northeast Tasmania from 1876 to 1940. Revel Munro, the manager of the museum, will provide a short talk on the history of tin mining in northeast Tasmania. The museum overlooks the Briseis mine, one of many mines which worked the Cainozoic alluvial tin deposits, including sub-basaltic deep leads, along the Ringarooma River valley early last century. The area has produced about 24 000 t of Sn metal and has an estimated resource of 0.85 Mt @ 1.4 kg/m^3 tin.

Stop 2.5 (time permitting)

Alkali-feldspar granite, Tulendeena: Mt Stronach Pluton (553 700 mE, 5 441 300mN)

This granite body is a part of the Scottsdale Batholith, and here is a medium-grained, pink, alkali feldspar granite, poor in plagioclase and mafic minerals and locally miarolytic, with medium to large cavities (to 200 mm), containing quartz, orthoclase and minor biotite crystals to 50 millimetres. The pluton hosts some molybdenum mineralisation at the Mt Stronach mine, some six kilometres to the west, and is probably the source of cassiterite and monazite-rich alluvial sediments in the area.

Stop 2.6 (time permitting)

Tonganah clay deposit, Tonganah road cutting (549 250 mE, 5 440 400 mN)

Kaolin was reported in railway cuttings at Tonganah in 1922 and a five acre reward lease was granted to the finder (C. Lutwyche). It was also observed in alluvial tin workings on the lower slopes of Mt Stronach and in cuttings on the Tasman Highway. A proline auger survey was conducted through the northeast of Tasmania in 1966 and several areas of leached, decomposed granite of potential filler clay quality were noted. This information was supplied to the Associated Pulp and Paper Manufacturers (APPM) geologist D. Dickinson, which led to a systematic drilling program in the area and ultimately to the Tonganah operation.

The mine produced about 765 kt of refined kaolin product from about 4.5 Mt of clay mined between about 1980 and 1998. The inferred resource in 1977 was 7.25 Mt (Dickinson, 1977), so nearly three million tonnes of resource may remain. The mine was closed in 1998, as the parent company changed from using kaolin as a filler in its paper production to calcium carbonate. The site has now been rehabilitated.

The Tonganah product

The plant produced a kaolinite filler clay. The particle shape and grain size would appear to be the principal causes of the Tasmanian kaolin being unsatisfactory as coating clay (see mineralogy, below). The 2 μm fraction constitutes only 30% of the product and electron micrographs show that the kaolin crystals occur as stacks measuring in excess of 10 μm rather than as the more desirable single pseudo-hexagonal crystals.

The production of coating clay in Tasmania may be possible by specialised treatment to disaggregate these stacks but it is unlikely that it would compete with the discovery of a higher quality naturally occurring kaolin. This latter possibility cannot be definitely excluded until all potential deposits have been examined.

Mineralogy of Tonganah clay

One sample of the filler clay (G400344) analysed by X-ray diffraction indicated a composition of 92 wt.% kaolinite, 5% mica, 2% quartz and 1% (orthoclase?) feldspar.

No halloysite was detected in the filler clay, but X-ray diffraction analysis of the clay fraction of a sample from the quarry floor indicated about 50% halloysite, 45% kaolinite and 5% illite.

A small sample of the filler clay was examined by Scanning Electron Microscopy. The kaolinite varies from very poorly crystalline through poorly crystalline to moderately crystalline, ragged and vermicular. Small amounts of tubular halloysite are locally common, and appear to be largely a surface alteration on kaolinite. Although the processing of the clay sample probably has caused some distortion and degradation of the kaolinite crystals, the preservation of these halloysite tubes overgrowing kaolinite suggests that such degradation was certainly not pervasive and was probably not highly significant overall. The ragged kaolinite flakes are also indicative of poor crystal growth rather than crystal damage.

Geology and genesis

(from N.C. Higgins and M. Solomon, 1986)

Mapping of the Tonganah deposits and the surrounding granites by Robinson (1982) and company data indicate that the kaolinisation occupies an elongate zone of about 18 km extent within coarse to medium-grained biotite-granite of the Scottsdale Batholith (Kr-Ar biotite 384 3 Ma; unpublished data). The Mt Stronach alkali feldspar granite (minor Mo, Cu, Ag) appears to underlie part of the deposit (Brown *et al*., 1977). Tertiary sedimentary rocks unconformably overlie the kaolinised granite and consist of epiclastic sandstone and siltstone and cobble units. Several lateritic horizons occur within the Tertiary rocks. Drilling data in the area of the clay mine indicate that the kaolinisation has a funnel shape with a maximum depth of 70 m below the present day surface.

Drill core from the Tonganah mine shows a transition upward from sericitised coarse biotite granite through a smectite zone to kaolinised granite over several metres. K-feldspar and sericitised plagioclase are altered to kaolinite and halloysite; chloritised biotite to smectite and kaolinite; while quartz remains unaltered. The clay beneath the Tertiary sedimentary rocks is white but is generally discoloured (red-brown) below twenty metres. This sequence is typical of a weathering profile through a granite body although the clay mineralogy is itself not unique to weathering situations. Tertiary weathering is postulated to have caused the formation of this and other clay deposits in Tasmania (van Moort, 1978).

The K-Ar biotite ages of unaltered granite samples average 384 3 Ma. K-Ar dating of clay size (-0.125 + 0.90 mm) fractions (containing K-feldspar and mica contaminants) of three kaolinised granite samples from Tonganah indicate that, with higher kaolin contents, the K-Ar ages are significantly younger (c.360 Ma). Rb-Sr data suggest that the mica contaminant passed through its blocking temperature around 365 Ma, indicating that there may have been a prolonged period of cooling and hydrothermal circulation within the granite after the initial granite emplacement.

Preliminary isotope data from Tonganah indicate significant variation in the D within the Tonganah deposit. Near-surface samples have ^{18}O values between 19.4 and 20.5‰ and D between -60 and -75‰, and fall on the kaolinite weathering line. The Pioneer and South Mount Cameron redeposited clays have D values greater than -75‰. The initial interpretation of these data is that the kaolinite with heavier D's represent hypogene clays formed at temperatures greater than 30°C. The isotopically lighter D values could have formed by equilibration with recent surface fluids (associated with recent weathering).

The preliminary isotope data, and the association of clay deposits with chemically specialised plutons (Sn, W, F, Li and high U, Th, K) in northeast Tasmania suggest that these alkali-feldspar granite plutons may have been the locus of prolonged hydrothermal convective circulation since granite emplacement, which is similar to the granites of southwest England (Durrance *et al*., 1982). The limited geochemical data available suggest that they contain sufficient radioelements to have produced a long-lived heat source capable of driving a convective system, although alternative heat sources such as the emplacement of the Tasmanian Jurassic dolerites (180 Ma) may have provoked rejuvenation of hot groundwater recirculation. In the latter case the alkali-feldspar granite plutons would have provided ideal sites for subsequent kaolinisation, in preference to other granites, because of their low-iron nature and enhanced permeability resulting from their earlier metasomatic history. How much hydrothermal circulation models can produce the flat-lying geometry, found at the Tonganah clay mine, remains unsolved.

Field observation, road cutting

The cuttings on both sides of the road expose some kaolin-rich zones in clayey gravel and possibly decomposed granite. Most of this material is a yellowish to off-white sandy clay, with angular granite-derived quartz, but there are some sizable patches of good pure white kaolinite. Some rounded quartz pebbles suggest that the material has been transported in part, possibly by fluvial activity, and the relationship to the main kaolinite deposits is uncertain. The main clay deposit just to the south has been rehabilitated.

Stop 3.1

Contact between granodiorite and Mathinna Supergroup, Bridport (532 700 mE, 5 462 000 mN)

An excellent coastal exposure, showing the contact between granodiorite (Scottsdale Batholith) and metasedimentary rocks of the Mathinna Supergroup. Migmatite-like structures in the metasedimentary rocks can be viewed at several places, and include quartzo-feldspathic veins and pods, replacing sediment, with some remnant dark, carbonaceous layering. The metasedimentary rocks appear highly brecciated, sheared and partially replaced by granitic material. The intrusion was presumably relatively dry and volatile-poor, as shown by a relative lack of micaceous minerals and tourmaline. Contact metamorphic minerals are lacking in hand specimen, possibly due to the psammitic nature of the Mathinna Supergroup. Mafic xenoliths, appearing to grade into diorite-like bodies, are locally abundant. These are large, usually tabular with some lamination, and may grade into the wall rocks.

Stop 3.2

Golconda Goldfield

By Tim Callaghan (Leader: Robert Reid, TasGold Exploration Manager, Tasmania)

Location and tenure

The Golconda goldfield is located in northeast Tasmania, about 30 km from Launceston (fig. 19). Two Exploration Licences (EL2/92 and EL41/02), held by TasGold Ltd, cover this area.

Topography

The maximum relief of the area is 400 metres. The main Lisle goldfield occupies a basin-like depression with steep ridges ringing it on all sides, except to the north where Lisle Creek passes through a gorge. The Lone Star, Golconda and Panama goldfields occupy similar but smaller depressions. The steep slopes are generally covered by talus deposits which obscure the bedrock geology.

Previous exploration

The Denison and Golconda alluvial fields were discovered in 1872 and the reefs of Denison and Golconda were first opened up in 1876 and 1877 respectively (Coroneos, 1993). The Lisle alluvial field was discovered by Charles Bessell and company in 1878 following their discovery of the Tobacco Creek goldfield in 1877 (Dickens, 1991).

Most production occurred between 1878 and 1909 but was mainly sporadic throughout that period

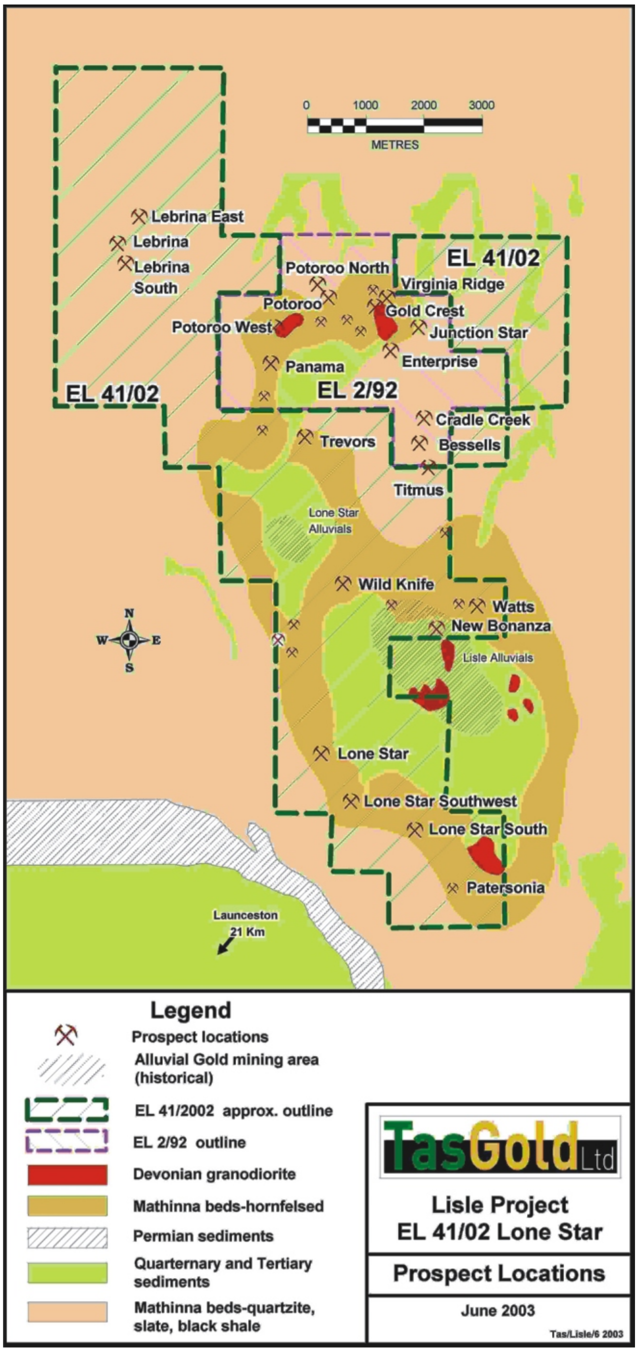


Figure 19

Prospect locations, Lisle–Golconda goldfield.

(Coroneos, 1993). The area officially produced 2.7 t of gold by 1925, mostly from the Lisle Valley alluvial sediments, although records are poor and it is rumoured that a large proportion of the production was taken directly to the Victorian Mint. Twelvetrees (1909) estimated the production to 1909 to be 250,000 ounces. Minor alluvial mining has continued until recent years (Bottrill, 1991).

Hard-rock mining in the Golconda and Panama goldfields continued periodically until the 1920s. Production records are poor but head grades are generally reported to be in the 8–15 g/t range with production mainly from small quartz veins hosted in granitoids and Mathinna Supergroup rocks (Bottrill, 1994).

Comalco Limited pegged EL25/76, covering the Lisle, Golconda and Denison goldfields, in 1976 and undertook a brief review of the area including a pan-con survey, geological mapping and bedrock sampling in the 1970s. Their target was a 10 Mt open-pittable, stratabound Au deposit in altered sandstone of the Mathinna Supergroup (Askins, 1977). The area was considered unprospective for their target after failing to find appreciable gold from their preliminary work. The EL was relinquished in 1977.

CRA Exploration (CRAE) carried out stream silt sampling of the EL area as part of later exploration of another license (EL53/80). This survey showed anomalous arsenic geochemical values in the southern part of the Lisle area (Broadbent, 1982). Some potential for disseminated gold in the metamorphic aureole was considered but no anomalies were followed up.

BP Minerals Australia and the Seltrust Mining Corporation P/L carried out a program of geological mapping, rock chip and stream silt sampling, aeromagnetic geophysical surveying and open-hole percussion drilling between 1983 and 1986. These companies were targeting a bulk tonnage, low-grade gold deposit hosted within the intrusive bodies.

The aeromagnetic survey results delineated the magnetic expression of the Lisle granitoid and also defined a zone of low magnetic intensity concentrically disposed around the granitoids. Small discrete magnetic highs were scattered throughout this zone (Storer, 1985).

Twenty-nine open-hole percussion holes targeted on magnetic and geological targets were completed in 1984. A total of 1037 m of drilling averaging 30–40 m in depth at seven localities was completed. The holes often collapsed, terminating in clay derived from granitoid, although some holes intersected both Mathinna Supergroup rocks and granitoids. Low order geochemical gold analyses were recorded in some places. It was concluded that the weakly altered granodiorites were the probable source of the Lisle alluvial sediments but the grades of the host rock were too low to be of economic interest (Storer, 1984).

Argyle Minerals NL carried out an aerial photograph interpretation between 1986 and 1988 (Cromer, 1987). This was followed up by limited rock chip sampling as well a bulk sampling of the alluvial sediments at the Denison River goldfield outside the EL area. The results indicated limited potential in this area.

Billiton Australia completed a number of programs between 1990 and 1991. These included:

- A regional BLEG stream sediment geochemical survey sampled 26 sites;
- A comprehensive BLEG stream sediment geochemical survey sampled 214 sites. Eleven anomalous sites were re-sampled by duplicate sampling upstream of the original site.
- A composite BLEG soil geochemical sampling program (264 samples) was undertaken over the ridges surrounding the Lisle valley. Three anomalous areas were re-sampled (28 samples) in more detail.

This work outlined two main exploration target areas, the principal one to the north of the Lisle basin with a subsidiary area to the south and west. The anomalies were not followed up in any detail.

MacMin NL completed a number of programs between 1993 and 2001 including:

- Reconnaissance soil geochemical sampling in 1994 across targets delineated from a review of existing data (MacDonald, 1994). This resulted in over 50 anomalous areas delineated by more than 2500 geochemical samples.
- Grid based B-horizon soil geochemical sampling in 1995 across five grids, follow-up power auger sampling, rock-chip geochemical sampling from selected adits and shafts (Hall, 1995).
- Reconnaissance drilling of four diamond core holes (195.3 m) at the old Enterprise and Gold Crest mines in late 1995 (Duncan, 1996).
- Reconnaissance drilling of four reverse circulation percussion (RC) holes (359 m) at the Enterprise Prospect in 1996 (Duncan, 1996).
- Further soil and auger geochemical sampling, wacker drill and excavator trench sampling in 1997 and 1998 in the Panama Valley, Enterprise Ridge and Tobacco Creek areas, as well as excavator trench sampling of other geochemical anomalies (Hall, 1995).

TasEx Resources Limited completed an RC drilling program on the EL in 2002 (McNeil, 2002) including:

- Fifteen RC holes for 571.5 m at Potoroo. Significant gold associated with disseminated sulphide mineralisation and quartz veining hosted in altered granitoids was identified.
- Five RC holes for 247 m and one diamond tail for 122.5 m at Enterprise. Many holes intersected old workings or the quartz reef, with best intersections of 2 m @ 2.9 g/t Au and 0.4 m @ 14.4 g/t Au from hole E005.

TasGold has recently completed RC, diamond and trenching programs on the Enterprise, Potoroo, Junction Star, Panama and Kelly's prospects.

- Eleven RC holes and six diamond-drill holes at Enterprise identified the Enterprise vein over a strike length in excess of 400 m and delineated a new mineralised vein (West Vein).
- Six RC holes at Potoroo extended the known mineralisation. Disseminated and vein mineralisation on altered granodiorite with grades such as 26 m @ 0.6 g/t Au in hole P018.

Ore deposit models

The majority of northeast Tasmanian gold deposits are typical slate belt style, mesothermal gold deposits similar to the Victorian goldfields. The best known and single largest reef (including Victoria) is the Tasmania Reef at Beaconsfield, which contains >2.91 Mt @ 19.8 g/t Au. The Tasmania Reef consists of a quartz + carbonate + sulphide-filled fracture that is transgressive to the host sediments and is fault controlled. The reef varies in width from less than one metre to approximately five metres and has a strike length of 350 to 400 metres. The reef remains open at depth.

Unlike most of the northeast Tasmanian gold deposits, the Lisle–Golconda reef deposits appear to be related to the reduced granodiorites of the Scottsdale Batholith. There is an obvious spatial relationship between late-stage intrusive rocks and gold mineralisation. Gold is hosted in quartz-sulphide veins and disseminations within intrusive rocks, and structurally-controlled veins within the contact aureole. Sulphide minerals include arsenopyrite and pyrite with lesser chalcopyrite, bismuthinite, stibnite and molybdenite. Geochemically the mineralisation has a Au, Ag, Bi and Mo association.

Intrusion-related gold deposits (associated with tungsten-tin deposits) are an under-recognised and economically important class of gold deposits. These deposits include sheeted veins, quartz stockworks and bulk mineable disseminated gold deposits spatially and geochemically associated with reduced intrusive rocks (fig. 20).

Examples of these styles of deposits are known in Alaska, the Czech Republic, Spain, Kazakhstan, Bolivia and Australia. The Kidston (Queensland) and Timbarra (New South Wales) deposits are Australian examples.

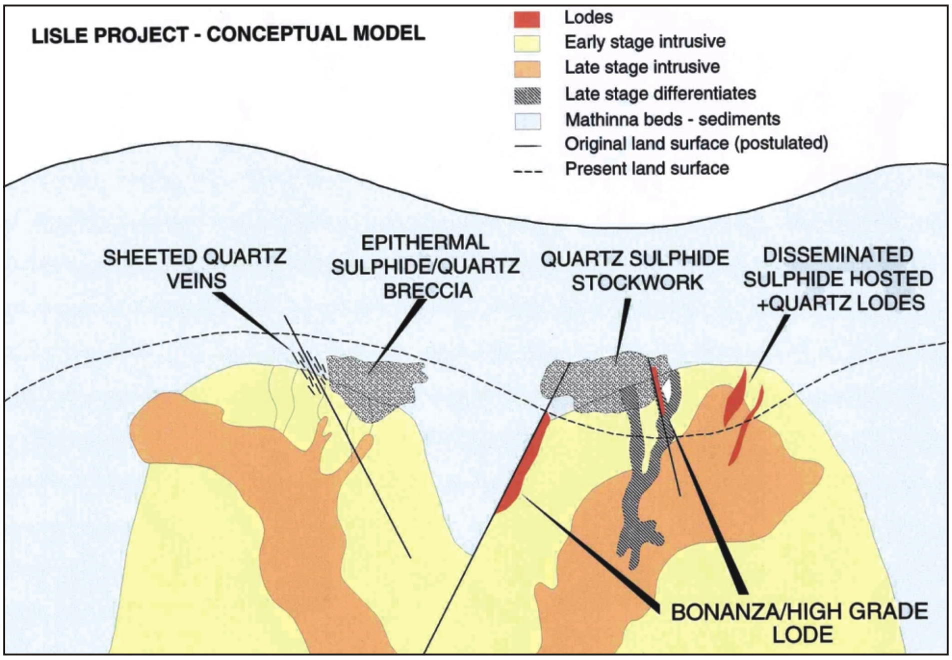


Figure 20

Conceptual model, Lisle Project.

World-class Alaskan deposits of this style include Pogo and Fort Knox. Pogo is reported to host more than nine million tonnes at 17.8 g/t Au for more than 5.0 million ounces contained gold. Mineralisation occurs in three or more tabular, gently-dipping quartz bodies associated with early biotite and later quartz-sericite stockwork and sericite-dolomite alteration. The quartz bodies occur 1.5 km south of a Cretaceous batholith and are hosted primarily in gneiss.

Fort Knox occurs as a structurally-controlled stockwork and shear quartz veins in a granodiorite pluton. It is reported to host 158.3 Mt at 0.83 g/t Au for more than four million ounces contained gold.

Recent reports of gold mineralisation discovered at the nearby Denison goldfield by Anglo Australian Resources in sandstone (Mathinna Supergroup) may also be a model worth considering for the Lisle Project.

Local geology

A good interpretive geological and structural map of EL2/92 has not yet been completed. The local geology is dominated by ridges of hornfelsed Mathinna Supergroup and basins of weathered granodiorite and diorite. Numerous granodioritic and dioritic dykes intrude the Mathinna Supergroup. Valleys and ridge slopes are covered by Quaternary talus and alluvial deposits, obscuring most of the recessive geology.

The Mathinna Supergroup generally consists of a monotonous sequence of graded, quartz-wacke turbidites with lesser siltstone and black shale. Where observed in outcrop these rocks appear to form NNW-trending folds with several fold closures apparent on the lease. A weak NNW-striking slatey cleavage is observed in some outcrops. Further structural mapping and interpretation is required. The Mathinna Supergroup is locally hornfelsed with chlorite after cordierite spotting common within hundreds of metres of contacts with Devonian granitic to dioritic intrusive rocks.

Granitic to dioritic intrusive rocks are generally deeply weathered and rarely crop out. Rare outcrop and core intersections indicate that the intrusive rocks are complex and inhomogeneous, with numerous inclusions of hornfelsed Mathinna Supergroup and dark diorite. Textures vary from equigranular, feldspar-biotite-quartz granodiorites to feldspar-hornblende-biotite porphyritic diorites. Intrusions occur as dykes and small cupolas or porphyritic apophyses, possibly off a larger buried body. The largest known intrusive of this type occurs in the Lisle Valley and measures approximately 4 by 4 kilometres.

Roach (1992) analysed 16 samples of the various granodiorites from Lisle, Golconda, Panama and the western margin of the Scottsdale Batholith, known as the Diddleum Pluton. There is a clear distinction between the rocks of the Scottsdale Batholith and the granodiorite from the Lisle area. In terms of Rb and Sr the Lisle granodiorites are the least fractionated of the Tasmanian Devonian granitoids.

There is a marked variability of the magnetic susceptibility of the granodiorites. This is probably a reflection of varying geochemistry between the complex intrusive rocks but may also represent areas of magnetite destruction associated with hydrothermal alteration.

Mineralisation and alteration varies between host rocks. Within the Mathinna Supergroup it occurs as thin (0.1 to 1.5 m) quartz veins with strike lengths of up to several hundred metres. Veins appear to be hosted in late brittle faults. Vein attitudes vary between prospects but are generally steeply dipping. Some reported stratabound mineralisation of silicified auriferous sandstone beds has been reported (Reid, 1926; Fulton, 2001) although these have not yet been observed by the author.

Mineralisation and alteration within the intrusive rocks is associated with intense sericite-silica alteration and variable disseminated pyrite and arsenopyrite. Quartz stockworks and sheeted veins are intimately associated with alteration zones within the intrusive rocks. Vein orientations and styles again appear to vary between prospects. Minor ankerite, siderite and sulphide minerals are associated with quartz veining and as pervasive and disseminated selvedge alteration. Sulphide minerals include dominantly pyrite and arsenopyrite, with lesser galena, molybdenite and chalcopyrite.

Enterprise prospect (526 000 mE, 5 441 100 mN)

The Enterprise area hosted the largest hard-rock workings within the Lisle–Golconda field and was also a significant alluvial producer. The prospect contains numerous shafts, pits and adits, many collapsed or subsequently disturbed. The main working is located on an extensive (>400 m) quartz-sulphide reef hosted in an approximately north-striking (350°), moderately west-dipping (40–50°) thrust fault (fig. 21, 22). The reef/fault is hosted in fine to medium-grained, multiphase granodiorite and diorite. Other parallel and spur veins are present, increasing the likelihood of a multi-veined system. A significant example is the West Vein located at the southern end of the Enterprise Prospect and the Gold Crest reef to the northwest.

Mineralisation is variably developed in dilatant zones along thrust faults, with a number of discontinuous veins varying between 0.3 and 1.5 m in width. The host rock is moderately to strongly silicified and sericite-chlorite altered, with minor ferroan carbonate. Strong alteration extends approximately five metres either side of the reef. Moderate stockwork veining with altered granodiorite selvedges extends into the footwall of the reef. Sulphide minerals comprise around 1% of the mineralisation and occur within quartz veins and as disseminations within altered granodiorite. Common sulphide minerals include pyrite and arsenopyrite with minor chalcopyrite,

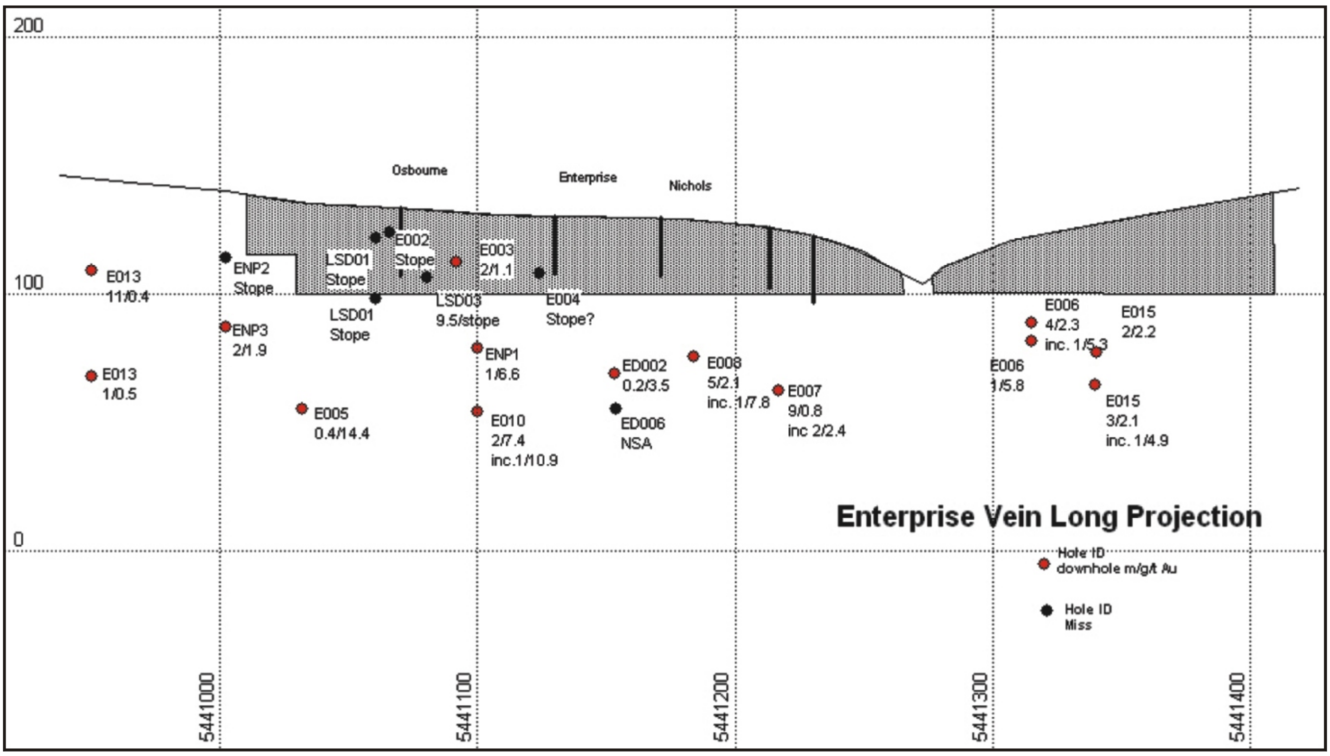


Figure 21

Enterprise vein, long projection.

galena and molybdenite. Gold occurs as fine electrum grains.

Earlier rounds of drilling produced variable results, with many holes intersecting old workings or missing the main structure. A further eleven RC holes and six diamond holes were completed this year for a total of 29 holes at the Enterprise Prospect. Drilling extends over a strike length of 400 m and to a depth of 80 m from surface. The mineralisation remains open down-dip and along strike to the north and south. There is a high potential for further resources to be defined from parallel structures and structural dilatant zones (e.g. West Vein) as well as extending the main Enterprise Reef.

Potoroo prospect (524 800 mE, 5 442 050 mN)

The Potoroo prospect is located at the eastern end of the Panama Valley. The prospect contains small granodiorite intrusions into hornfelsed Mathinna Supergroup rocks. The granodiorite has been deeply weathered and forms topographic lows largely obscured by Quaternary talus.

Mineralisation consists of sheeted quartz-pyrite-arsenopyrite veins and disseminated sulphides in altered granodiorite. The Mathinna Supergroup is apparently less mineralised than the granodiorite. Several trenches and drill holes have been completed at the prospect. Unfortunately no maps from previous rounds of trenching are available so geological control on the prospect is poor. Recent trenching indicates that granodiorite dykes and veining within the granodiorite trend northwest and dip steeply northeast. If this is the case then earlier rounds of drilling were not optimal to the strike of the mineralisation.

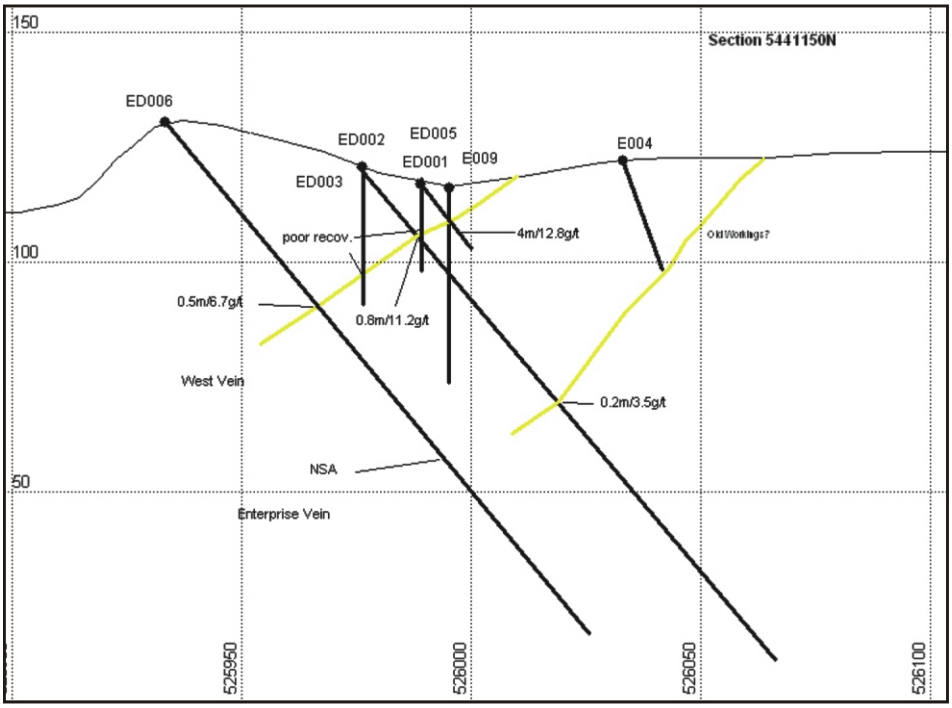


Figure 22

Section 5441150N, Enterprise prospect.

Consistent gold mineralisation is indicated by channel sampling in many trenches, with significant intersections of greater than 0.1 ppm Au (Table 2).

RC drilling of the prospect has defined significant low-grade intersections within a limited area. The mineralised zone remains open to the south and west. Results suggest that the prospect has good potential to host a bulk tonnage low-grade deposit (Fort Knox style). The drill spacing needs to be increased to scope the size potential of the deposit.

Table 2

Gold and silver results from costeans at Potoroo prospect, northeast Tasmania

Costean No.	*Length (m)*	*Au (g/t)*	*As (ppm)*
TP1	64	0.6	1443
TP2	50	0.3	8
TP3	25	0.1	13
TP4	74	0.2	1134
TP5	98	0.4	3485
TP7	32	0.1	269
TP9	14	0.1	807
TP9	8	0.1	275

Lisle goldfield

Introduction

Several goldfields were discovered in this area in about 1877 and were dominated by the Lisle alluvial field, with a probable gold production of nearly ten tonnes. A small production of gold continues from Lisle.

The workings are spatially closely related to hornblende-biotite-magnetite-sulphide bearing granodioritic intrusions in the Mathinna Group. The metamorphic aureoles are commonly sharply defined, varying from 800 m to about 5 km in width, depending upon the dip of the contact (McClenaghan *et al.*, 1982). Within these aureoles the sedimentary rocks are commonly spotty and/or hornfelsed, and may contain biotite and cordierite, as well as recrystallised quartz, muscovite and chlorite. Some hornfels contains up to 80% iron cordierite (sekaninaite). Quartz and greisen veins and granitic dykes locally occur in the aureole near the contact.

Lisle

Workings at Lisle included alluvium and eluvium in a basin-shaped depression, possibly representing an old lake bed of Tertiary age (Reid, 1926; Marshall, 1969). There were numerous patchy gold-rich horizons in the possible lacustrine sediments, and in carbonaceous horizons underlying talus, which produced relatively pure, free, angular (crystalline?) gold (Noldart *in* Marshall, 1969). This type of gold suggested a secondary origin (i.e. *in situ* reprecipitation of dissolved gold from groundwater; Reid, 1926; Bottrill, 1986). Some gold grains are highly porous and/or colloform, while some have silver-rich cores and silver-depleted rims (Bottrill, 1991; Roach, 1991), confirming that some gold is detrital and some reprecipitated.

Auriferous quartz was relatively rare, both in alluvium and bedrock. Twelvetrees (1909) found evidence for gold originating in the contact metamorphosed sandstone of the Mathinna Supergroup surrounding the basin, near the contact with Devonian granitoid intrusive rocks. Inclusions of mica, rutile and magnetite in the gold grains suggest that the gold was more likely to have been disseminated in the hornfels or granitoids than in quartz veins, while rare gold-limonite composites in placers suggest that gold-bearing pyrite may have originally been present (Bottrill, 1986). Some gold was found in small quartz veins in the granitic rock underlying the alluvial sediments (Thureau, 1882; Montgomery, 1894). Drilling by the Department of Mines revealed very minor quartz-carbonate-pyrite alteration zones in the magnetite-pyrite bearing granodiorite, with trace gold (to 0.05 g/t Au; Bottrill, 1996).

Stop 3.3

Pipers Brook Winery

Pipers Brook Vineyard was founded in 1974 and is one of the leading producers of cool-climate wines in Australia. Architecturally interesting, the winery at Pipers Brook is one of the most advanced in Australia and is functionally efficient for fine wine production. It has been recently expanded and now boasts cellars for barrel storage, gravity feed wine transfer, modern bottling facilities and sparkling wine cellars.

Stop 3.4 (time permitting)

Mineralisation in the Moina area: Shepherd and Murphy mine (423 277 mE, 5 406 370 mN)

(summarised from Collins, 1979)

The Moina area has long been considered to be one of the most prospective areas in Tasmania, and has attracted many exploration companies and prospectors during the last 120 years.

The area is underlain by a sequence of Ordovician shallow marine sedimentary rocks, correlates of the Denison Group and Gordon Limestone. These rocks overlie Cambrian volcanic rocks, correlates of the Mt Read Volcanics, the source and host rocks to a range of significant precious and base metal deposits in western Tasmania. The rocks have been folded, and intruded by Late Devonian to Early Carboniferous Dolcoath Granite over a large area, although the granite is poorly exposed.

Sediments of Tertiary age are interbedded with the Tertiary basalt flows which cover most of the area. Quaternary alluvial deposits can also be widespread in some areas.

The intrusion of the Dolcoath Granite developed a contact aureole characterised by a wide range of alteration types and associated mineralisation, including fluorite-magnetite skarn, base metal-gold skarn (Hugo Skarn) and tungsten-tin-bismuth and molybdenum veins (Shepherd and Murphy mine). The mineralisation in the area is spatially and genetically related to the intrusion of the Dolcoath Granite and resulted from ore-bearing, fluorine-rich hydrothermal fluids dispersing through intensely fractured, volcanic and highly siliceous rocks.

Shepherd and Murphy mine

There have been approximately thirty mines and prospects in the Moina and Round Mount districts (fig. 23), but only three mines have achieved significant production. These are the Shepherd and Murphy tin-tungsten-bismuth mine at Moina, the Round Hill silver-lead mine at Round Mount, and the Bell Mount alluvial goldfield. The Shepherd and Murphy mine is situated on the south bank of Bismuth Creek at Moina, about 2.5 km west of the nearest Dolcoath Granite outcrop.

The lode deposits of the Shepherd and Murphy mine were discovered in 1893 by Thomas Shepherd and Thomas Murphy. The mine was basically developed by a series of adits and shafts on six main east-west trending, near vertical, mineralised quartz veins. The mine contributed the greatest part of the total production of tin, tungsten and bismuth from the Moina and Round Mount districts. During periods of intermittent production between 1893 and 1957, an estimated 525 t Sn, 255 t WO_3 and 71 t Bi were recovered from the underground and surface workings at this mine. The mine has been abandoned since 1957, but sporadic interest has been shown in the deposit since then, particularly the surrounding skarns.

The lodes of the Shepherd and Murphy mine are essentially quartz fillings of tension fractures, occurring as a series of almost vertical (dip about 85°S), east-trending quartz veins up to 395 m in length and varying up to 900 mm thick, but generally 350–500 mm thick. Six parallel veins (fig. 24), numbered from south to north, occur within a width of 275 metres. Of these, veins 2, 4, 5 and 6 have been developed and worked. An additional vein, the northwest branch lode, trends northwest and intersects the No. 6 lode west of the main shaft. The veins penetrate the dense quartzite and overlying skarn.

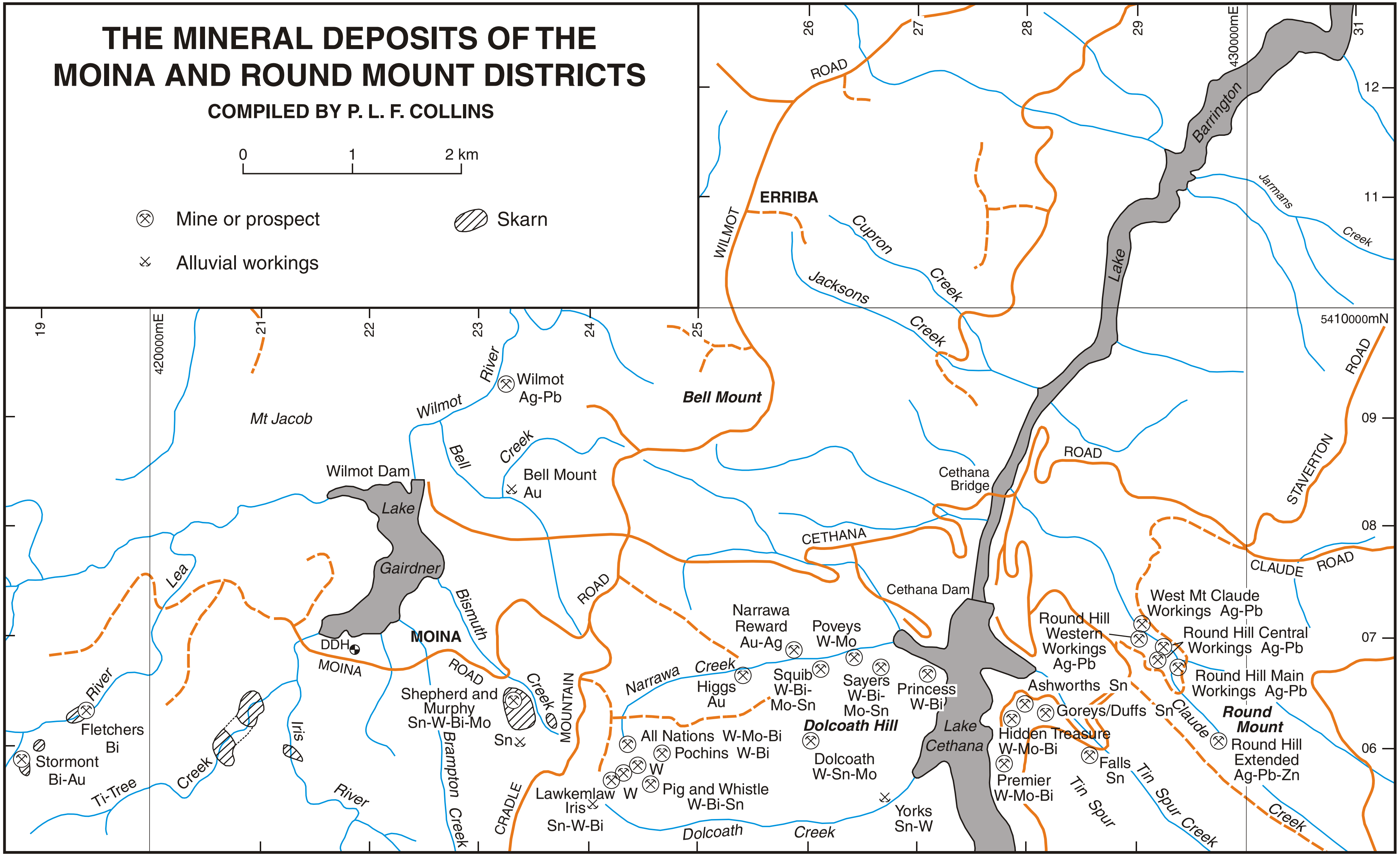


Figure 23. *Moina mineral field, showing locations of deposits and different types mineralisation (from Collins, 1979).*

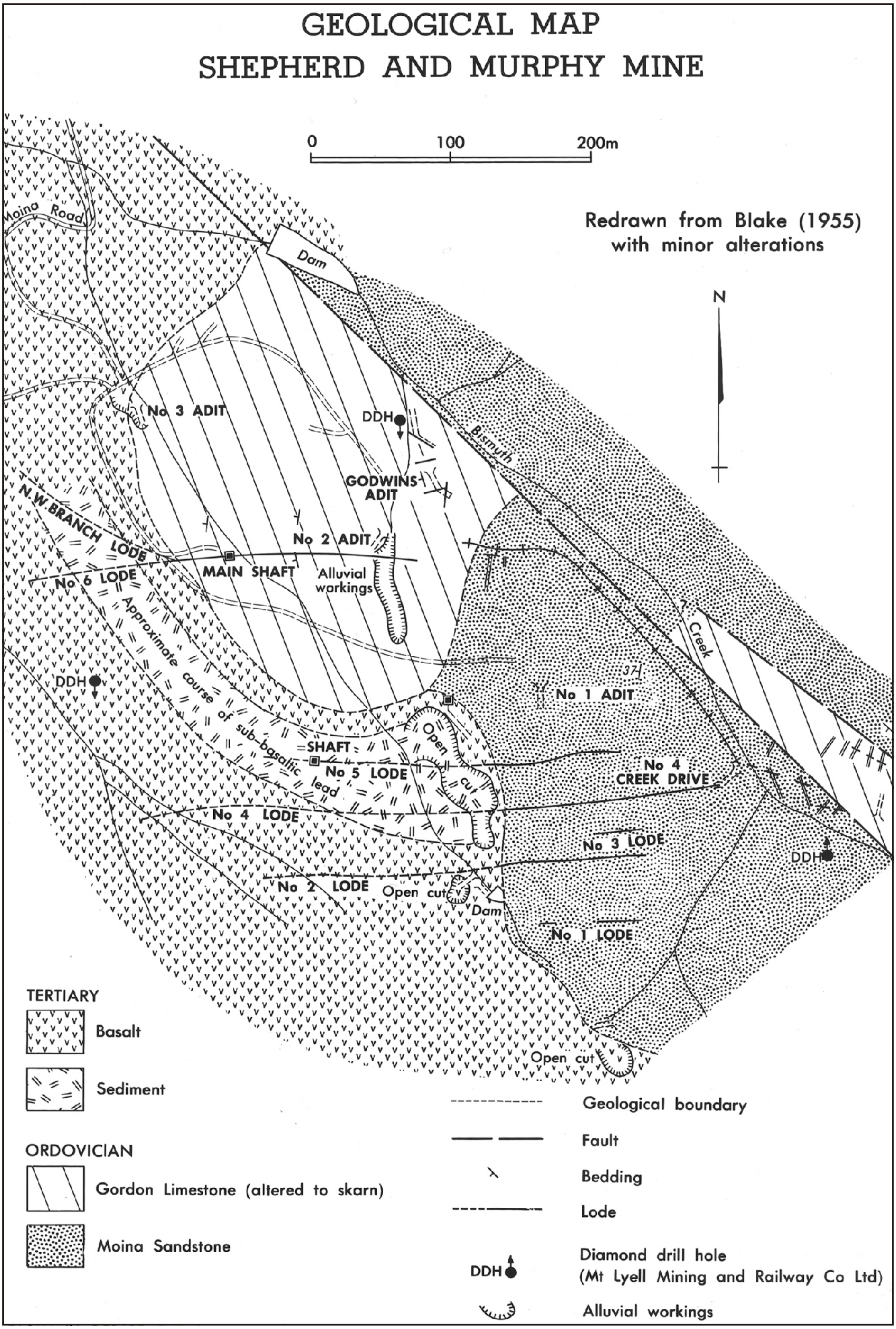


Figure 24

Simplified map of Shepherd and Murphy mine area, showing the location of the six main mineralised lodes.

The veins consist predominantly of quartz with fluorite, topaz, phlogopite, muscovite, chlorite, laumontite, calcite, siderite and beryl as accessory gangue minerals. The metallic minerals of economic importance comprise cassiterite, wolframite and bismuthinite. Subordinate metallic minerals include pyrite, marcasite, magnetite, hematite, pyrrhotite, arsenopyrite, molybdenite, chalcopyrite, sphalerite, galena and scheelite (as an alteration product after wolframite). Native bismuth and bismuthinite also occur near the surface (Blake, 1955; Williams, 1958). A detailed description of the mineralogy of the veins is given by Williams (1958). Wall-rock alteration is a minor feature only noted occasionally (Blake, 1955).

The depth and lateral extensions of the veins have been tested by different exploration companies including Comalco-Shell and the Mt Lyell Mining and Railway Co., but the results were discouraging. The veins have been cut to the east by a large transverse fault (Bismuth Creek Fault), but recent drilling by Goldstream-Titan on the eastern side of this fault indicated an abundance of narrow, mineralised veins developed both in skarn and sandstone.

In addition to the main east-west lodes there are two other lode types: 'banded lodes' and northwest branch lodes. The 'banded lodes' consist mainly of quartz with some chalcopyrite and fluorite and are frequently banded, consisting of laminations of fine siltstone and quartz varying from 3 mm to 12 mm thick (Burns, 1958). The banding is interpreted by Burns to be silicification of fracture cleavage in shale bands interbedded with the quartzite. The 'banded lodes' are considered to be contemporaneous with the east-west lodes (Burns, 1958). The northwest branch lodes are considered to be mineralised transverse faults which generally cause negligible displacement of the east-west veins.

Field observations

Outcrops and dumps showing some complex skarns, containing mostly andradite, fluorite, magnetite, diopside, minor sulphide minerals (pyrrhotite and chalcopyrite) and hastingsitic amphiboles, with pink feldspar and white quartz veins, can be seen at this site. Some of the quartz veins are vuggy and contain variable amounts of coarse-grained muscovite, fluorite, topaz, sulphides (arsenopyrite, bismuthinite, molybdenite and pyrite), wolframite and cassiterite.

Some similar small veins occur in Moina Sandstone in the Dolcoath Hill quarry a few kilometres to the east, along the same mineralising structure. These veins are of two main types: quartz-wolframite and quartz-topaz-cassiterite-kaolinite.

Day 4

Stop 4.1

Mt Bischoff tin deposit (376 600 mE, 5 412 200 mN)

(Adapted from Halley, in Turner and Taheri, 1990)

The Mt Bischoff tin deposit was the first major mineral resource developed in western Tasmania. Mining commenced in 1872 and quickly reached full production, with the recovery grade for the first twenty years or so being almost 3% Sn.

Production and reserves

Up until 1921 the Mt Bischoff Tin Mining Company treated 4.598 million tonnes of open-cut and underground ore for an average recovery grade of 1.17% Sn (Reid, 1923). From 1921 to the end of major hard-rock mining in 1947 the company, and then the Commonwealth and Tasmanian governments, treated about 0.8 Mt of lower grade ore from the same sources. The Wheal Giblin (fissure) Lode and associated workings immediately west of the main deposit produced about 0.15 Mt of underground ore, mostly at a recovery grade of 0.94% Sn, in the period to 1918 (Reid, 1923) and about 0.09 Mt thereafter. About 0.2 million cubic metres of North Valley alluvium, mainly derived from the Waratah River flats north of Mt Bischoff, were treated by the Mt Bischoff Tin Mining Company in the period 1928–1942, whilst another 0.556 million cubic metres were treated for 570.167 tonnes of tin in concentrates by Ringarooma Mining Pty Ltd in the period 1972–1976. Altogether about 62 000 t of metallic tin in concentrates were produced from the Mt Bischoff tin field to 1978.

Following a period of detailed prospecting a 4 km^2 area surrounding the old mines was granted as a Retention Licence in 1988 to a joint venture syndicate comprising Comstaff Pty Ltd, Preussag and Metals Exploration Ltd. Geological mapping, trenching, auger and rock-chip sampling, aeromagnetic and ground magnetic studies were followed by diamond, percussion and reverse circulation drilling on 20 20 m and 40 40 m grids to establish a basis for calculation of ore reserves. Reserves were classified into two ore types, namely dolomite-sulphide lode ore (0.3% Sn cut-off) and quartz porphyry ore (0.2% Sn cut-off). Combined proved, probable and possible reserves of dolomite-sulphide ore are about 1.3 million tonnes at 1.00% Sn whilst probable and possible reserves of porphyry ore total about 3.40 Mt at 0.47% Sn.

Geology and genesis of the Mt Bischoff tin deposit

Setting

The mineralisation occurs within an inlier containing a succession of late Precambrian quartzite, shale and minor dolomite which is correlated with the Oonah Formation (Burnie Formation). These rocks were intruded by a number of quartz-feldspar porphyry dykes of Devonian age. Hydrothermal 'pebble dykes' were formed at the same time as the intrusion of the dykes. The mineralisation is spatially related to the porphyry dykes and occurs as replacement of a dolomite horizon within the sedimentary rocks, as greisenisation of the dykes, and as veins and fracture linings. The mineralised dolomite has been dissected by erosion, with a few isolated bodies preserved in synclines (fig. 25).

The porphyry dykes have been dated at 349 4 Ma (Rb-Sr). This is a similar age to the Meredith Granite (353 7 Ma) which crops out about eight kilometres to the southwest. A gravity anomaly suggests that a buried ridge of granite extends from the Meredith Granite and runs beneath the Mount Bischoff deposit.

Altered porphyry dykes

All of the dyke rocks have undergone at least one stage of alteration. The least altered rocks were affected by an early potassic alteration event. They are characterised by quartz and orthoclase phenocrysts and micrographic intergrowths of quartz and orthoclase. They contain up to 8% K_2O, but only traces of Na and Ca. The potassically altered rocks are variably overprinted by later sericitisation.

The early potassic alteration in the dykes and the overlapping temporal relationships between the dykes and hydrothermal breccias suggest that the earliest event in the history of the Mt Bischoff deposit was the intermittent explosive release of volatiles and synchronous intrusion of dyke material from an underlying water-saturated granitic melt. The tin mineralisation post-dates and overprints this early magmatic-hydrothermal event.

The greisenised dyke rocks are characterised by a hard, white, fine-grained groundmass of quartz and topaz with orthoclase phenocrysts pseudomorphed by a variety of minerals. At the margins of the greisen, the pseudomorphs are composed of siderite. Through most of the greisen, the pseudomorphs are composed of intergrowths of pyrrhotite, quartz, topaz, pyrite, fluorite and cassiterite. With increasing intensity of alteration, the rocks become progressively silicified. Quartz phenocrysts develop optically continuous overgrowths of quartz and the relict porphyritic texture is overprinted by coarsening of the groundmass. Alteration zonation along and across the dykes suggests that the dykes were major fluid conduits.

Dolomite replacement mineralisation

The dolomite replacement mineralisation was formed in two separate and distinct events. The mineralogy of the replacement rocks is usually dominated by

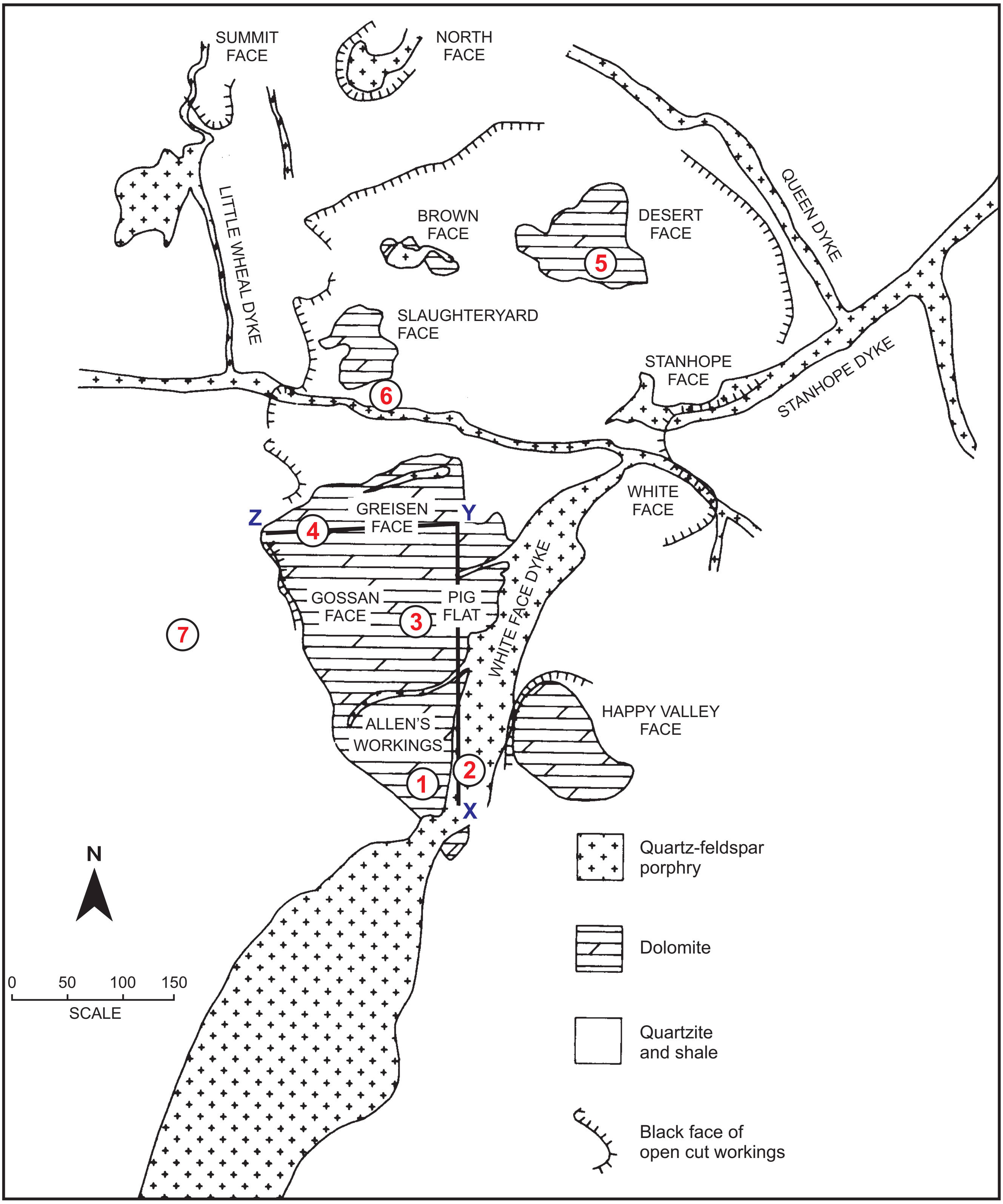


Figure 25

Geology of the main mineralised area at Mt Bischoff mine. The lines X-Y-Z correspond to the block section in Figure 26. Bold numbers are the localities for inspection.

pyrrhotite plus carbonate or one of a number of silicate minerals. The rocks can be divided into a number of assemblages characterised by the dominant gangue mineral (fig. 26). These assemblages are:

Stage 1	(i)	chondrodite
	(ii)	serpentine
Stage 2	(iii)	carbonate
	(iv)	phlogopite
	(v)	talc
	(vi)	quartz

The stage 1 mineralisation was formed by the reaction of dolomite with a hot (400–460°C), saline (30 to 36 wt.% NaCl) fluid. A zonation from serpentine to chondrodite to magnesite to unreplaced dolomite developed in response to decreasing silica activity in the fluid as it reacted with dolomite and precipitated magnesium silicates. The stage 1 replacement assemblages are barren, commonly containing less than 100 ppm Sn.

The stage 2 mineralisation is very similar to the Renison Bell deposit in terms of its mineralogy and chemistry. It formed at 300–360°C from a fluid with about 10–12 wt.% NaCl. The greisenisation of the dykes was produced by the same hydrothermal event. The stage 2 replacement assemblages are zoned from quartz to talc to carbonate (averaging 65% $MgCO_3$, 35% $FeCO_3$) to unreplaced dolomite. The phlogopite assemblage is relatively minor and, where present, occurs towards the outer margin of the talc zone. The phlogopite-bearing rocks commonly exhibit fine, contorted banding and have been termed wrigglite.

The zonation again reflects decreasing silica activity in the fluid as wall rock reaction proceeded. The boundaries between the assemblages are very sharp chemical reaction fronts which were continuously pushed outwards as more fluids passed through the system such that, for example, the quartz assemblage was replacing the talc assemblage in the inner part of the system while at the same time talc was replacing the carbonate assemblage in the outer part of the system.

The highest tin grades occur within the quartz assemblage. Relatively high grades also occur within parts of the talc assemblage. Cassiterite shows a close association with the silicate gangue minerals rather than with pyrrhotite. Paragenetic relationships indicate that most of the cassiterite was precipitated in the quartz and talc assemblages, after carbonates had been replaced. The most strongly altered dolomite replacement rocks, the quartz assemblage, have the same mineralogy as the most strongly greisenised dyke rocks, that is quartz + pyrrhotite + topaz + fluorite. This shows the influence of the fluid composition rather than initial rock composition in determining the ultimate alteration mineralogy.

Fissure lodes

Mineralised fissure-filling veins, which cross cut the sedimentary rocks and porphyry dykes, formed at a

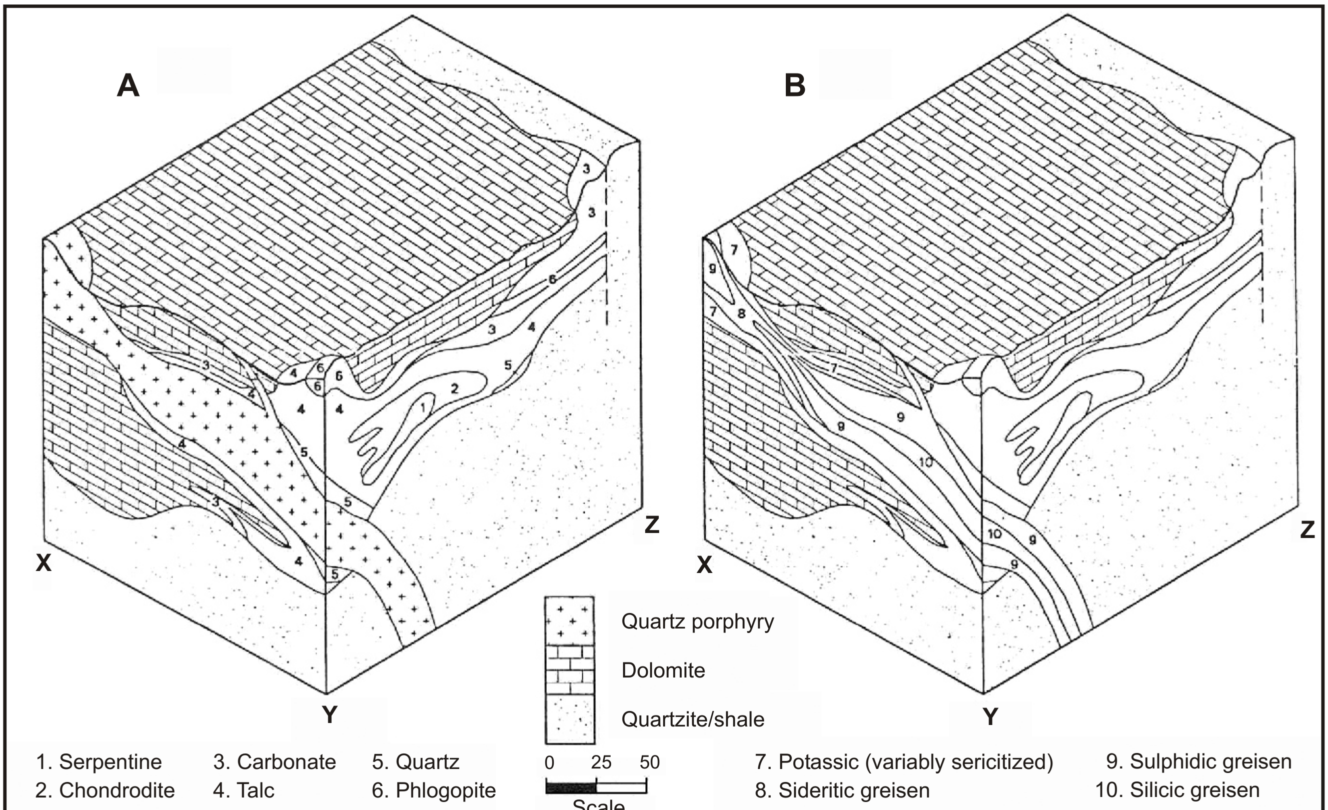


Figure 26

Block section of Greisen Face looking southwest, showing: (a) the distribution of dolomite replacement assemblages, and (b) the distribution of alteration in the White Face Dyke. Section lines correspond to X-Y-Z in Figure 25.

slightly lower temperature than the stage 2 mineralisation. They contain a complex assemblage of pyrite, base metal sulphides, As-Sb-Bi sulphosalts, quartz, topaz, fluorite and cassiterite.

Tin precipitation

Fluid inclusions from the stage 2 mineralisation in the altered dolomite indicate that all the assemblages from this stage were formed from the same fluid type with consistent temperatures and composition. The fluid contained significant amounts of CO_2 and CH_4 (1.5 m CO_2 and 0.3 m CH_4). Fluid inclusions also provide evidence of boiling, particularly within the quartz assemblage. The chemical changes resulting from boiling, particularly increasing fO_2 and pH, would efficiently precipitate tin from the fluid. Boiling has probably been responsible for the precipitation of most of the cassiterite in this deposit. This mechanism is consistent with the distribution and paragenetic relationships of cassiterite.

Field observations (for locations see Figure 25)

1. Unaltered dolomite, Allens workings: High oxygen isotope values are consistent with an evaporitic deposit and environment. Some minor siderite veins and pods are evident.
2. White Face Dyke: Abundant topaz, sulphide, tourmaline, fluorite and other greisen alteration of Devonian quartz porphyry.
3. Pig Flat – Wrigglite: Quartz-tourmaline- sulphide–rich skarn. Also some late-stage siderite-sulphide veining.
4. Greisen Face: Chondrodite/talc/pyrrhotite-carbonate skarn alteration.
5. Desert Face: Chondrodite-pyrrhotite skarn alteration, contact between magnesian skarn and Burnie Formation (fig. 27).
6. Western Dyke: Greisenised (quartz-topaz-pyrite altered) porphyry with cassiterite (disseminated and in coatings on joint surfaces).
7. Excellent exposure of porphyry dyke and hydrothermal breccia. Quartz/cassiterite veinlets in fractures in Burnie Formation.

Figure 27

The Desert Face, Mt Bischoff, showing variably oxidised tin and sulphide-rich magnesian skarns (mostly red and yellow), in contact with Burnie Formation siltstone and quartzite (pale grey).

Stop 4.2

Kara (scheelite-magnetite) mine (402 625 mE, 5 425 400 mN)

Alwyn Neubacher

Tasmania Mines Limited (Tasmines) operates the Kara mine which is located seven kilometres south of Hampshire and 40 km south of Burnie in northwest Tasmania. Tasmines is a publicly listed company listed on the Australian Stock Exchange. The mine operates within Consolidated Mineral Lease 1371P/M which has an area of 728 ha and which was granted on 1 May 1989 for 21 years. A total of 505 ha of the CML are Crown Land, 142 ha are freehold land owned by Australian Forest Holdings, and 81 ha are freehold land owned by Tasmines.

Tasmines produces magnetite, an iron ore, and scheelite, a tungsten ore, from the Kara mine and is a key supplier of iron-ore products to other Tasmines operations, Renison Limited, and the Cornwall Coal Company. Tasmines exports a value added, specialist iron-ore product (dense medium magnetite) to coal mines in NSW and Queensland, and scheelite to overseas markets. The value of Tasmines' sales in 2002 was over $7.2 million.

Tasmines has 15 permanent employees at the Kara mine and there can also be up to 16 people on site who are employed by various consultants and contractors.

During 2002, 149 140 t of ore were mined and 66 382 t of magnetite were sold. No scheelite was sold because of low prices. Planned production from Kara No. 1 for 2003 to 2005 is 147 000 t of ore per year from which 70 000 t of dense medium magnetite concentrate will be produced.

The Kara ores occur in skarn deposits which have replaced limestone and which occur immediately adjacent to the Housetop Granite. In this sense, they occupy a setting similar to the tin deposits at Renison Bell and Mt Bischoff. The ores consist of more or less massive magnetite (Fe_3O_4) which, in places, contains up to about 1% scheelite ($CaWO_4$).

Mining commenced on the site in 1977 and has continued since. The Kara No. 1 mine is Tasmania's second oldest, continuously operated, metal mine.

The mine is an open cut. Ore is concentrated in a mill which operates principally on gravity and magnetic separation methods. A small sulphide flotation circuit is used to remove sulphide minerals from the scheelite concentrate.

A northern extension to the Kara No. 1 pit has commenced, with the pit to be extended for about 200 m north of its current footprint.

Geology and mineralisation

(Adapted from Khin Zaw and Singoyi, 2000)

The major ore bodies at Kara are hosted by the Ordovician Gordon Limestone at the southern end of the Devonian Housetop Granite, adjacent to the granite, or separated from it by the Ordovician Moina Sandstone.

The Kara deposit formed as a proximal skarn assemblage in carbonate host rocks following the emplacement of the Devonian Housetop Granite and was characterised by early, high-temperature mineral assemblages dominated by anhydrous minerals, and late low-temperature assemblages with abundant mineralisation (scheelite, magnetite) and hydrous minerals.

The geological setting of the Kara skarn deposit area is shown in Figure 28. The oldest rock units are Cambrian and are exposed to the northwest of the deposit area. The Cambrian sequence consists of laminated cherty mudstone interbedded with carbonate rocks (Baillie *et al*., 1986), with contact rocks that are commonly transformed to hornfels, marble, and in places to metamorphic skarn assemblages. This sequence may also contain pyrrhotite (Barrett, 1980). The Ordovician units comprise the Owen Conglomerate, sandstone and argillite (correlative with Moina Sandstone), and limestone and impure limestone (correlative with Gordon Subgroup). The Owen Conglomerate and the Moina Sandstone form the Denison Group of Banks *et al*. (1989). The Owen Conglomerate is massive and poorly bedded, with quartz and rounded to sub-rounded quartzite pebbles and cobbles in an argillite matrix.

The overlying Moina Sandstone unit is extensively exposed in the mine area and is massive to weakly bedded, coarse to fine grained, and argillaceous. It is predominantly composed of quartz clasts with a matrix of finer quartz and clays and shows reddish colouration of iron oxide and hydroxide. Close to the granite contact, the rock is massive, compact, and hornfelsed, with the development of epidote and chlorite. The overlying Gordon Limestone is restricted to the north of the deposit area and appears to have been converted to skarns in the south where carbonate rocks are not intersected by drilling. The impure limestone is interpreted to overlie the Moina Sandstone. The unit consists of sandstone intercalated with carbonate rocks, overlain by limestone, which commonly occurs as coarse-grained, recrystallised, massive marble.

Devonian granite is exposed in open cuts (Kara 1, Bob's Bonanza and Lohreys Pit; fig. 29). The granite may occur in contact with skarns and is intersected in drill core from several skarn bodies. It is pink, massive, coarse grained and usually equigranular, although porphyritic and medium-grained varieties are observed in places. The granite is composed of K-feldspar, quartz, plagioclase and accessory minerals consisting of biotite, muscovite, epidote, hornblende, and magnetite. Epidote, black to green amphibole (hornblende to hastingsite) and some magnetite occurs as secondary minerals in the granite. Amphibole appears to replace biotite. The granite generally shows increasing intensity of alteration towards its margin, with feldspars being altered to sericite or epidote.

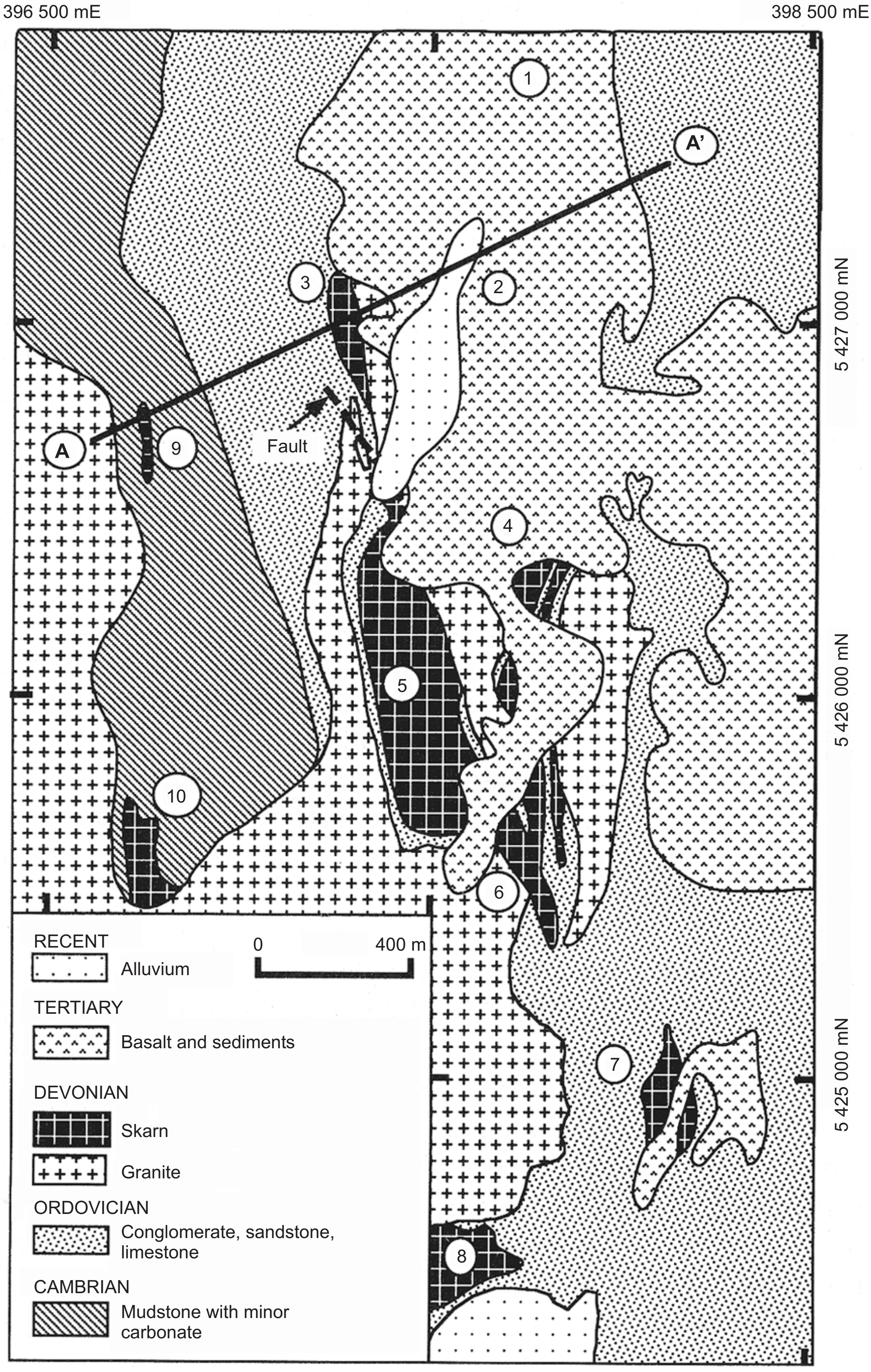


Figure 28

Deposit geology and skarn bodies of the Kara area, northwestern Tasmania. 1 = L5; 2 = Kara North magnetic zone; 3 = Kara North 266 zone; 4 = Eastern Ridge; 5 = Kara 1; 6 = Bob's Bonanza; 7 = Lohreys Pit; 8 = Kara South; 9 = L10; 10 = L9. Adapted from geological maps of Tasmania Mines NL. A–A′ = location of cross section shown in Figure 31 (from Khin Zaw and Singoyi, 2000).

Figure 29

Granite and epidote alteration, showing a very irregular granite surface, partly epidotised, overlain by an epidote skarn.

Figure 30

Contact of granite (centre, yellowish) with hornfelsed Moina Sandstone (LHS, grey) and a fault contact with skarn (RHS, dark grey), Kara mine.

The Kara magnetite-scheelite deposit consists of several skarn bodies, including Kara South, Kara 1, Bob's Bonanza, Kara North 266, Kara North magnetic zone, and L5 lenses (fig. 28). The majority of these bodies have developed within the Ordovician sequence of the Gordon Limestone Subgroup, including the lower unit of intercalated carbonates and sandstone which are mostly comprised of limestone in the southern skarn bodies (e.g. Kara 1 and Bob's Bonanza). The skarns are in direct contact with the granite but may be separated from it by a thin layer of Owen Conglomerate or Moina Sandstone away from the contact.

The major structural feature of the deposit area is mid-Palaeozoic folding of Cambrian and Ordovician sequences (fig. 31). The fold structures have been intruded by Devonian granite, and the beds in the folded strata are either semi-concordant to the granite intrusion or truncated by it. The folds generally trend north-south and have limbs that are shallow to very steep dipping and locally overturned. At the Kara 1 ore body, Ordovician sandstone beds on the western side of the pit are nearly vertical to overturned and dip to the east, indicating that a syncline is situated to the east of the pit. This syncline has been extrapolated to extend northward and is enveloped by the underlying granite intrusion and contains skarns (fig. 31). Mine geologists consider the syncline to plunge gently (15°) towards the north. Faults do not constitute a major deformation feature at the Kara deposit area, although minor faults have been mapped and interpreted by mine geologists. Most of the faults tend to displace the granite and, therefore, post-date granite emplacement.

Calcite and fluorite occur in veins in the altered zone. The contact zone between the granite and the Moina Sandstone is commonly fractured and consists of broken granite and sandstone fragments (fig. 30). This fragmentation is probably due to hydraulic pressures induced by late hydrothermal fluids emanating from the granite intrusion. Within the granite, the fragments contain matrix or linings of skarn minerals (epidote, quartz, and amphibole).

The principal skarn mineral assemblages at Kara are magnetite, garnet, vesuvianite, clinopyroxene, epidote and amphibolite, with fluorite, calcite, quartz,

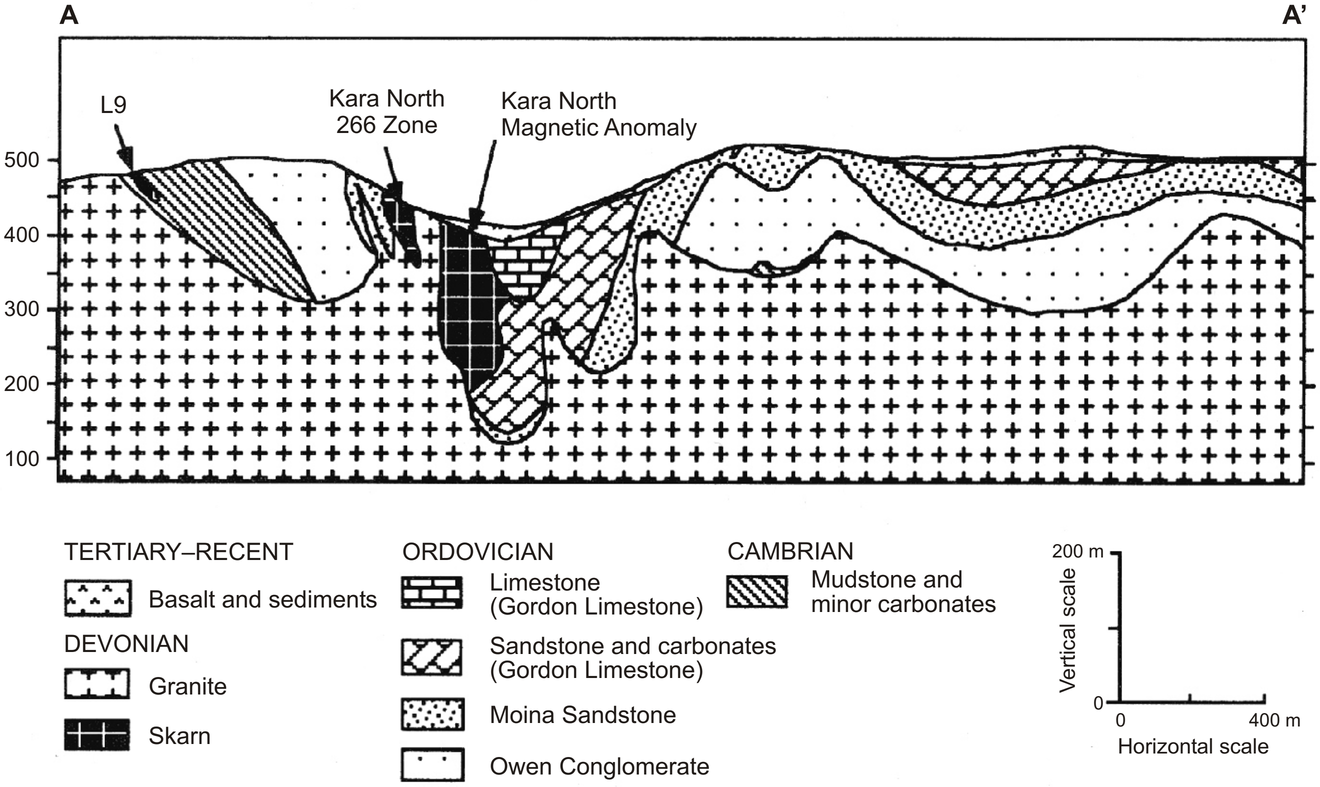


Figure 31

Cross section showing the geological structure on line A–A′ of Figure 28 (from Khin Zaw and Singoyi, 2000).

scheelite, hematite, chlorite, wollastonite, sphene, pyrite, chalcopyrite and apatite as subordinate or accessory minerals. At least four stages of skarn formation and ore deposition have been recognised at Kara (Table 3). Early mineral facies are largely anhydrous, whereas later ones are predominantly hydrous.

Stages I and II mineral assemblages represent early skarn formation and are dominated by anhydrous minerals of clinopyroxene and garnet. Stages III and IV minerals represent late skarn-forming phases and pervasively replace early mineral assemblages. Scheelite occurs in stages I through to III and generally shows a close spatial association with hydrous minerals (vesuvianite in stage I and/or II and amphibole in stage III). Abundant scheelite is found in stage III with amphibole, where it forms very coarse grains locally in excess of 50 millimetres.

In summary, the Kara deposit formed as a proximal skarn assemblage in carbonate host rocks following the emplacement of the Devonian Housetop Granite and is characterised by early, high-temperature mineral assemblages dominated by anhydrous minerals and late low-temperature assemblages with abundant mineralisation (scheelite, magnetite) and hydrous minerals. Chemistries of clinopyroxene, garnet, and scheelite suggest that the Kara skarn deposit was formed under relatively oxidised conditions. The deposit differs significantly from other scheelite deposits, such as Can Tung (Canada), which formed under relatively reduced conditions in the abundance of magnetite (up to >90 vol.%), lower quantities of sulphide minerals, and absence of pyrrhotite. Figure 32 shows the proposed genetic modelling of the Kara deposit.

Field stops, Kara mine

The Kara 1 deposit is the largest known magnetite-scheelite deposit in this area (fig. 12). Contacts with the Housetop Granite and Moina Sandstone can be observed, as well as a range of skarn types and mineralisation as described below.

1. **Housetop Granite**: Aplitic intrusions in skarn occurring in the pit floor are medium-grained and irregular bodies with calcic alteration and veining near the skarn contact. The alteration takes the form of coarse-grained hedenbergitic pyroxene, hornblende and quartz-epidote aggregates.

2. **Epidote skarn** (retrograde skarn): Yellow-green, massive, granular and often with quartz, bavenite and fluorite; mostly near the granite and may vein and partly replace granite.

3. **Granite** (dyke?) on western wall: Medium to coarse grained, aplitic, pink, irregular pods in hornfels.

4. **Cambro-Ordovician hornfels**: At the granite contact these form hard green rocks composed of quartz with lesser clinopyroxene, chlorite, epidote and mica.

5. **Fault Zone**: The structure of the area is poorly understood, but there appears to be a major fault on the western side of the Kara 1 pit, that may partly

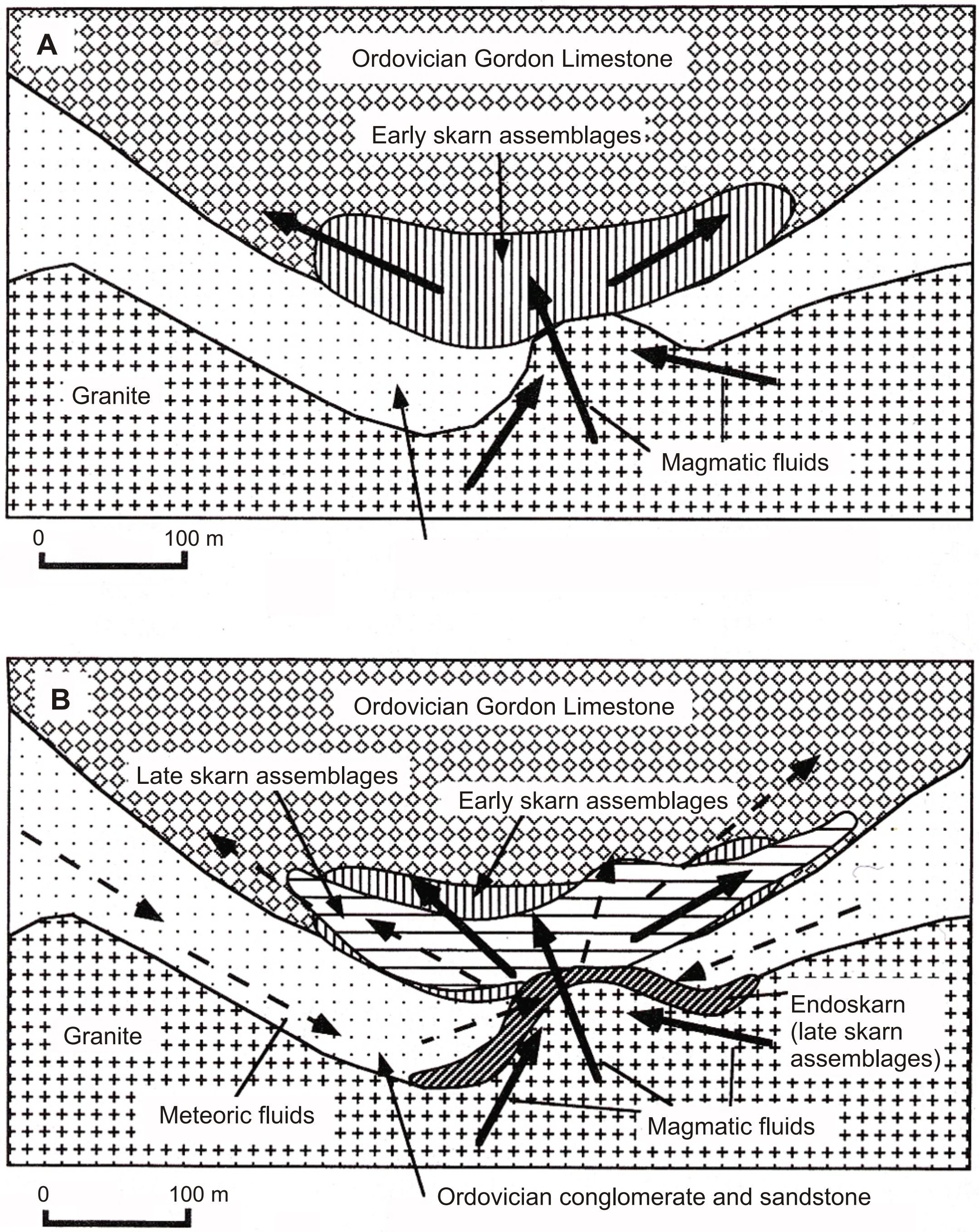


Figure 32

Idealised cross sections showing the genesis of Kara skarn deposit.

control the mineralisation and is gold-anomalous (McKeown, 1994). Here it is clayey and contains massive goethite with minor quartz and magnetite.

6. **Pyroxene skarn** (diopside to hedenbergite): Massive, granular, pale to bright green, with minor magnetite and amphibole.

7. **Vesuvianite-magnetite skarn**: Magnetite can form cross-cutting veins and large pods in the fine-grained grey-greenish vesuvianite, usually mixed with some andradite.

8. **Garnet skarn** (andradite): Coarse grained, dark to pale brown or greenish, it can form pods and cross-cutting veins, and may be mixed with magnetite, calcite, amphibole and minor scheelite.

9. **Amphibole skarn** (retrograde skarn): Uncommon, dark green to white, may be asbestiform. Compositions include tremolite, actinolite, hastingsite, magnesiohastingsite, ferroedenite, and possibly others (Khin Zaw and Singoyi, 2000). It can include significant amounts of the rare, red Be-mineral danalite.

10. **Mineralisation**: At the Kara No. 1 deposit, the ore-grade scheelite and magnetite mineralisation is hosted by andradite and vesuvianite-rich skarn which forms an irregularly-shaped blanket draped 15–25 m above the granite. This mineralisation is separated from the granite by tungsten-poor, epidote and pyroxene-rich skarns. The highest tungsten values are mostly within the lower parts of the magnetite-rich zones. The prograde skarns are variably retrogressed to amphibole skarns epidote and fluorite. Scheelite is disseminated mostly in magnetite-rich skarn. It is usually milky white but rarely yellow, and usually contains minor MoO_3 (<3 wt.%; Khin Zaw and Singoyi, 2000). Tin is anomalous (<0.1%) in the skarn and appears to be present within the structure of various silicates, especially andradite, and is not recovered. Minor sulphide minerals are present in the skarns, and include pyrite, aikinite, molybdenite, chalcopyrite, sphalerite, arsenopyrite, bismuthinite and galena.

Table 3

Schematic diagram showing paragenetic relationships in skarn assemblages, Kara mine (adapted from Khin Zaw and Singoyi, 2000).

Mineral	Stage 1	Stage 2	Stage 3	Stage 4
Clinopyroxene	———		- - - - -	
Garnet	- - - - -	———	- - - - -	
Vesuvianite	- - - - -	———	- - - - -	
Amphibole			———	
Epidote			———	
Quartz			———	- - - - -
Fluorite			———	- - - - -
Calcite			———	———
Chlorite			———	
Wollastonite	- - - - -			
Apatite		- - - - -		
Sphene			- - - - -	
Magnetite		- - - - -	———	
Scheelite	- - - - -	- - - - -	———	
Hematite			- - - - -	———
Pyrite			———	
Chalcopyrite			- - - - -	- - - - -

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