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Explanatory notes for the Time–Space Diagram and Stratotectonic Elements Map of Tasmania

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CONTENTS

Introduction

For the purposes of this report, Tasmania is divided into seven Proterozoic–lower Palaeozoic regions or 'Elements' with differing geological histories (fig. 1). These elements are summarised in a Time-Space Diagram (Plate 1) and comprise King Island, the Rocky Cape Element, the Dundas Element, the Sheffield Element, the Tyennan Element, the Adamsfield–Jubilee Element, and the Northeast Tasmania Element. A very generalised west-to-east geographic variation within each element is indicated in the left-to-right arrangement of events shown within each column. The flat-lying rocks of the Tasmania Basin and younger successions overlap Element boundaries. Thus, column margins are shown as dashed lines in the upper part of the time-space diagram and retained only as a guide to geographic position. Discussion in the text is subdivided in a corresponding way, i.e. according to Element boundaries up to and including the Devonian–Early Carboniferous granitoid intrusives, but beyond this according to major basin sequences which overlap the Element boundaries. A new Stratotectonic Elements Map of Tasmania at a scale of 1:500 000 is included as Plate 2.

This work is a compilation of existing published and unpublished data and interpretations, in preparation for the interpretative phase of the 'TASGO' Project of the National Geoscientific Mapping Accord (NGMA). Substantial input and editorial comment has been provided by Tony Yeates of the Australian Geological Survey Organisation (AGSO); by Tasmania Development and Resources (TDR) geoscientists Tony Brown, Keith Corbett, Steve Forsyth, Marcus McClenaghan, Ralph Bottrill, Bob Richardson and Geoff Green; and by Nic Turner, previously of TDR. Tony Yeates and Barry Drummond of AGSO also had substantial input in the design and planning phase of the project.

The new AGSO time-scale used as a template for the Time-Space Diagram (Plate 1) is a provisional draft as at May 1994.

Values of decay constants for radiometric dates are after Steiger and Jäger (1977) and Dalrymple (1979), and are as follows:–

^{40}K: $\lambda\beta = 4.962 \times 10^{-10}$ yr^{-1};
$\lambda\varepsilon = 0.581 \times 10^{-10}$ yr^{-1};
^{40}K/K $= 1.167 \times 10^{-4}$ moles/mole.

^{87}Rb: $\lambda = 1.42 \times 10^{-11}$ yr^{-1}.

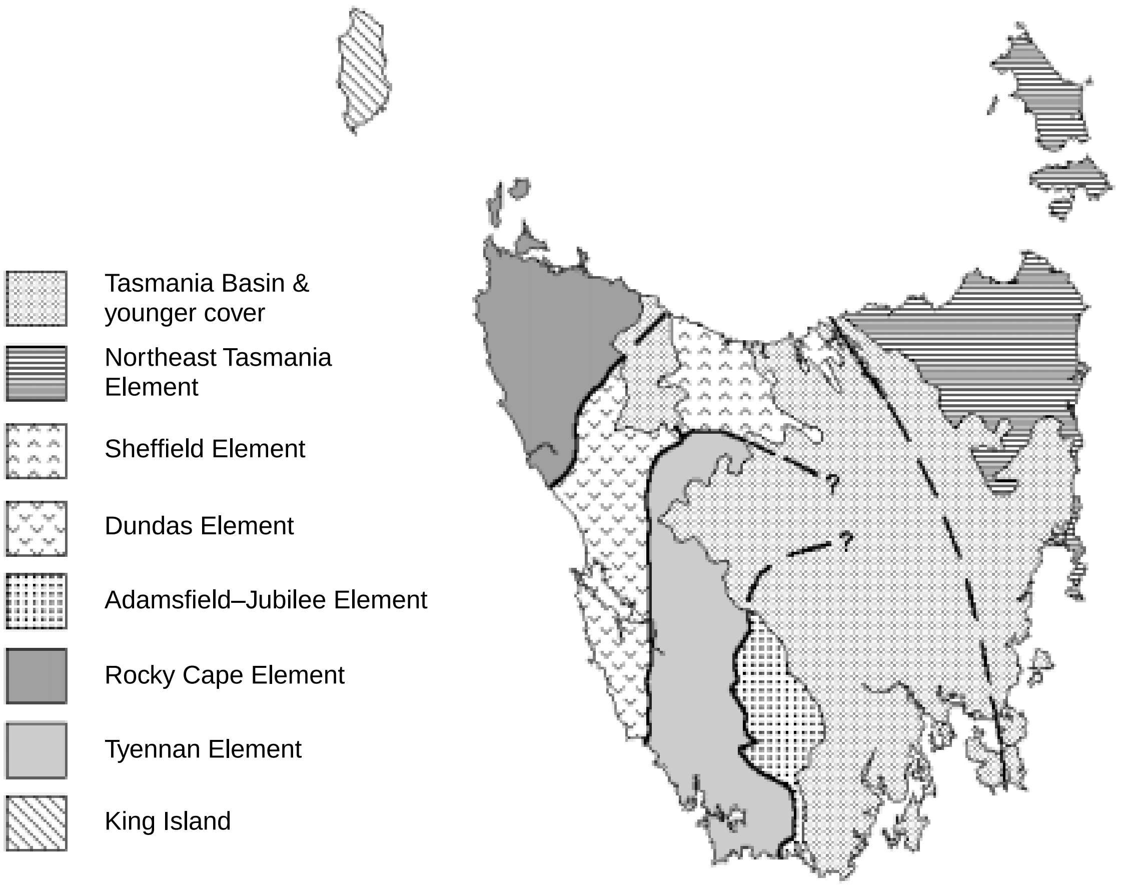

Figure 1

Main subdivisions or 'Elements' used for the Time-Space diagram. Heavy dashed lines are concealed Element boundaries

?Mesoproterozoic

The oldest exposed rock sequence on King Island, forming the western half of the island, is a succession more than 1000 m thick of multiply-deformed amphibolite facies metasedimentary rocks with minor mafic intrusives (Cox, 1973, 1989). Turner *et al.* (1992) suggest that the original depositional age of the sequence may be similar to that of the sedimentary protolith of the metamorphic rocks of the Tyennan Element, i.e. 1100–1150 Ma indicated by Rb-Sr model ages (Råheim and Compston, 1977). The lithologies are dominantly quartzofeldspathic schist with minor quartzite, pelitic schist, and rare thin calcareous lenses. The typical schist mineral assemblage is quartz + muscovite + biotite (+ plagioclase). The dominant lithology forms massive to laminated units up to several metres thick, and its protolith was probably a feldspathic quartz sandstone. Quartzite and micaceous quartzite are also generally massive to finely laminated, or rarely cross laminated. Hornblende amphibolite sills, with compositions similar to tholeiitic basalt, were emplaced prior to regional deformation (Cox, 1989).

Neoproterozoic orogenesis and granitoid intrusives

Polyphase deformation of the ?Mesoproterozoic sequence was in part broadly synchronous with a period of Precambrian granitic intrusive activity and metamorphism to amphibolite facies (Cox, 1989), at ca. 760 Ma (see below). Garnet-biotite geother- mometry indicates temperatures of 470–580°C, while the presence of andalusite and rare phengite suggests low pressures of 100–300 MPa (Blackney, 1982; Turner, 1989*a*). These conditions are consistent with high-level contact metamorphism.

According to Cox (1989), the first major deformation phase (D_1) produced tight to isoclinal folds with penetrative axial surface foliation. Prograde metamorphism commenced during D_1, and S_1 microfabrics are defined by amphibolite facies mineral assemblages. Major granitic intrusive activity post-dated D_1 folding, but may have commenced in late D_1. The dominant and earliest granitic intrusive phase is an S-type, K-feldspar porphyritic biotite adamellite, although a number of other later minor phases are present. U-Pb geochronology of igneous zircons from the granitoids indicated an age of 760 ± 12 Ma (Turner, 1993*a*; Black, 1994). Previous Rb-Sr muscovite ages (McDougall and Leggo, 1965) were reported in Cox (1989) as 730 Ma and 726 Ma. The U-Pb method also revealed the presence in the intrusive of inherited zircon grains, with ages ranging from 1200–1800 Ma (one grain yielded an age of 2900 Ma). D_2 structures (including open to tight minor folds) deformed the metamorphic assemblages as well as many of the granitic rocks, however minor granitic intrusive activity also post-dates D_2 and D_3. D_3 folds are moderately to gently inclined open structures. Upright D_4 folds post-date all granitic intrusive activity, but appear to be cut by a swarm of tholeiitic dolerite dykes which may be related to the ?Late Proterozoic mafic extrusive rocks on the east coast of the island (Cox, 1989).

?Neoproterozoic sequences

An inferred unconformity separates the metamorphic complex with its granitoid intrusive rocks from a sequence of relatively unmetamorphosed argillaceous sedimentary rocks, presumed to be of Neoproterozoic age, which forms the majority of the eastern half of King Island (Waldron *et al.*, 1993). Along the southeast coast this siltstone sequence is overlain, with apparent conformity, by a sequence of (from bottom to top) siliceous sandstone, siltstone, diamictite, dolomite, tuff and mafic volcanic rocks, which has generally been assumed to correlate with lithologically similar sequences in the Smithton Synclinorium of northwest Tasmania. However, Waldron *et al.* (1993) cast doubt on a direct one to one correlation, largely on the basis of comparisons of the geochemistry of the mafic volcanic rocks. The stratigraphy of the southeast coast sequence, summarised from Waldron *et al.* (1993), is shown below:–

TOP	(Not exposed)
Unit 7	Tholeiitic basalt, in porphyritic and non-porphyritic flows.
Unit 6	Picritic pillow lavas, flow units, breccia and hyaloclastite units.
	.... *DISCONFORMITY* ....
Unit 5	Tholeiitic mafic volcanic rocks, with lava phases dominated by massive and pillow basalt flows.
Unit 4	Laminated siltstone with minor volcaniclastic lithicwacke and tuff.
Unit 3	Finely laminated dolomite with some well-developed cross lamination associated with channel-fill structures. Remnant oolitic and pelletal textures and microstylolites.
Unit 2	Diamictite with subangular clasts of quartzite, meta-siltstone and carbonate rocks in either a red hematitic or grey calcareous matrix.
Unit 1	Shallow-water, clean or muscovitic quartz sandstone, cross-bedded in part.
BASE	(Conformable contact with relatively unmetamorphosed siltstone sequence)

Some workers have considered the diamictite to be of glaciogenic origin (Waterhouse, 1916; Carey, 1947), while Jago (1974) suggested it may represent a submarine density flow deposit derived from an area affected by glaciation. However, Waldron *et al*. (1993) favour a non-glaciogenic, density-flow origin. The dolomite is believed to have originated by early diagenetic dolomitisation of limestone, and relict textures indicate a subtidal to intertidal depositional environment (Waldron *et al*., 1993). In the mafic volcanic succession, Unit 5 shows intrusion of lava lobes into hydroplastic sediment, while in Unit 7 flows occur in repetitive cycles separated by beds of pebble-cobble conglomerate, and the non-porphyritic flows have uniform bases and amygdaloidal tops. Numerous dykes of picritic composition have been recorded, and dykes petrochemically similar to the Unit 7 tholeiitic volcanic rocks intrude Units 5 and 6 as well as the underlying sedimentary successions. Waldron *et al*. (1993) suggest an intracontinental rift setting for the King Island volcanosedimentary sequence, with the volcanic rocks generated by variable degrees of partial melting of the upper mantle.

The southeast coast section shows fairly consistent dips of about 50° to the east and southeast, and is offset by numerous faults, indicating that its deposition was followed by at least one further period of significant folding and faulting, of unknown age(s). Several thick augite syenite dykes cut the lower sedimentary rock succession, and are themselves offset by faults in places. The dykes are essentially of unknown age but Waldron *et al*. (1993) suggest a possible genetic relationship with the volcano-sedimentary succession.

Early Carboniferous granitoid intrusive rocks

In the Early Carboniferous the volcano-sedimentary sequences in the eastern part of King Island were intruded by several small granitic stocks and associated aplitic and pegmatitic dykes. The granitoids are porphyritic with large pink K-feldspar phenocrysts, and are classified as adamellite- granodiorite based on modal analyses (Camacho, 1989). Fission-track dating of sphene (Gleadow and Lovering, 1978) yielded results mostly close to 350 Ma, in good agreement with previous K-Ar ages (McDougall and Leggo, 1965).

Two of the stocks (the Grassy Granodiorite and the Bold Head Adamellite) intrude the southern extension of the volcano-sedimentary sequences, where contact metamorphism is associated with scheelite skarn mineralisation formed by selective metasomatic replacement of carbonate horizons (Large, 1971; Danielson, 1975; Kwak, 1978; S. G. Brown 1989). The Bold Head intrusive is bounded to the west by a major north-trending fault which post-dates intrusion and mineralisation, and it has been suggested that this body is probably a faulted sliver of the Grassy intrusive (Wesolowski, 1981). The Sea Elephant Adamellite, located in the northeast of the island, is more felsic than the Grassy and Bold Head bodies, and is considered to be their fractionated equivalent (Camacho, 1989).

Rocky Cape Element (DBS)

?Mesoproterozoic: Rocky Cape Group

In current interpretations (e.g. Turner *et al*., 1992), the oldest exposed sequence in the Rocky Cape Element (fig. 1) comprises the Rocky Cape Group and correlates, which form the most areally-extensive outcrop in the region. No present data directly indicate the age of deposition of this sequence, but Turner *et al*. (1992) suggest an age in the vicinity of the ca. 1100 Ma depositional/provenance age determined for metapelite in the Tyennan Element (Råheim and Compston, 1977). The nature of the basement to the Rocky Cape Group is unknown. In its type section on the north coast (Gee, 1968; Gee *in* Turner, 1989*a*), the exposed sequence is 5700 m thick, and comprises four main units:–

TOP	(Not exposed)
Jacob Quartzite	Supermature quartzarenite with silica cement, with abundant planar crossbedding.
Irby Siltstone	Dominantly siltstone with black shale, dolomite, sub-greywacke and hematitic breccia.
Detention Subgroup	Dominantly supermature cross-bedded quartz sandstone with interbedded siltstone.
Cowrie Siltstone	Black pyritic shale with interbedded siltstone.
BASE	(Not exposed)

Gee (1971) considered the initial basin to have formed on a stable shelf, initially starved of clastic input which increased with time, culminating in the accumulation of unusually thick blankets of shallow marine mature sands. Palaeocurrent directions are bimodal, with northwesterly and southeasterly transport directions.

Burnie and Oonah Formations

A northeasterly-trending metamorphic belt known as the Arthur Lineament (Gee, 1967) separates the Rocky Cape Group (to the west) from the Burnie and Oonah Formations (to the east). The Burnie Formation consists dominantly of sandy turbidite-facies quartzose wacke and slaty mudstone with minor altered mafic pillow lava, whereas the Oonah Formation, while of similar general character, is more varied and contains additional lithologies including relatively clean quartz sandstone, dolomite, chert and conglomerate (Turner, 1989I).

Gee (1967) considered that the Rocky Cape Group represents the oldest exposed sequence in the Rocky Cape Element, while the Burnie and Oonah Formations are younger and probably coeval. Current thinking (e.g. Turner *et al*., 1992; 1994) also favours a considerable time-span between the deposition of the Rocky Cape Group and that of the Burnie and Oonah Formations, while at least partial equivalence of the latter two units is supported by similarity in the oldest K-Ar slate ages obtained from them (690 ± 10 Ma, Adams *et al*., 1985; Turner, 1989*a*). Furthermore, a possible correlation mentioned by Gee (1967) is supported by Turner *et al*. (1992; 1994), who suggest that the Oonah and Burnie Formations are laterally equivalent to basal shallow-water siliciclastic rocks in the Togari Group of the Smithton Synclinorium, and to basal proximal turbidite fan deposits in the Ahrberg Group (see below), both of which sequences rest unconformably on the Rocky Cape Group.

The Oonah and Burnie Formations thus represent an important link between Neoproterozoic, mainly shallow-marine shelf sedimentation in the Rocky Cape Element, and the commencement of Neoproterozoic–Late Cambrian mainly deeper marine sedimentation in the Dundas and Sheffield Elements. Further discussion of these formations appears in the appropriate sections.

Smithton Synclinorium

In the northwestern part of the Rocky Cape Element, the Rocky Cape Group is unconformably overlain by a sequence of clastic sedimentary rocks, mafic volcanic rocks, dolomite and chert which forms the Smithton Synclinorium. Williams (1979) and Baillie (*in* Brown, 1989*a*) ascribed the unconformity to the Penguin Orogeny of Spry (1962), an event thought to have been responsible for most of the deformation in the Rocky Cape Group, the Arthur Lineament, and the Burnie and Oonah Formations, but to have pre-dated deposition of units in the Smithton Synclinorium. However, Gee (1967, 1968) argued that the folding of the Rocky Cape Group (and thus the Penguin Orogeny) can be shown to have affected at least the lower units in the Smithton Synclinorium. This relationship is supported by subsequent regional geological mapping (Seymour and Baillie, 1992; Everard *et al*., in press), which has demonstrated a general similarity in structural complexity and tightness of folding within the Smithton Synclinorium compared with its immediate Rocky Cape Group basement. This is consistent with current interpretations based on correlations of deformed and metamorphosed rocks in the Ahrberg Group with units in the Smithton Synclinorium (Spry, 1964; Turner, 1990*d*; Turner *et al*., 1992, 1994), which imply that the major regional deformation and metamorphism in the Rocky Cape Element is relatively young and post-dates deposition of most of the units in the Smithton Synclinorium. This conclusion is supported by geochronological studies in the Arthur Lineament, which date the peak regional metamorphism at around 500 Ma (Turner *et al*., 1992, 1994; Turner, 1993*b*).

The unconformity at the base of the lowermost sequence in the Smithton Synclinorium (the Togari Group, see below) is best exposed at the northeastern margin of its present outcrop area. Considerable erosion is evident at the contact, but the maximum discordance is 22° (Gee, 1967, 1968), which Gee (1967) suggested could result from a depositional dip component in the overlying conglomerate. In the western part of the basin the contact is strongly erosional and commonly concordant, with a maximum recorded discordance of less than 10° (Seymour and Baillie, 1992). However, the recent detailed mapping has confirmed that regionally the basal units of the Togari Group rest on different parts of the Rocky Cape Group basement along the contact, as was first suggested by the reconnaissance mapping of Longman and Matthews (1961). This suggests that the unconformity represents at least a period of gentle regional folding prior to Togari Group deposition. This early deformation may correlate with the 760 Ma folding on King Island (Turner *et al*., 1994; Berry, 1994). Also provisionally grouped with this event on the Time-Space Diagram (Plate 1) is an episode of wrench faulting, based on the recognition of a major east-west trending dextral wrench fault, with an offset of some 8 km, which cuts the Rocky Cape Group (Gee, 1971; Lennox *in* Brown, 1989*a*) but which, based on equivocal field evidence, appears to intersect but not affect the Smithton Synclinorium (Lennox *et al*., 1982).

As shown by a regional cross-section based on recent detailed mapping (Seymour, p.473 *in* Williams, 1989), the Smithton Synclinorium is a large, open, generally upright structure, and its present outcrop limits bear no necessary relationship to the original extent of deposition of the units within it. The lower four main units in the synclinorium form the Togari Group (Everard *et al*., in press), which is separated by a considerable hiatus and inferred disconformity from the fossiliferous uppermost unit, the late Middle–early Late Cambrian Scopus Formation. The Togari Group stratigraphy, with approximate thicknesses, is shown below.

The Forest Conglomerate and Quartzite is discontinuous and contains clasts derived predominantly from Rocky Cape Group quartzites. Chert lithologies in the Black River Dolomite are apparently derived from silicification of carbonate, and preserve primary oolitic textures which indicate a shallow marine (shelf) depositional environment. The Black River Dolomite includes discontinuous units of dolomitic diamictite containing clasts of dolomite, stromatolitic dolomite and oolitic chert in a dolomitic matrix (Everard *et al*., in press). The stromatolite *Baicalia* cf. *B. burra* is present in clasts in the diamictite (Brown, 1985; Griffin and Preiss, 1976) and more recently *Conophyton garganicum* has also been found (C. Calver, pers. comm.). These forms are known from the Willourian and Torrensian of South Australia (Preiss, 1987).

The Kanunnah Subgroup is correlated with the Crimson Creek Formation of the Dundas Element (Brown, 1989*a*). The lower Kanunnah Subgroup includes discontinuous units of diamictite containing clasts of mafic and felsic volcanic rocks, dolomite, chert and mudstone in a fine-grained non-dolomitic matrix (Everard *et al*., in press). Basaltic lavas in the Kanunnah Subgroup are generally clinopyroxene and/or plagioclase-phyric tholeiites (Brown, 1989*a*). However, olivine-phyric tholeiitic lavas in the Smithton area show distinctive REE patterns which indicate that they belong to a different geochemical group, which may be younger than the main phase of Kanunnah Subgroup volcanism (Brown and Waldron, 1982; Brown and Jenner, 1989; Brown, 1989*b* and unpublished data). Control on the age of the main phase of mafic volcanism is provided by dating of dolerite dykes intruding the Rocky Cape Group basement, as one of four geochemical groups of these dykes has been interpreted as feeders for the Smithton Synclinorium basalts, and has yielded K/Ar ages of 600 ± 8 Ma and 588 ± 8 Ma (Brown, 1989*a*).

The Smithton Dolomite shows oolitic and other textures indicative of a shallow marine sedimentary environment, but it is less commonly silicified than the Black River Dolomite.

Recent carbon and strontium isotope chemostratigraphic studies of the Smithton Synclinorium carbonate sequences suggest a Cryogenian age for the Black River Dolomite (in

TOGARI GROUP STRATIGRAPHY (SMITHTON SYNCLINORIUM)

TOP		(Inferred disconformity at base of Scopus Formation)
Smithton Dolomite	~1.5 km	Interbedded dolomite and dolomitic limestone.
Kanunnah Subgroup	~1 km	Interbedded mudstone, siltstone and turbiditic lithicwacke with mafic volcanic detritus. Several major phases of mafic lava. Discontinuous lower units of polymict diamictite.
Black River Dolomite	600 m	Interbedded dolomite, chert, siltstone and black mudstone. Discontinuous units of dolomitic diamictite.
Forest Conglomerate and Quartzite	0–500 m?	Interbedded siliceous conglomerate and cross-laminated orthoquartzite.
BASE		(Unconformity on Rocky Cape Group)

broad agreement with the stromatolites), and an age of ca. 580 Ma (Neoproterozoic III) for the Smithton Dolomite. The diamictite at the top of the Black River Dolomite is a possible correlate of the Sturtian glacials in the Adelaide Fold Belt (C. Calver, pers. comm.).

The uppermost unit in the Smithton Synclinorium, the Scopus Formation, rests with inferred disconformity on the Smithton Dolomite (Everard *et al.*, in press). This unit consists of interbedded mudstone, siltstone and lithic wacke, and contains fossils indicating biostratigraphic ages between Boomerangian and Idamean (Jago and Buckley, 1971; Jago, 1976; Baillie, 1981), straddling the Middle–Late Cambrian boundary. The Scopus Formation is correlated with the Dundas Group of central western Tasmania. A sedimentological study of part of the succession by Baillie and Jago (1995) indicated deposition within a submarine basin-floor turbidite-fan complex. Two lithofacies associations were recognised: a coarse-grained association of thick-bedded coarse-grained sandstone, pebbly sandstone and granule conglomerate, interpreted as channel-fill deposits; and a thin-bedded association of fine-grained sandstone, siltstone and mudstone, interpreted as overbank deposits consisting predominantly of mud turbidites. Palaeocurrents from both associations are directed towards the north and are statistically indistinguishable. Petrographic data indicate that the sandstones were derived predominantly from felsic and mafic volcanic sources, but that a metamorphic source similar to that in the oldest exposed Tasmanian Precambrian terrains also supplied some sediment.

Ahrberg Group

Near Corinna, on the western flank of the southern part of the Arthur Lineament, the Ahrberg Group comprises an east-facing succession of quartzose sandstone and conglomerate, followed by mudstone, dolomite and tholeiitic metavolcanic rocks, resting unconformably on basement equivalent to the Rocky Cape Group (Spry, 1964; Turner, 1990*d*; Turner *et al.*, 1991). Despite the unconformable relationship, the major deformation in the succession is equivalent to that in the underlying rocks, and so deposition of the Ahrberg Group apparently pre-dated the Penguin Orogeny (Turner, 1990*d*). This is consistent with the current interpretation of the Togari Group in the Smithton Synclinorium (see above).

Turner (1990*d*) describes the Ahrberg Group as follows. The basal formation is interpreted as a proximal, marine fan deposit. It fines upward into mudstone with cherty beds, then into the fine-grained, stromatolitic Savage Dolomite which was deposited in very shallow marine conditions. The overlying Bernafai Volcanics is schistose, with a mineral assemblage of albite-epidote-actinolite-chlorite. This unit includes lavas and fragmental rocks, and poorly cleaved, slaty or phyllitic pelite. Other formations, possibly dolomitic, may overlie the Bernafai Volcanics.

High-purity silica deposits in the Corinna area consist of mainly angular, incoherent quartz silt and fine sand ('silica flour'), and form an irregular layer between dolomitic rocks of the Ahrberg Group and unconsolidated Tertiary fluvial gravel (Turner, 1990*d*). A residual origin after silicified dolomite is indicated, and the deposits represent a significant industrial mineral resource.

Regional deformation and metamorphism

Introduction: the Penguin and Delamerian Orogenies

Based on correlation of the Ahrberg Group in the Corinna area with the Togari Group, the major phase of deformation and metamorphism in the Rocky Cape Element is currently considered to post-date deposition of the Togari Group (Turner *et al.*, 1992, 1994). This event is equivalent to the Penguin Orogeny of Spry (1962), particularly in the way it was interpreted by Gee (1967, 1968). It is believed to be responsible for the bulk of the deformation and metamorphism in the Arthur Metamorphic Complex (see below), as well as much of the polyphase deformation of the Burnie and Oonah Formations, and much of the folding and cleavage development in the Rocky Cape Group and the Smithton Synclinorium.

The ca. 500 Ma revised age for the Penguin Orogeny is constrained by K-Ar mineral ages on metamorphic amphibole from the Arthur Metamorphic Complex, supported by similar Ar-Ar and K-Ar mineral ages from the metamorphic rocks of the Forth Inlier in the Sheffield Element, and similar K-Ar and Rb-Sr mineral ages from the metamorphic rocks of the Tyennan Element (Turner *et al.*, 1992, 1994). The Penguin Orogeny has been correlated with the earliest phase of the Delamerian Orogeny of South Australia, and the apparently equivalent Ross Orogeny of North Victoria Land, Antarctica (Turner, 1993*b*; Turner *et al.*, 1994; Corbett, 1994; Turner and Crawford, in prep.). There were very strong regional variations in the intensity of deformation and metamorphism, the highest strain zones occurring in the Arthur Lineament, while the Smithton Synclinorium shows negligible metamorphism and relatively low to moderate strain, perhaps because it was cratonic at the time of deformation, as implied in the palaeogeography suggested by Turner *et al.* (1992, 1994).

Arthur Lineament

The rocks comprising the Arthur Lineament are known as the Arthur Metamorphic Complex (Turner, 1989*a*), and consist in part of metamorphic equivalents of the Rocky Cape Group and the Oonah Formation. The metamorphic complex is best known in the south (Spry, 1964; Urquhart, 1966; Turner, 1990*d*; Turner *et al*., 1991, 1992, 1994), particularly around the Savage River iron ore mine (Spiller, 1974; Coleman, 1975; Matzat, 1984). It is also well known in and north of the Arthur River area (Gee, 1967, 1971, 1977; Everard *et al*., in press), but the central part of the belt is poorly known.

In the northern part of the Arthur Lineament, the western part of the metamorphic complex comprises a narrow belt of phyllite and schistose quartzite which is transitional into the relatively unmetamorphosed Rocky Cape Group (Gee, 1977; Everard *et al*., in press; Turner, 1990*d*). The western boundary is different near Corinna in the southern part of the lineament, where Turner (1990*d*) noted a sharp increase in the degree of metamorphic recrystallisation across a fault at the eastern edge of the Ahrberg Group. However, the presence of schistose volcanic rocks, slate and phyllite in the eastern part of the Ahrberg Group led Turner (1990*d*) to conclude that a transitional metamorphic boundary, equivalent to that forming the western margin of the complex in the north, lies somewhere within the Ahrberg Group. In both the Pieman River area in the south (Turner, 1984, 1990*d*; Brown, 1986) and the Arthur River area in the north (Gee, 1977; Everard *et al*., in press), the eastern part of the metamorphic complex consists of interlayered micaceous quartzite, schist and phyllite which show a transitional decrease in grade eastward into their relatively unmetamorphosed equivalents in the Oonah Formation.

The main lithological association in the southern part of the metamorphic complex (Timbs Group of Turner *et al*., 1991) appears to be extensively developed throughout the Arthur Lineament, where it consists of pelitic and carbonate-rich schist with subordinate amphibolite and minor quartzose schist and carbonate. The amphibolite is generally of tholeiitic composition, and its protolith included mainly extrusive and probable shallow intrusive mafic rocks (Turner, 1990*d*). The schist shows greenschist mineral assemblages of quartz-white mica-chlorite-albite-carbonate ± biotite, while the amphibolite contains assemblages of actinolitic amphibole-albite-epidote-chlorite-carbonate-quartz -magnetite (Turner, 1990*d*). However, according to Turner *et al*. (1992), prograde assemblages in the complex are strongly retrogressed to actinolitic assemblages which are syntectonic with respect to the main deformation (D_2). High pressure, prograde mineral assemblages are locally preserved in blueschist facies metabasite, which contains glaucophane/crossite with compositional parameters consistent with crystallisation at about 700 MPa (Turner *et al*., 1992; Turner and Bottrill, 1993), and which match compositional parameters of blue amphiboles in the Sambagawa metamorphic belt of Japan.

The eastern part of the Timbs Group in the vicinity of the Savage River mine forms a distinct unit, the Bowry Formation (Turner, 1990*d*; Turner *et al*., 1991), which consists of pelitic schist with amphibolite and associated pyrite-magnetite lenses, magnesite and dolomite. Similar rocks occur in a similar position in the Arthur River area in the north (Everard *et al*., in press). Sparse bedding data suggest that the Bowry Formation is east-facing and concordant with the metamorphosed equivalents of the Oonah Formation (Turner, 1990*d*).

The most significant mineralisation in the Arthur Metamorphic Complex is the iron ore deposit at the Savage River mine (Coleman, 1975; Weatherstone, 1989). The orebody is a subvertical sheet, up to 150 m thick, consisting of magnetite-silicate- sulphide rocks in an association of metamorphosed mafic rock, serpentinite and carbonate, near the eastern boundary of the Bowry Formation (Turner, 1990*d*). The mafic rocks are of tholeiitic composition and include intrusives, tuffaceous rocks and minor pillow lavas. Ore minerals are magnetite and pyrite with minor chalcopyrite and trace sphalerite, ilmenite and rutile (Coleman, 1975). The mineralisation is currently thought to be of marine, volcanogenic origin, but previous genetic theories included magmatic segregation, late magmatic residual fluid, and hydrothermal replacement (Weatherstone, 1989). Other smaller lenses of magnetite-pyrite- minor chalcopyrite mineralisation indicate stratiform control over a strike length in excess of 70 km (Turner, 1990*d*). Minor gold occurs in quartz and carbonate veins at a number of places in, and to the east of, the Bowry Formation.

Magnesite deposits of considerable size occur within the Bowry Formation in both the Savage River area in the south of the Arthur Lineament, and in the Arthur River area in the north (Turner, 1990*d*). The magnesite most commonly occurs as a fine-grained, equigranular marble, and Frost (1982) favoured an origin by metasomatism of original dolomite by solutions containing $MgCl_2$, rather than a sedimentary origin.

Relationship to the Scopus Formation

The relationship of the ca. 500 Ma tectonothermal event to the uppermost unit in the Smithton Synclinorium (the Scopus Formation, Everard *et al*., in press) is enigmatic. It contains faunas straddling the Middle–Late Cambrian boundary, i.e. just after the major 500 Ma tectonothermal event, based on the time-scale used herein. However field relationships suggest that despite an inferred major disconformity at the base of the formation, it is structurally concordant with the underlying Smithton Dolomite (Seymour and Baillie, 1992). The regional structural cross-section (Seymour, p.473 in Williams, 1989) suggests that the Scopus Formation is affected by the same style and degree of folding as the underlying units. This enigma may indicate that the ages attributed to Stage boundaries in this part of the column are too young. Alternatively, the folding and cleavage development in the Scopus Formation may be Middle Devonian, which would imply a component of Devonian deformation in the rest of the Smithton Synclinorium and in its Rocky Cape Group basement.

Dolerite dyke swarms

Brown (1989*a*) recognised four groups of dolerite dykes in the northern part of the Rocky Cape Element, based on chemistry, petrography, and structural and stratigraphic constraints. His Group 1, intruding the Rocky Cape Group, are tholeiitic and are foliated and therefore pre-tectonic. Group 2, also intruding the Rocky Cape Group, are massive unfoliated tholeiites and may be feeders for the main phase of basaltic volcanism in the Kanunnah Subgroup of the Smithton Basin. Group 4 comprises olivine-two pyroxene-feldspar cumulates which form a dyke intruding part of the eastern Smithton Basin, and which show a distinctive REE pattern matching that of the olivine-phyric lavas in the Smithton area, and so may be feeders for the latter (A. V. Brown, unpublished data). Group 3, intruding the Rocky Cape Group, are massive, unfoliated, calc-alkaline and leucocratic, and are probably the youngest but least constrained in age.

A. J. Crawford (*in* Turner, 1992) also noted that the Rocky Cape Element dyke swarms include dolerites with trace element and REE signatures overlapping with the Smithton Synclinorium basalts, but he also recognised dyke samples with more strongly LREE-enriched patterns, lower Ti/Zr and Zr/Nb, and higher Ti/V compared with the Smithton basalts. He considered the latter dyke compositions to be transitional to P-MORB (plume-type mid-ocean ridge basalts). Following a generalised temporal sequence of decreasing enrichment with time, Crawford suggested that many of the Rocky Cape Element dykes were an early manifestation of an extended period of crustal attenuation that led to aborted rift development at the site of the present Smithton Synclinorium, and eventuated in ocean opening at some location east of the Rocky Cape Element.

Wurawina Supergroup

Correlates of the Owen, Gordon and Eldon Groups are exposed in the east-west trending Duck Creek Syncline, which intersects the west coast at the southern end of the Arthur Lineament (Blissett, 1962; Brown *et al*., 1994).

The basal unit is a siliciclastic pebble to cobble conglomerate which rests unconformably on metamorphosed equivalents of the Oonah Formation, and is considered to be a correlate of the Mt Zeehan Conglomerate (Blissett, 1962). The conglomerate is thin or absent in places, and is succeeded by cross-bedded quartzite considered equivalent to the sandstone sequence overlying the Mt Zeehan Conglomerate in its type area. The conglomerate-sandstone sequence is up to 150 m thick.

The Gordon Group correlate is represented by ~90 m of calcareous siltstone, shale and micrite (Blissett, 1962). Conodonts from the base of the section suggest an age equivalent to faunal assemblage OT12 (Banks and Baillie, 1989), which may imply the presence of a hiatus or condensed section between the Gordon Group section and the underlying siliciclastic rocks.

The Eldon Group section in the Duck Creek Syncline includes equivalents of the lower five formations of the group in its type area at Zeehan in the Dundas Element (Blissett, 1962). The stratigraphy, with approximate thicknesses, is shown below.

The thicknesses of all but the Crotty Quartzite are comparable to those of the equivalent units in their type area (see *Dundas Element*). Blissett (1962) believed that the Crotty Quartzite follows the Gordon Group conformably at Duck Creek, but the only age data on the relatively thin Gordon Group section suggest the presence of a hiatus or condensed section between the two sequences in this area (see Plate 1).

ELDON GROUP STRATIGRAPHY (DUCK CREEK SYNCLINE)

TOP		(Not exposed)
Florence Quartzite	≤490 m	Massive, thin-bedded, or flaggy grey quartzite, and laminated siltstone, with marine macrofossils.
Austral Creek Siltstone	≤60 m	Thin bedded grey and green-grey siltstone and fine quartzite.
Keel Quartzite	15 m	Cross-bedded coarse grey quartzite with bands of purplish grit.
Amber Slate	≤290 m	Green-grey shale and flaggy siltstone with crinoid ossicles and *Tentaculites* sp.
Crotty Quartzite	~24 m	Massive pale grey quartz sandstone with bands of pebbly grit and quartzose conglomerate.
BASE		(Apparent conformity on Gordon Group correlate)

Devonian deformation

The full extent of the regional Middle Devonian deformation in the Rocky Cape Element is not well established. Williams (1979) attributed late folds of apparent northwesterly trend in the southern part of the Arthur Lineament to the late deformation phase of the mid-Devonian orogenesis. However, based on subsequent detailed regional mapping (Turner *et al.*, 1991), Turner (1992) showed that Devonian structures in the southern Arthur Lineament and adjacent Oonah Formation are steep crenulation cleavages of west-northwesterly to westerly trend, and are statistically parallel to Devonian slaty cleavage in the Wurawina Supergroup of the Duck Creek Syncline adjacent to the west coast at the southern end of the Arthur Lineament. Several late phases of open upright folding and associated crenulation cleavage development recognised in the Oonah Formation between the Heemskirk Granite and the middle Pieman River (Brown *et al.*, 1994), and similar structures in the northern part of the Arthur Lineament (Everard *et al.*, in press), are almost certainly also Devonian in age.

Devonian–Carboniferous granitoids

The oldest dated post-tectonic felsic intrusive in the Rocky Cape Element is an unfoliated quartz-feldspar porphyry intruding the Timbs Group, a unit within the southern part of the Arthur Metamorphic Complex (Turner *et al.*, 1991). This porphyry appears to be significantly older than the major post-tectonic granitoids in the Rocky Cape region. Magmatic zircons from this body yielded a U-Pb age of 380 ± 6 Ma (Turner, 1993*a*; Black, 1994), closely following the Devonian deformation. The porphyry also contains inherited zircon grains with U-Pb ages ranging from ca. 500–1600 Ma, including a cluster of ages at ca. 1200 Ma.

The two major post-tectonic granitoid bodies intruding the Rocky Cape Element are the Early Carboniferous Heemskirk Granite and Pieman Granite. The Heemskirk Granite is a large body intruding the southernmost part of the Rocky Cape Element, and consists of red and white granite phases which are almost contemporaneous (McClenaghan, 1989). A tourmaline nodular facies developed in the white granite on its contact with the red granite suggests the trapping of a fluid-rich phase (Klominsky, 1972). The mechanism of emplacement of the pluton was by intrusion of granite sheets into space created by subsidence within a semi-circular cauldron-like structure (Hajitaheri, 1985). A contact aureole is developed in Cambrian and Silurian to Devonian sedimentary rocks on the southern margin of the body (Brooks and Compston, 1965). Isotopic dating summarised by McClenaghan (1989) shows a range of ca. 330.5–362 Ma. The Pieman Granite, cropping out on the west coast where it intrudes Rocky Cape Group rocks, is petrologically similar to the white phase of the Heemskirk Granite, and shows an isotopic age range of ca. 338.5–356.5 Ma (McClenaghan, 1989).

Granitoid-related mineralisation

Significant mineralisation is associated with the Heemskirk Granite. Iron sulphide-cassiterite replacement deposits, and tin-tungsten skarns, are hosted by carbonate rocks in the Neoproterozoic Oonah Formation, the Success Creek Group, and the Ordovician Gordon Group (see also *Dundas Element*, below). A large number of generally minor Ag-Pb-Zn vein deposits occur over the shallow-shelving eastern subsurface extension of the Heemskirk Granite (Zeehan field), hosted mainly by Oonah Formation and Wurawina Supergroup units (Blissett, 1962; Collins and Williams, 1986). No significant mineralisation appears to be associated with the Pieman Granite, in part because of the steepness of its eastern (onshore) boundary (Leaman and Richardson, 1989).

Near Balfour, Rocky Cape Group correlates host fracture-related, chalcopyrite-bearing, quartz and sulphide lodes with a NNW trend, and tin-bearing quartz veins (Yaxley, 1981). These are also probably Devonian, granite-related deposits (Turner, 1990*d*), although distant from known granitoids.

Dundas Element (DBS, with contributions from K. D. Corbett)

Introduction

Historically, the Dundas Trough was defined as a palaeogeographic feature lying between the Tyennan 'Geanticline' to the east and the Rocky Cape 'Geanticline' (including the Burnie and Oonah Formations) to the west — a Cambrian 'geosynclinal trough' with Success Creek Group at the base on the western side, followed by the Crimson Creek Formation, the Dundas Group and the Mt Read Volcanics, then sequences now known as the Owen Group as the final filling (Campana and King, 1963; Corbett, Green and Williams, 1977).

However, current interpretations of the Late Proterozoic–Cambrian palaeogeography and tectonics are quite different. An early 'trough margin' may be represented by the Burnie and Oonah Formation turbidite wedges (off-lapping from the Rocky Cape Element, see Turner *et al*., 1992, 1994), but there is no clearly defined *eastern* limit for this early basin. There was no 'trough' at the time of deposition of the Success Creek Group and Crimson Creek Formation, but perhaps rather a series of broad basins (or laterally continuous deposition) extending across most of Tasmania (Berry and Crawford, 1988). In the model of Berry and Crawford (1988) the Dundas 'trough' only developed in the Middle Cambrian after their proposed major arc-continent collision event, and rebound of the Tyennan Block. It marked the site of the main Mt Read Volcanics belt and associated volcano-sedimentary sequences, which were largely deposited on top of various allochthonous rock units which had been tectonically emplaced as a result of the collision event. At this time the Mt Read Volcanics/Sticht Range Beds appear to form an eastern 'trough' margin, but the *western* margin is difficult to define. It presumably lies somewhere west of Zeehan, but may be buried under the thrust-faulted margin of the Oonah Formation, as implied by regional cross-sections based on geophysical interpretations (see summary in Leaman *et al*., 1994) and new regional mapping (Brown *et al*., 1994). An equivalent of the western 'trough' margin in the Middle Cambrian may be represented by megabreccias at the western margin of the Dial Range Trough at Penguin on the north coast (Burns, 1964).

A Dundas 'trough' can also perhaps be visualised in the Late Cambrian, when deposition of the thickest parts of the Owen Group and correlates was confined between the Tyennan Element to the east and the Rocky Cape Element to the west.

The view of the palaeogeography thus varies considerably depending on the time-slice taken, and so the usefulness of the Dundas Trough concept has diminished considerably. Consequently, the non-palaeogeographic and non-genetic term Dundas Element (fig. 1) is used herein only as a convenient sub-heading for description of sequences in the area approximating the original Dundas Trough of Campana and King (1963).

Neoproterozoic sequences

An early basin margin may be buried beneath the Mt Read Volcanics belt at the eastern margin of the Dundas Element (Corbett and Turner, 1989), as the oldest exposed post-basement sequence here is the Middle Cambrian Sticht Range beds (Baillie, 1989*c*), which rest unconformably on metamorphic rocks of the Tyennan Element to the east. Alternatively, the Neoproterozoic sedimentary sequences may have extended more or less continuously across the Rocky Cape, Dundas, Tyennan and Jubilee Elements, with the 'Dundas Trough' being essentially a later, Middle Cambrian feature (Berry and Crawford, 1988).

Oonah Formation

In early tectonic models (Campana and King, 1963) the Oonah Formation was considered to form basement to the western margin of the 'Dundas Trough'. However, palaeogeographic reconstructions based on current stratigraphic correlations portray the Oonah and Burnie Formations as integral parts of the early sedimentation in the Dundas Element (Turner *et al*., 1992; 1994). Furthermore, the depositional age of the Oonah Formation, which is constrained by K-Ar ages of 690 ± 10 Ma from slate (Adams *et al*., 1985) and 708 ± 6 Ma from detrital muscovite (Turner, 1993*b*), is significantly younger than the depositional age of ca. 1100 Ma determined for metapelite in the Tyennan region by Råheim and Compston (1977). This suggests the possibility of an older basement (?Rocky Cape Group, Turner *et al*., 1992, 1994) underlying the Oonah Formation and the western part of the Dundas Element. The question of what forms basement to the western part of the Dundas Element is complicated by recent interpretations arising from geophysics (e.g. Leaman *et al*., 1994) and from new regional mapping (Findlay and Brown, 1992; Brown *et al*., 1994), which indicate that substantial parts of the Oonah Formation have been tectonically re-emplaced over and within younger sequences by east- or southeast-directed low-angle thrusting.

The Oonah Formation is more varied than the Burnie Formation with which it is correlated. In a number of areas subdivisions have been recognised which include lithologies such as fine-grained muscovitic quartz sandstone, quartzite, turbiditic quartzwacke, volcaniclastic lithic wacke, pebble conglomerate, laminated siltstone, mudstone, carbonate rocks (commonly recrystallised dolomite) and basaltic lavas (Turner, 1989*a*; Brown, 1986). Chemically the volcanics are high-Ti alkali basalts (Brown, 1986, 1989*b*). Three major inliers of Oonah

Formation correlates occur within the Dundas Element; these are, from north to south, the Mt Bischoff Inlier, the Ramsay River Inlier and the Dundas Inlier. Part of the Dundas Inlier is metamorphosed to phyllite grade (Concert Schist of Blissett, 1962). The Oonah Formation and its correlates in the inliers are commonly structurally complex; for example, four deformation events were recognised in the Mt Bischoff Inlier by Williams (1982). However recent structural studies suggest that the apparent complexity may be due, at least in part, to overprinting of early recumbent and/or isoclinal folds (with associated penetrative foliation) by Middle Devonian upright folds and associated axial-plane crenulation cleavages of at least three separate generations (Brown, 1986; Brown *et al.*, 1994; Everard *et al.*, in press).

Success Creek Group

According to Brown (1986), the Oonah Formation in the western part of the Dundas Element is unconformably overlain by the 750–1000+ m thick Success Creek Group. The only good exposure of the unconformity, in the Pieman River, has since been inundated by a power development, and elsewhere the contact with older rocks is commonly faulted. According to Brown (1986, 1989*b*) the transgressive onlap nature of the unconformity can be inferred from field relationships traced northward from the Pieman River locality, and the unconformity represents a structural and low-grade metamorphic break as well as an hiatus in sedimentation. The interpretation of Brown (1986) is in agreement with Taylor (1954), but contrary to Blissett (1962), who did not recognise any distinction between the Oonah Formation and the Success Creek Group, and considered that there was no evidence for an unconformity between the two sequences. A further complication is that south of the Pieman River, the eastern boundary of the Oonah Formation against younger units immediately west of Zeehan is a major thrust fault (Tenth Legion Thrust: Findlay and Brown, 1992; Brown *et al.*, 1994), and it is unclear what role this structure may play in the Pieman River area. In any case, the existence of a major structural break (equivalent to the Penguin Orogeny) at the Oonah Formation–Success Creek Group contact would represent a contradiction of the regional stratigraphic correlations and radiometric dating in Turner *et al.* (1992, 1994).

The stratigraphy and characteristics of the Success Creek Group are summarised by Brown (1986, 1989*b*). The sequence, with approximate thicknesses, is shown below.

The basal diamictite and the Renison Bell Formation are relatively thin units. Clasts in the basal diamictite are dominantly locally derived, from the Oonah Formation. In the un-named unit above the Dalcoath Formation, soft-sediment deformation is present on all scales from localised slumping to large-scale sliding, resulting in local development of highly deformed melange zones. The deformation is ascribed to incompetent behaviour during post-depositional, localised large-scale slump movements which indicate tectonic instability of the basin (Brown, 1986, 1989). Volcanic detritus in granule and pebble conglomerate units in the Renison Bell Formation is probably derived from volcanic units within the upper Oonah Formation (Brown, 1986, 1989). Oolitic texture and clasts, consisting of fragments of stromatolites (see below), are present in the 'red rock' member (Brown, 1986).

The depositional environment of the Success Creek Group is interpreted as a shallow-water, tidal flat–flood plain setting which was relatively unstable and subsiding at the time (Brown, 1986). The only non-trace fossils recorded, occurring in recrystallised and brecciated oolitic chert units, are fragments of the stromatolite *Baicalia* cf. *B. burra*. These are similar to stromatolite fragments in diamictites associated with the Black River

SUCCESS CREEK GROUP STRATIGRAPHY

TOP		(Conformable contact with Crimson Creek Formation)
Renison Bell Formation	≤150 m	**'Red rock' member**: hematitic chert and mudstone with minor carbonate, lithic wacke and conglomerate units.
		Lower member: Thin-bedded siliceous siltstone interbedded with minor sandstone, calcareous siltstone, laminated mudstone, pebble conglomerate and carbonate units.
Un-named formation	75 m	Laminated mudstone and siltstone, with minor sandstone and conglomerate, showing severe soft-sediment deformation and localised melange development.
Dalcoath Formation	550–800 m	Clean, shallow water quartz sandstone interbedded with minor siltstone, pebbly sandstone and conglomerate.
Un-named formation	50 m	Diamictite (poorly sorted immature polymict conglomerate) with sandstone lenses.
BASE		(Inferred unconformity on Oonah Formation)

Dolomite of the Smithton Synclinorium (Griffin and Preiss, 1976; Brown, 1985; Everard *et al.*, in press).

Crimson Creek Formation

The Success Creek Group is conformably overlain by the Crimson Creek Formation, which consists of 4000–5000 m of interbedded, commonly volcaniclastic turbiditic wacke and siltstone-mudstone, with numerous tholeiitic basalt lava horizons and associated intrusive sills. The sequence shows a ratio of wacke to siltstone-mudstone of at least 60:40 (Brown, 1986, 1989*b*). The sand-grade rocks include hyaloclastite, tuffaceous wacke, volcanic and feldspathic greywackes, and volcaniclastic lithic wacke. Brown (1989*b*) notes that the sequence may not necessarily have been deposited in very deep water.

The mafic volcanic rocks within the Crimson Creek Formation, and those previously correlated with it, have recently been re-evaluated based on major and trace element chemistry, REE patterns and Sm-Nd isotope data (Brown and Jenner, 1988, 1989; Brown, 1989*b*). This work has shown that there are four distinct suites present. Only one of these suites, consisting of sub-alkaline to alkaline basalts with Within Plate Basalt geochemical signatures, is now thought to be primarily associated with the deposition of the Crimson Creek Formation (Brown and Jenner, 1988), and it is believed to correlate with mafic lavas of the Kanunnah Subgroup in the Smithton Synclinorium. The other three suites are believed to be associated with exotic sequences tectonically emplaced into the Dundas Element (see below).

?Allochthonous sequences

Mafic volcanics and associated rocks

Three of the four suites of basaltic rocks distinguished by Brown and Jenner (1988) in the Dundas Element are spatially and genetically related to each other, and have Island Arc–Ocean Island characteristics. The first such suite includes basaltic rocks in the Cleveland-Waratah area north of the Meredith Granite and along the eastern flank of the Huskisson Syncline southeast of the same granite (see Plate 2). This suite is a sub-alkaline basalt association with Ocean Floor Basalt geochemical affinities. The second suite comprises high-magnesium andesitic/boninitic rocks containing distinctive pseudomorphed clinoenstatite phenocrysts and abundant chrome spinel grains, while the third suite comprises low-Ti basalt-andesite with tholeiitic characteristics and extreme LREE depletion. The latter two suites resemble, chemically and isotopically, Eocene–Recent boninitic and associated lavas from the Bonin Islands and Cape Vogel area in Papua New Guinea.

All three basaltic suites, together with associated sedimentary sequences, and the ultramafic-mafic complexes (see below), are considered to be remnants of exotic assemblages tectonically emplaced into the western Tasmanian terrane as a result of a major collision event in the late Early–early Middle Cambrian. The present-day surface remnant of the line of contact of the allochthonous and autochthonous rock assemblages is considered to approximate to a line joining the ultramafic-mafic complexes in the Dundas Element (Brown and Jenner, 1988).

Ultramafic-mafic complexes

Ultramafic-mafic complexes occur in at least ten separate areas in the western half of the Dundas Element (Brown, 1989*b*). They are commonly fault-bounded, and are believed to have been tectonically emplaced (see below). Brown (1986) recognised three ultramafic-mafic rock associations which are commonly in fault juxtaposition within the complexes: Layered Pyroxenite-Dunite (LPD); Layered Dunite-Harzburgite (LDH); and Layered Pyroxenite-Peridotite and associated Gabbro (LPG). Igneous layering is common in all three associations, and pseudo-sedimentary structures have been observed in the LPG succession. Pervasive serpentinisation is common. All of the ultramafic rocks are orthopyroxene-rich, and this feature distinguishes them from the dominantly clinopyroxene-rich sequences which world-wide are usually associated with mid-ocean ridge and back-arc environments (Brown, 1989*b*).

Brown and Jenner (1988) demonstrated genetic links between the ultramafic rocks and the three mafic volcanic suites forming their Island Arc–Ocean Island association (see above). The LDH and LPG ultramafic associations were shown to be high-temperature, low-pressure cumulates formed from the magmas which produced, respectively, the boninitic and low-Ti tholeiitic volcanics. The LPD ultramafic association, which is the oldest based on field relationships, was considered to have formed as a cumulate from the parent magma of the earliest volcanic suite of the Island Arc–Ocean Island association, the Ocean-Floor Basalt suite.

Age constraints on the tectonic emplacement of the complexes are provided by the following:–

- The basal conglomerate units of the Dundas Group contain some detritus of ultramafic derivation (Rubenach, 1974; Padmasiri, 1974; Brown, 1986), indicating that some ultramafic rocks had been tectonically emplaced into the Dundas Element by about the middle Middle Cambrian.

- Recent U-Pb geochronology on igneous zircons from a tonalite in the Heazlewood Ultramafic Complex yielded an age of 510 ± 6 Ma (Turner, 1993*a*; Black, 1994). This is assumed to represent the original age of crystallisation of the last magmatic phase of the rocks in the Complex (i.e. prior to its tectonic emplacement).

In past interpretations, the complexes have been described as disrupted ophiolites (Solomon and Griffiths, 1972; Corbett *et al*., 1972; Rubenach, 1973, 1974). However, after extensive studies, Brown (1986, 1989*b*) believes that no ultramafic complex within Tasmania can be described as an ophiolite or 'ophiolitic', nor can it be inferred that the tectonic environment for their formation was part of a mid-ocean ridge or back-arc setting.

Berry and Crawford (1988) were the first to propose an obduction model, in which the ultramafic-mafic complexes (together with the Island Arc–Ocean Island basaltic suites and associated sedimentary sequences; Brown and Jenner, 1988) are allochthonous relics of a fore-arc terrain which collided with, and was thrust over, a passive continental margin in the Middle Cambrian. The timing of this event is revised to late Early to early Middle Cambrian in the recent synthesis of Berry (1994), on the basis of new and re-interpreted geochronology in Turner *et al*. (1994). The Berry and Crawford (1988) model involves emplacement of all of the complexes as part of a single allochthonous sheet, with a source to the east or north. One problem with such a model is explaining how the Tyennan Element could later rise through the obducted sheet of ultramafic-mafic material without shedding large piles of ultramafic-derived clastic material into the adjacent sedimentary troughs (Brown, 1989*b*; Corbett and Turner, 1989). However, as pointed out by Brown and Jenner (1988), this problem disappears if the Precambrian 'basement' rocks of the Tyennan Element also formed a component of the exotic terranes involved in the collision event.

Osmiridium alloys (Os, Ir, Ru) have been mined from first and second cycle alluvial and eluvial deposits derived from LDH-association rocks in a number of ultramafic complexes in the Dundas Element (but more significantly from similar deposits derived from the Adamsfield body in the Adamsfield–Jubilee Element) (see summary in Green, 1990). Minor amounts of nickel and copper have been mined from the Heazlewood River, Trial Harbour and Serpentine Hill bodies (Collins and Williams, 1986; Brown, 1989*b*), and asbestos from the Cape Sorell and Serpentine Hill complexes.

Post-collisional Middle Cambrian sequences

Introduction

The emplacement of the ultramafic-mafic complexes and associated allochthonous sequences into the Dundas Element was followed by a period of intense volcanism, sedimentation and tectonism extending through the Middle and Late Cambrian. Major Middle Cambrian units developed at this time were the Mt Read Volcanics and associated volcano-sedimentary sequences, and the mixed-derivation sedimentary sequence of the Lower Dundas Group and its various correlatives. Late Cambrian sedimentation was dominated by siliciclastic sequences such as the Owen Group and Upper Dundas Group, which show significant derivation from the metasedimentary rocks of the Tyennan Element.

Mt Read Volcanics and associated rocks

The Mt Read Volcanics form a 10–15 km wide volcanic belt lying along the eastern side of the Dundas Element, and extending in an arcuate shape into the southern margin of the Sheffield Element before disappearing under the younger cover of the Tasmania Basin. The belt is approximately 250 km long, without including possible major extensions beneath the Tasmania Basin. Along the eastern margin of the volcanic belt, the Sticht Range beds form a thin discontinuous basal sequence of clastic sedimentary rocks resting on Precambrian metasedimentary rocks of the Tyennan Element and showing a gradational contact up into the overlying volcanic rocks (Corbett and Turner, 1989). The sequence comprises fluvial to shallow marine conglomerate, sandstone and siltstone, with a probable middle Cambrian biostratigraphic age (Baillie, 1989*c*), but probably does not pre-date all of the Mt Read Volcanics.

The Mt Read Volcanics comprise felsic, intermediate and minor mafic volcanic rocks, dominantly of calc-alkaline type (Corbett and Solomon, 1989). The belt is intruded by high-level, sub-volcanic granitic stocks in several places. In its central part the belt is bisected longitudinally by the major NNE-trending Henty Fault system, across which major lithological and stratigraphic differences are evident in the volcanic sequences. Detailed regional mapping over the last decade has enabled the recognition of several major volcanic and volcano-sedimentary sequences or associations of regional extent, most of which are still informal groupings (see summaries in Corbett and Solomon, 1989; Corbett, 1992). The major units are:–

1. The "Western Volcano-Sedimentary Sequences", comprising extensive successions of interbedded tuffaceous sandstone, siltstone, shale and volcaniclastic mass-flow conglomerate and breccia (including the Yolande River Sequence, Henty Fault Wedge Sequence, White Spur Formation, part of the Dundas Group, and the Mt Charter Group).

2. The "Central Volcanic Complex" (CVC), a relatively proximal volcanic sequence rich in submarine lavas and pumice breccias which interfingers with the Western Sequences and contains a number of major polymetallic orebodies.

3. The "Eastern Quartz-phyric Sequence", which overlies the Sticht Range Beds and probably also interfingers with the CVC.

4. The Tyndall Group and correlates, a younger sequence of mainly volcaniclastic rocks with minor lavas, ignimbrites and limestones.

Recent studies show that the bulk of the Mt Read Volcanics were deposited in a submarine environment, and that many rocks with "pseudo-ignimbritic" textures are pumice-rich mass-flow deposits (Allen and Cas, 1990; McPhie and Allen, 1992; Corbett, 1992). Welded ignimbrite occurrences are restricted to lenses (possibly representing large slide blocks) within the upper part of the Tyndall Group (White *et al.*, 1993).

Age control on the deposition of the Mt Read Volcanics is derived from a number of sources:–

- Recent U-Pb dating of magmatic zircons, in combination with $^{40}Ar/^{39}Ar$ dating of magmatic hornblende, from the Mt Read Volcanics yielded a concordant age of 502.6 ± 3.5 Ma (Perkins and Walshe, 1993). A volcaniclastic unit in the Tyndall Group yielded a slightly younger, $^{206}Pb/^{238}U$ age of 494.4 ± 3.8 Ma. It should be noted that the age of the Central Volcanic Complex in particular was not extensively tested by this dating.

- A single trilobite recovered from the Sticht Range beds indicates a probable Middle Cambrian age (Baillie, 1989*c*).

- The Que River Shale of the Mt Charter Group contains a trilobite-rich fauna of Boomerangian (middle Middle Cambrian) age (Jago, 1977; 1979).

- A limestone in the basal part of the Tyndall Group contains late Middle Cambrian fossils (Jago *et al.*, 1972).

- The Owen Group, which unconformably overlies the Mt Read Volcanics, includes the Newton Creek Sandstone which contains middle Late Cambrian fossils of post-Idamean, pre-Payntonian age (Jago *in* Corbett, 1975*b*).

Perkins and Walshe (1993) also found that the Mt Read Volcanics contain inherited zircons, falling into two age groups. The oldest group comprises sub-rounded abraded crystals in the age range 800–1600 Ma, and a source in the metamorphic rocks of the Tyennan Element (which underlie at least part of the Mt Read Volcanics) is suggested. The younger group comprises euhedral grains in the range 530–600 Ma, which the authors suggest may be derived from early sequences within the Dundas Element. However the Time-Space diagram (Plate 1) suggests that the only autochthonous sequence in this age range in the Dundas Element is the Crimson Creek Formation.

Mineralisation in the Mt Read Volcanics

The Mt Read Volcanics form the most conspicuously mineralised belt of rocks in Tasmania, being host to several world-class base-metal and precious-metal deposits (Hellyer, Que River, Rosebery, Hercules, Mt Lyell, Henty) and numerous medium-sized and small deposits. All of the major deposits have many of the characteristics of volcanic-hosted massive sulphide (VHMS) deposits, believed to have formed on the seafloor during volcanism (Corbett and Solomon, 1989). The Mt Lyell deposit is unusual in having most of its metal reserves in disseminated, epigenetic, replacement-type ore probably formed beneath the seafloor. Interpretation of the Rosebery orebody remains controversial, with recent models invoking contrasting origins, by Cambrian synvolcanic sub-seafloor replacement of permeable pumiceous mass-flow deposits (Allen, 1994), or by Devonian syntectonic metal mobilisation and wallrock replacement in structural traps (Aerden, 1994).

Recent interpretations are consistent with the major VHMS deposits forming within one or two relatively short time intervals (Corbett, 1994). The Rosebery-Hercules ore horizon, hosted by the northern CVC, may be time-equivalent, or nearly so, to the Que-Hellyer horizon within the Mt Charter Group (McPhie and Allen, 1992; Perkins and Walshe, 1993; Corbett, 1994). The large disseminated Cu-Au orebodies of the Mt Lyell field are genetically related to nearby small massive sulphide lenses of seafloor-exhalative origin such as Comstock (Walshe and Solomon, 1981). The latter lie at or close to the contact between the CVC and overlying Tyndall Group. Mineralised horizons further north, such as Howards Anomaly, occupy a similar stratigraphic position (Corbett, 1994). The Mt Lyell mineralisation may be younger than the Que-Hellyer horizon as the basal Tyndall Group is late middle Cambrian, while the Que-Hellyer horizon pre-dates the Que River Shale which contains a middle Middle Cambrian fauna.

Cu-Au mineralisation south of the Henty Fault may be genetically related to the intrusion of Cambrian granites (Large *et al.*, 1994) which intrude the Eastern Quartz-Phyric Sequence and the CVC but pre-date the Tyndall Group (Corbett, 1979; Perkins and Walshe, 1993).

Lower Dundas Group

The Dundas Group is a thick conglomeratic flysch sequence, which can be considered to comprise two parts: a lower sequence of Middle Cambrian age containing detritus from various sources including felsic volcanic rocks (probably Mt Read Volcanics), ultramafic-mafic rocks and intra-basinal cherts; and an upper sequence of essentially Late Cambrian age in which the detritus is predominantly siliceous and derived from Precambrian metasedimentary rocks. The latter sequence is equivalent to the Owen Group.

The amount of felsic volcanic detritus in the Lower Dundas Group in its type area is relatively small according to Brown (1986), being most prominent in the Razorback Conglomerate and in the Brewery Junction Formation at the top of the sequence. Time-equivalent rocks along strike, however, such as the Huskisson Group just to the north, are predominantly of felsic volcanic (Mt Read Volcanics) derivation (Brown, 1986), and it seems likely that the Lower Dundas Group in its type area represents deposition in a local basin dominated by input from local sources.

The base of the Dundas Group is difficult to define. Contacts against the ultramafic-mafic complexes and Crimson Creek Formation in the Dundas area are typically faulted, although the presence of ultramafic detritus in the Red Lead Conglomerate suggests that this unit may have been close to the base, at least locally. Detailed re-mapping led Selley (1994) to conclude that the Red Lead Conglomerate lies with erosional contact upon a succession of intercalated basaltic pillow lavas and volcanic breccias which forms part of the Serpentine Hill Complex, one of the ?allochthonous ultramafic-mafic complexes (Berry and Crawford, 1988). In places the Red Lead Formation is underlain by a siltstone-sandstone sequence, referred to as the Judith Formation, which contains the oldest Middle Cambrian fauna (*P. gibbus* Zone) known in Tasmania.

Corbett and Lees (1987) suggested that the basal part of the Dundas Group is represented by the White Spur Formation, a west-facing volcano-sedimentary sequence linking the Dundas area to the Mt Read Volcanics. They interpreted the contact between the White Spur Formation and underlying Central Volcanic Complex rocks as an unconformity, but later work by Allen suggests an interfingering relationship is more likely, with the base of the White Spur Formation being represented by the 'hangingwall epiclastics' at the Hercules and Rosebery mines (Allen and Cas, 1990; McPhie and Allen, 1992).

Probable equivalents of the Lower Dundas Group occur extensively in the Cuni area southwest of Renison and in the Stonehenge area west of Zeehan, and include felsic to intermediate volcaniclastic sandstone, siltstone and chert-rich granule-pebble conglomerate (Brown *et al*., 1994).

Late Cambrian – Early Ordovician siliciclastic sequences

Introduction

Along the West Coast Range in the eastern part of the Dundas Element the Mt Read Volcanics are overlain, typically unconformably, by the Owen Group, a thick sequence of Tyennan-derived siliciclastic conglomerate and sandstone of Late Cambrian to Early Ordovician age (Corbett, 1990). Facies within the sequence range from possibly non-marine (alluvial fan), through shallow marine (typically with abundant marine trace fossils), to deeper marine proximal turbidites probably deposited in submarine fans.

The siliciclastic sequences are also present in the central parts of the Dundas Element, where they tend to be more marine in character, with a higher proportion of flysch-type deposits. They include the Upper Dundas Group, upper Huskisson Group, and extensive sequences in the Hatfield River area. Conglomerate-rich sequences of more proximal type are again evident in the Mt Zeehan–Professor Range area, possibly indicating the presence near here of a western 'trough' margin in the Late Cambrian.

Owen Group (West Coast Range)

The Owen Group siliciclastic sequence has recently been mapped, at 1:25 000 scale, from the West Coast Range to the Black Bluff area. The sequence has a maximum thickness of about 1.5 km on the Tyndall Range, against the Great Lyell Fault, but shows marked thinning to the east away from the fault (Corbett, 1990). Five main units are recognised, based on a type sequence in the Queenstown area (Wade and Solomon, 1958) (see over).

At the base of the Owen Group, the Jukes Conglomerate consists of discontinuous lenses of volcaniclastic conglomerate, sandstone and breccia, commonly of local derivation and indicating rapid erosion of the volcanic pile.

Above the Jukes Conglomerate, the Owen Group consists predominantly of siliciclastic conglomerate and sandstone in which the dominant clast lithologies are quartzite and quartz schist derived from the Precambrian metasedimentary rocks of the Tyennan Element. Chert clasts, presumably of intra-basinal origin and probably derived from the allochthonous sequences, are a significant component in some units. The two major conglomerate units (Lower Owen and Middle Owen Conglomerate) are typically thick-bedded to massive, and up to boulder grade. Trace fossils and evidence for marine conditions are lacking, and the units possibly represent alluvial fan deposits. Intra-formational unconformities are present within the Lower Owen Conglomerate. The Newton Creek Sandstone is a proximal turbidite sequence of interbedded sandstone and siltstone, with lesser conglomerate and minor limestone. It contains trilobites and brachiopods of middle Late Cambrian age. The Upper Owen Sandstone is a shallow marine sequence, which in the Queenstown area is divided into two parts by the Haulage Unconformity. This angular discordance results from folding of the lower beds, and is attributed to movements on the Great Lyell Fault. Similar movements were probably responsible for other unconformities within the Owen Group, and major control on Owen Group sedimentation by the Great Lyell Fault can be inferred from the dramatic thickening of units which occurs towards the fault (Corbett, 1990).

The uppermost Owen Group unit, the Pioneer Beds, contains abundant chert detritus as well as detrital chromite in many places, suggesting re-activation of intra-basinal ultramafic and chert sources by the Haulage Movement. The Pioneer Beds transgress westwards across the Great Lyell Fault at Queenstown, whereas all previous Owen units were apparently ponded against the fault (Corbett *et al*., 1974). Field relationships in the Queenstown area suggest that the Pioneer Beds have a gradational contact into overlying Gordon Group limestone via a thin sequence of calcareous siltstone and sandstone (Calver *et al*., 1987; Corbett *et al*., 1989). This is consistent with the early Caradoc age of the earliest faunas recorded in the Gordon Group in the Dundas Element (faunal assemblage OT12; Banks and Baillie, 1989), and with preliminary assessment of marine macrofossils from the Pioneer Beds, which has indicated a middle Ordovician age (pers. comm. J. R. Laurie, AGSO). The new data on the age of the Pioneer Beds (which was previously believed to be lowermost Ordovician) may imply that a considerable hiatus is represented by the Haulage Unconformity (see also discussion in Banks and Baillie, 1989, p.199).

Upper Dundas Group (Central Dundas Element)

The Upper Dundas Group, from the upper part of the Brewery Junction Formation through to the base of the Gordon Group, is a correlate of the Owen Group. The lower part of the sequence, up to the Misery Conglomerate, is marine and flysch-like, with sandstone-siltstone units alternating with units rich in pebble to boulder-grade conglomerate (Brown, 1986). Clasts in the conglomerate consist predominantly of quartzite, with lesser chert. The Misery Conglomerate is about 160 m thick, and is possibly a correlate of the Middle Owen Conglomerate (Corbett, 1990). It overlies the fossiliferous Climie Formation, which is of pre-Payntonian age and appears to be a correlate of the Newton Creek Sandstone, and is overlain (possibly disconformably) by a shallow marine sequence of sandstone and siltstone which contains fossils of latest Late Cambrian age (Jago and Corbett, 1990). The latter unit is possibly a correlate of the Linda Sandstone.

Professor Range – Mt Zeehan area

A siliciclastic sandstone-siltstone-conglomerate sequence of Late Cambrian age is also present in the Professor Range–Henty River area. The lower part of the sequence is flysch-like in character, with abundant green to grey siltstone, and contains trilobites of probable Post-Idamean age suggesting correlation with the Newton Creek Sandstone and Climie Formation (Baillie and Corbett, 1985). Overlying this, disconformably in places, is a unit of pebble-cobble conglomerate and sandstone, up to 300 m thick, showing some cross-bedding and indications of shallow-water deposition. This is followed by a thick (1000 m+) upper sequence of well-sorted siliceous sandstone with some conglomerate beds, cross-bedding, ripple marks and intensely bioturbated beds with worm-burrows perpendicular to bedding. The latter unit has been correlated with the Moina Sandstone of the Sheffield Element (Blissett, 1962; Brown *et al*., 1994) and a littoral to immediately sub-littoral depositional environment has been suggested (Banks and Baillie, 1989). However it is probably more accurately correlated with the Upper Owen Sandstone, of which the Moina Sandstone represents only the uppermost part in the Black Bluff area at the southwestern margin of the Sheffield Element (Corbett, 1990).

Thick accumulations of siliciclastic pebble to cobble conglomerate at Mt Zeehan (Mt Zeehan Conglomerate) are almost certainly predominantly

OWEN GROUP STRATIGRAPHY (QUEENSTOWN AREA)

TOP	(Disconformably overlain by Gordon Group)
Upper Owen Sandstone	**Pioneer Beds:** Chert-rich granule-pebble conglomerate, with detrital chromite in places, passing up into grey sandstone with pipestem burrows and minor shale.
	~~HAULAGE UNCONFORMITY~~~~~~~~~~~~~
	Linda Sandstone: Shallow marine sequence of thin-bedded, pink to grey, bioturbated sandstone and siltstone.
Middle Owen Conglomerate	Thick-bedded to massive pebble-cobble to cobble-pebble conglomerate with thin lenses of pink sandstone.
Newton Creek Sandstone	Proximal turbidite sequence of grey sandstone, siltstone and lesser conglomerate, with Late Cambrian fossils.
Lower Owen Conglomerate	Two units of boulder-grade conglomerate (one with an erosional unconformity at the base), and a thin unit of siltstone and sandstone.
Jukes Conglomerate	Locally-derived volcaniclastic conglomerate.
BASE	(Erosional unconformity on Mt Read Volcanics)

alluvial fan deposits, probably deposited in a separate small basin distinct from the main areas of deposition of the Owen Group in the West Coast Range area. Palaeocurrent and provenance data indicate that conglomerate detritus at Mt Zeehan was derived from the northwest (provenance in the Rocky Cape Element?), and is of different character to that in the siliciclastic conglomerate at the Professor Range which was derived from a northeasterly to easterly direction (D. B. Seymour, unpublished data). The conglomerate sequence at Mt Zeehan is overlain by a cross-bedded and bioturbated sandstone unit similar to that forming the upper sequence at the Professor Range (Blissett, 1962; Brown *et al.*, 1994).

Cambrian deformation

One of the main problems encountered in deciphering the tectonic history of the Dundas Element is separating the strong regional Devonian structural overprint from earlier structures, particularly as the stratigraphic column contains much direct and indirect evidence of Cambrian-age tectonism. Typically the Cambrian rocks show up to three phases of cleavage development, but commonly these are parallel to cleavages in adjacent post-Cambrian rocks, and have thus been attributed to Devonian deformation. Unequivocal evidence for the regional development of a penetrative cleavage in the Cambrian has not been found, although rare local Cambrian cleavage development has been inferred in the Mt Read Volcanic belt (Corbett and Turner, 1989). However, development of open upright N-trending folds (which were later tightened during Devonian orogenesis) in the Dundas Element is now believed to have occurred in the Late Cambrian as part of the last phase of the Delamerian Orogeny (510–490 Ma) (Berry, 1994).

Kinematic studies aimed at determining the movement histories of major faults have also had to contend with the problem of Cambrian versus Devonian movements. The Great Lyell Fault has a complex movement history including a number of major phases of movement in the Cambrian, yet local foliation fabrics in the fault zone are commonly dominated by the youngest Devonian overprint. Greater understanding of pre-Devonian movement history was achieved in a study of the Henty Fault (Berry, 1989), where computer analysis of fault striations revealed three major phases of movement:–

- Early east-directed thrusting which created a ramp through the Owen Group. This movement pre-dated Devonian folding and may correlate with Early Ordovician movement on the Great Lyell Fault.

- A second reverse movement (on a steep west-dipping reverse fault) which post-dated NW-trending Devonian folds.

- A final Devonian movement involving a sinistral wrench displacement of less than 5 km.

Gordon and Eldon Groups

In the Dundas Element, the Pioneer Beds and the Moina Sandstone are succeeded by a sequence up to at least 600 m thick, of predominantly limestone with subordinate siltstone and minor sandstone, which comprises the Gordon Group (Banks and Baillie, 1989). There may be a depositional hiatus (or condensed section) between the Pioneer Beds and the Gordon Group. The Gordon Group was deposited over an extensive area of western Tasmania, including the Dundas and Sheffield Elements, the Florentine Synclinorium area of the Adamsfield–Jubilee Element, and at least part of the Tyennan Element. The most common lithologies are micrite and dolomitic micrite, deposited in supratidal, intertidal, and subtidal shallow marine environments. Common fossiliferous horizons constrain the span of deposition of the succession, which based on the preliminary biostratigraphic system for the Ordovician of Tasmania set up by Banks and Burrett (1980), ranges between faunal assemblages OT12 (early Caradoc) and OT20 (mid-Ashgill) in the Dundas Element (Banks and Baillie, 1989). The base of the carbonate sequence is regionally diachronous, deposition apparently having started later in the Dundas Element than in the Sheffield Element (Banks and Baillie, 1989).

The Gordon Group carbonates are host to Pb-Zn-Ag mineralisation, the greatest concentration being in the Zeehan mineral field. The most significant deposit, at the Oceana mine south of Zeehan, was thought to have originated by deposition in structurally controlled fissure veins from mineralising fluids emanating from the Devonian Heemskirk Granite, but is now regarded as an Ordovician sedimentary exhalative deposit (S. Taylor in Banks and Baillie, 1989; Taylor and Mathison, 1990), an interpretation supported by recent Pb isotope determinations on galena. Two types of mineralisation occur at Oceana: one, grossly discordant to the host rock, consisting mostly of generally coarse galena, sphalerite and siderite occurring as open-space infillings of veins, cavities and intraclastic areas of breccias in a recrystallised, silicified and dolomitised limestone; the other comprising two separate stratiform horizons of semi-massive fine-grained galena, sphalerite and siderite, with replacement-indicative textures, lying at the top and bottom of a 30 m thick unit of limestone breccia at a stratigraphic level about halfway through the Gordon Group. The distinctive limestone breccias are now thought to be of submarine debris-flow origin during deposition of the Gordon Group.

The Gordon Group is succeeded by the predominantly clastic succession of the Eldon Group, which in the Dundas Element is probably disconformable in some places on the Gordon Group (Banks and Baillie, 1989). The Eldon Group

ELDON GROUP STRATIGRAPHY (ZEEHAN AREA)

TOP		(Not exposed)
Bell Shale	>425 m	**Upper**: Dominantly mudstone, with marine macrofossils. **Lower**: Thin-bedded, very fine-grained quartz sandstone, siltstone and mudstone; marine macrofossils.
Florence Quartzite	490 m	Massive to plane-laminated fine-grained quartz sandstone with minor mudstone; marine macrofossils.
Austral Creek Siltstone	60 m	Mudstone.
Keel Quartzite	60 m	Cross-bedded, ripple-marked fine-grained quartz sandstone; rare tentaculitids.
Amber Slate	245 m	Interbedded mudstone, siltstone and fine-grained sandstone; marine macrofossils and ostracodes.
Crotty Quartzite	490 m	Quartz sandstone, conglomerate and mudstone; marine macrofossils and worm casts.
BASE		(Disconformable in some places on Gordon Group)

sequence consists predominantly of major units of quartz sandstone and siltstone with minor conglomerate horizons and limestone lenses. Horizons with marine macrofossils are common, and biostratigraphic constraints on the span of deposition in the Dundas Element range between mid-Llandovery (Aeronian) and Pragian (Banks and Baillie, 1989). A lack of fossil occurrences in the Ludlow and early Pridoli suggests a period of non-deposition in this time range. Commonly the Eldon Group shows an easily recognisable stratigraphy, which in the type area near Zeehan (Blissett, 1962; Banks and Baillie, 1989) is as shown above.

A number of depositional environments have been suggested for various parts of the sequence, including the following (Banks and Baillie, 1989): very shallow marine, sandy tidal flat (Crotty Quartzite and Amber Slate); storm-influenced barrier bar complex (Florence Quartzite); storm deposits formed by fluctuating storm-wave action and waning events (parts of the lower unit of the Bell Shale); and deeper water marine (upper unit of the Bell Shale).

Devonian deformation

In the late Early to early Middle Devonian all of the rock sequences of the Dundas Element, up to and including the Eldon Group, were widely affected by at least two regional phases of upright to inclined, open to tight folding and associated cleavage development and faulting. These two events formed structures of two trend groups, respectively N-S (earlier), and NW-SE to WNW-ESE (later), and are believed to represent the regional D_3 and D_4 events of an orogeny of late Early to early Middle Devonian age in western Tasmania (Williams *et al*., 1989). Folds belonging to the later NW-SE trend also affected significant areas of the Precambrian rocks of the Rocky Cape and Tyennan Elements bordering the Dundas Element (Williams, 1976, 1979). The northern part of the Dundas Element was also affected by structures of character similar to the other two trend groups but of NE-SW trend, which are believed to represent the regional D_2 event of the same orogeny (Seymour, 1980; P. R. Williams *in* Seymour, 1989). The Devonian orogenesis in western Tasmania has commonly been correlated with the Tabberabberan Orogeny of eastern mainland Australia (Williams, 1979).

Devonian–Carboniferous intrusive rocks

The largest outcropping post-tectonic granitoid body within the Dundas Element is the Meredith Batholith. This is a large composite body composed predominantly of two textural types; equigranular, fine-grained to medium-grained grey biotite adamellite, and porphyritic biotite adamellite. Recent work has resulted in the recognition of ten separate plutons within the batholith, grouped into mafic type and felsic type based on a number of characteristics (A. Camacho *in* Williams *et al*., 1989). The contact aureole of the batholith is up to 2.5 km wide and contains hornfels of albite-epidote, hornblende, and locally of pyroxene hornfels facies (Groves *et al*., 1973). The maximum range (including uncertainty limits) of Rb-Sr and K-Ar radiometric ages from the batholith is ca. 338.5–366 Ma (Williams *et al*., 1989).

As well as the outcropping granitoids, geophysical studies have indicated the presence, at depth, of a very large ENE-trending ridge of granitoid almost linking the outcropping Heemskirk and Granite Tor Granites (Leaman and Richardson, 1989). Indicated depths to granite along this structure are as shallow as 1 km and less just south of a line between the Renison and Rosebery mines, and the structure underlies much of the Zeehan, Renison-Dundas and Rosebery mineral fields. A Department of Mines drill hole at Colebrook Hill, near Rosebery, penetrated this granite at a depth of one kilometre (K. D. Corbett, pers. comm.).

Probably related to the Heemskirk–Granite Tor subsurface ridge is the Renison Complex, a composite granitoid stock which has been intersected below tin-mineralised rocks in the Renison mine (A. Camacho in Williams *et al*., 1989). The Complex intrudes both the Success Creek Group and the Crimson Creek Formation, and three mineralogical types have been distinguished by Camacho (*in* Williams *et al*., 1989). The granitoid has been related to the strongly greisenised Pine Hill stock, which crops out two kilometres south of the mine (Patterson *et al*., 1981).

In the vicinity of the inlier of Oonah Formation rocks at Mt Bischoff in the northern Dundas Element, radiating quartz porphyry dykes intrude both the Oonah Formation correlate and the overlying early basin sequences. It has been suggested that these dykes emanated from the cupola of an underlying granitoid body (Groves and Solomon, 1964; Williams *et al*., 1989). Extreme greisenisation of the porphyries has resulted in the formation of topaz, tourmaline, muscovite and cassiterite pseudomorphing primary feldspar.

Post-tectonic lamprophyre dykes occur in the southern part of the Dundas Element, and one of these at Queenstown has yielded a Late Devonian K-Ar radiometric age of 363 ± 3 Ma (Baillie and Sutherland, 1992).

Granitoid-related mineralisation

A number of important deposits are associated with the subsurface ridge of granite between the Heemskirk and Granite Tor plutons. Most significant are the large stratabound iron sulphide-cassiterite replacement deposits hosted by the Success Creek Group and Crimson Creek Formation at the Renison mine, Australia's largest primary tin producer (Morland, 1990). Concentrations of vein deposits are found where the granite ridge intersects major faults (Bamford and Green, 1986), for example the Mt Farrell Pb-Zn deposits along the Henty Fault (Collins and Williams, 1986).

Important iron sulphide-cassiterite replacement deposits lie above the gently sloping northern flank of the Meredith Granite at Mt Bischoff (Wright, 1990) and Cleveland (Collins and Williams, 1986; Leaman and Richardson, 1989). The former is hosted in Oonah Formation; the latter in a probably-allochthonous sedimentary succession associated with mafic volcanic rocks of Island Arc–Ocean Island affinities (Brown and Jenner, 1988). Silver-lead-zinc vein deposits, the largest being at Magnet, have also been sourced from the Meredith Granite.

Southern extensions of the Dundas Element

South of Macquarie Harbour (see Plate 2), the triangular region bounded by Cape Sorell, Point Hibbs, the D'Aguilar Range and Elliott Bay has not been as intensively studied as the core part of the Dundas Element. However, it is considered to consist largely of units equivalent to sequences which have been defined or recognised within the Dundas Element north of Macquarie Harbour. Because of these similarities and due to space restrictions, the region has not been specifically illustrated on the Time-Space Diagram. As well as the many similarities with the core part of the Dundas Element there are some important differences in the geology, some of which are summarised in this section.

The structural framework of the region comprises a series of north- to northeast-trending lithological associations, comparable to the major subdivisions in the core part of the Dundas Element, but which here are mostly in mutual fault contact. The major bounding faults range from steeply dipping to very shallowly dipping, and some are known or inferred major thrust faults.

?Mesoproterozoic units

In the northwestern corner of the region, the Cape Sorell Block is in contact with Neoproterozoic units to the southeast on a shallowly northwest-dipping thrust fault (McClenaghan and Findlay, 1989). The Cape Sorell Block comprises a structurally complex lower greenschist facies assemblage of interlayered orthoquartzite, micaceous quartzite, phyllite and minor siliceous conglomerate (Baillie and Corbett, 1985). The orthoquartzite commonly shows cross-bedding and ripple marks. Herringbone cross-bedding suggests shallow-water deposition under tidal conditions. Four major deformation events have been recognised, the earliest of which produced large-scale isoclinal folds with axial plane penetrative foliation, and common transposition of bedding into the plane of the foliation. Microstructures and mineral assemblages associated with the axial plane foliation are typical of syntectonic lower greenschist facies metamorphic conditions (Baillie and Corbett, 1985).

The lithofacies, structural complexity and metamorphic grade of the Cape Sorell Block suggest correlation with the metasedimentary rocks of the Tyennan Element, which have an inferred Mesoproterozoic age of original deposition (Råheim and Compston, 1977).

The inferred northwest-dipping geometry of the bounding fault of the Cape Sorell Block has led to the interpretation that the block was emplaced by southeast-directed thrusting (e.g. Leaman *et al*., 1994). However, an outlier of similar rocks to the southeast of the main block appears to be a small klippe resting above an almost horizontal fault surface on Neoproterozoic sequences (see Plate 2, and McClenaghan and Findlay, 1989). This suggests that the overall geometry could also have arisen from northwest-directed thrust emplacement involving a ramp geometry, as suggested in Brown *et al*. (1991). Such an interpretation may more easily accommodate the suggested Tyennan correlation of the Cape Sorell Block.

?Neoproterozoic sequences

Two northeast-trending belts of rocks of inferred Neoproterozoic age separate the Cape Sorell Block from a major NNE-trending fault zone involving ultramafic and other rock types, extending from Hibbs Bay to Macquarie Harbour (see Plate 2). The two belts of ?Neoproterozoic rocks are in mutual fault contact.

The northwestern belt of ?Neoproterozoic rocks contains numerous internal major fault contacts, and includes three lithological associations:–

- Interbedded mudstone and lithicwacke, with major units of basaltic lava (including pillow lavas) known as the Lucas Creek Volcanics (McClenaghan and Findlay, 1989).
- Variably calcareous, interbedded mudstone, siltstone and sandstone, with minor conglomerate.
- Crystalline dolomite with minor chert and mudstone.

The first association is correlated with the Crimson Creek Formation (central Dundas Element) and the Kanunnah Subgroup (Smithton Synclinorium), based on similarities in overall lithology, and in major, trace and rare earth element geochemistry of the basaltic lavas (McClenaghan and Findlay, 1993). The lithological characteristics of the other two assemblages suggest correlation with the Success Creek Group (Dundas Element) and the bottom two formations of the Togari Group (Smithton Synclinorium).

The southeastern belt of ?Neoproterozoic rocks comprises two lithological assemblages, namely metamorphosed interbedded quartzwacke and mudstone/siltstone (dominant), and metamorphosed impure dolomite-rich sequences of mudstone, siltstone and sandstone (McClenaghan and Findlay, 1989). Graded bedding, flute casts and load casts in the dominant assemblage indicate turbidity current deposition (McClenaghan and Findlay, 1993). Lithological similarities indicate correlation with the Oonah Formation in the main part of the Dundas Element.

?Allochthonous units

As in the core part of the Dundas Element, the Cape Sorell–Elliott Bay region contains fault-bounded ultramafic assemblages, which are believed to represent allochthonous units which were tectonically emplaced in the late Early to early Middle Cambrian (Berry and Crawford, 1988). These occur in a relatively narrow zone of complex major faulting and shearing, extending from Macquarie Harbour to the west coast at Hibbs Bay, and which has been called the Point Hibbs Melange Belt (McClenaghan and Findlay, 1993). This shear zone includes slivers of fossiliferous Late Cambrian and Ordovician rocks, and its apparent southern extension at Hibbs Bay includes thrust slices of mid-Early Devonian limestone (Carey and Berry, 1988; Seymour *in* Brown *et al*., 1991). The zone has probably therefore had a complex history of repeated movements, the youngest being no older than Middle Devonian.

Apart from the ultramafic rocks, the region also contains at least one other possibly allochthonous rock association. This is a largely fault-bounded basaltic volcanic association occurring in two areas, one adjacent to the southwestern shore of Birchs Inlet, the other in a belt trending north from near Veridian Point on the west coast until it becomes concealed by Tertiary cover south of the Wanderer River (see Plate 2). These have been collectively known as the Birchs Inlet–Mainwaring River Volcanics (Brown *et al*., 1991). The association contains vesicular, pillow and sheet flows with interbedded sedimentary rocks which increase in proportion southwards. The sedimentary rocks include associations of interbedded mudstone and chert, interbedded volcaniclastic siltstone and lithicwacke with minor mudstone and carbonate, siliceous pebbly quartzwacke, and lithicwacke with minor siliceous pebble conglomerate (Brown *et al*., 1991). Petrography of the clastic rocks indicates a mixed volcanic and low-grade metamorphic source. Two geochemical groups of lavas are present, a lower picritic basalt association, and an upper tholeiitic basalt association, and these are believed to represent successive batch melts of the same source. The geochemistry also indicates the presence of units similar to island-arc tholeiites, as well as others similar to intra-plate basalts (Brown *et al*., 1991).

Middle Cambrian units

Between the D'Aguilar Range and Elliott Bay, bounded to the west by a fault against the Mainwaring River Volcanics and to the east by a mostly faulted boundary against Tyennan metasedimentary rocks, is a north-trending belt of mostly felsic volcanic, volcaniclastic, epiclastic and intrusive rocks which represents the southward extension of the Mt Read Volcanic belt (Pemberton, Vicary, Bradbury and Corbett, 1991; Vicary *et al*., 1992; Bradbury *et al*., 1992). At the eastern margin of the central part of the belt, a correlate of the Sticht Range Beds rests with apparent unconformity on metasedimentary rocks of the Tyennan Element (Vicary *et al*., 1992). The belt is intruded by

Cambrian granites at Low Rocky Point and on the eastern side of Elliott Bay.

Also possibly related to the Mt Read Volcanics are the Noddy Creek Volcanics and correlates, which extend in a somewhat discontinuous series of outcrops, from a fault-bounded area adjacent to the Point Hibbs Melange Zone near the Macquarie Harbour coast, southward to the west coast near High Rocky Point (see Plate 1). The larger outcrop areas in the northern part of this belt consist largely of pyroxene and feldspar-phyric andesitic intrusive rocks, lavas and ?autobreccias (McClenaghan and Findlay, 1989; Brown *et al*., 1991), but suggested correlates of the Noddy Creek Volcanics in the High Rocky Point area consist of basaltic andesites including pillow, sheet and autobreccia flows and agglomerate, with interbedded clastic sedimentary rocks (Brown, 1988). The Noddy Creek Volcanics are calc-alkaline and have the geochemical characteristics of island-arc volcanics (Brown *et al*., 1991).

In a broad north-south belt between the Point Hibbs Melange Zone and the western fault boundary of the Birchs Inlet–Mainwaring River Volcanics (see Plate 2), a dominantly clastic sedimentary sequence apparently partially encloses the Noddy Creek Volcanics, and consists of a variety of rock types including polymict and volcaniclastic to epiclastic conglomerate, sandstone, siltstone and mudstone. The sequence is commonly east-facing, and may correlate with the Lower Dundas Group (Middle Cambrian) in the core part of the Dundas Element.

Wurawina Supergroup

Siliciclastic conglomerate and sandstone, correlated with the Owen Group of the West Coast Range, occur in faulted north-trending synclines at the D'Aguilar Range and in and south of the upper reaches of the Wanderer River, and in a northwest-trending syncline southwest of Birchs Inlet (see Plate 2). The latter structure also has Gordon Group carbonates in its core.

Correlates of the lower Wurawina Supergroup and Upper Dundas Group also form a major fault sliver on the western side of the Point Hibbs Melange Zone (McClenaghan and Findlay, 1989). At its southern end this major shear zone appears, in part, to fan out into a southeast-dipping thrust stack involving sheets of ?Middle Cambrian clastic sedimentary rocks, Owen Group and Gordon Group correlates, and Devonian carbonates and clastic sedimentary rocks (Carey and Berry, 1988; Brown *et al*., 1991; McClenaghan *et al*., 1994). Blocks in this thrust stack are commonly in reversed stratigraphic order. The latest thrust movements in the development of the stack post-dated the Lower Devonian (Pragian) Point Hibbs Formation (Carey and Berry, 1988) and are attributed to the regional early Middle Devonian orogenesis. However, a structural interpretation based on the most recent detailed mapping in the area (Seymour *in* Brown *et al*., 1991) suggests that the stack of thrust sheets is folded by later north-trending folds, which are probably also attributable to the Middle Devonian deformation.

Devonian intrusive rocks

A few isolated outcrops of adamellite occur on the west coast south of Cape Sorell (Baillie *et al*., 1977). Gravity and seismic interpretation by Leaman and Richardson (1989) indicated that the small outcrops are parts of a large granitoid mass extending at shallow depth some considerable distance offshore (see 2 km contour on Plate 2). The latter authors named the body the Grandfathers Granite, and argued a Devonian age for it on structural and rock property evidence.

Post-tectonic dykes and concordant sheets of lamprophyre occur in a number of structural settings in the Cape Sorell–Elliott Bay region, including the Point Hibbs Melange Zone and the thrust stack at Hibbs Bay. Radiometric age determinations on a number of samples have yielded K-Ar ages with a weighted average of 373.4 ± 4.1 Ma (McClenaghan *et al*., 1994), or late Middle Devonian on the time-scale used herein. At Hibbs Bay this dating constrains the lower limit on the age of the regional Middle Devonian orogenesis to a degree at least equal to the best previous constraint, provided by undeformed spelean deposits of the Eugenana Beds in the Sheffield Element.

Sheffield Element (DBS)

Introduction

For the purpose of this compilation, the term Sheffield Element refers to the large rectangular area of Neoproterozoic–Early Palaeozoic basin sequences in the central north of the state, bounded to the west by the northern part of the Arthur Lineament, to the south by the Tyennan Element, and to the east by the River Tamar, but also overlapped by Tasmania Basin and younger sequences in the east, southeast and west (fig. 1)). A somewhat arbitrary boundary is drawn with the Dundas Element. The region includes the Burnie Formation, and the previously defined Dial Range Trough (oriented north-south between the Forth Massif and the main outcrop area of the Burnie Formation) and Fossey Mountains Trough (oriented east-west adjacent to the Tyennan Element in the south of the region). The Dial Range and Fossey Mountains Troughs are cited specifically where appropriate.

The oldest exposed rocks around the margins of the Sheffield Element are the metaquartzite and metapelite of the Tyennan Element in the south, and the probably equivalent Ulverstone and Forth Metamorphic Complexes which form inliers in the central north. These metamorphic rock sequences are assumed to be of broadly similar age, perhaps in the vicinity of the ca. 1100 Ma depositional/ provenance age determined for metapelite in the Tyennan Element (Råheim and Compston, 1977; see also Turner *et al*., 1992).

Burnie Formation

Early Palaeozoic sedimentary sequences rest on the weakly metamorphosed but polydeformed turbiditic quartzose wacke – mudstone sequence of the Burnie Formation in the northwestern part of the Sheffield Element.

Control on the depositional age of the Burnie Formation is provided by a K-Ar date of 725 ± 35 Ma for the Cooee Dolerite (Crook, 1979), contact relationships of which suggest that it intruded the Burnie Formation while the sediments were still wet and relatively unconsolidated. U-Pb dating of zircons from the Cooee Dolerite (Turner, 1993*b*; Black, 1994) yielded only inherited grains with ages ranging from 1400–2000 Ma, with over 50% of grains in the range 1700–1800 Ma and one grain at ca. 2400 Ma. Turner *et al*. (1994) consider that these inherited zircon ages provide good evidence that the mafic magma was derived by melting of a 1700–1800 Ma source, and that since the source apparently contained common zircon it may have been a mafic granulite. This may indicate the presence of a previously unknown old basement at depth. However, secondary inheritance of the old zircon grains is also a possibility, and if so they would indicate only the original provenance of this basement.

Correlates of the Burnie Formation appear to structurally overlie the Forth Metamorphic Complex around the western and southern margins of the Forth Inlier, but the contact is not well exposed. The contact between the Ulverstone Metamorphics and correlates of the Burnie Formation is well exposed at Goat Island on the north coast, and is a generally shallowly-dipping but re-folded major thrust fault called the Singleton Thrust (Burns, 1964).

Polydeformed correlates of the Burnie Formation form the largely fault-bounded Badger Head Inlier near the northeastern margin of the Sheffield Element. Geophysical and structural studies (Leaman *et al*., 1973; Elliott *et al*., 1993) have indicated that the western boundary of the Badger Head Inlier is a major west-directed thrust fault against a complex sequence known as the Port Sorell Formation, which is considered to be a tectonic melange (Elliott *et al*., 1993; see further discussion below).

?Early Cambrian sequences

The Late Precambrian–Early Palaeozoic stratigraphy of the Sheffield Element is less well known than that of the main part of the Dundas Element, and current opinion differs significantly from the stratigraphic relationships proposed in the first major study of the region (Burns, 1964). Interpretation of field relationships led Burns to believe that the Barrington Chert and Motton Spilite are autochthonous units lying above an unconformity on the fossiliferous Cateena Group and below a possible disconformity at the base of the fossiliferous Radfords Creek Group. A similar interpretation was adopted in a recent major structural study of the region (Woodward *et al*., 1993). However, current belief is that the Barrington Chert and Motton Spilite are the oldest units in the post-Burnie Formation sequence. Furthermore, the trend of opinion is that they are probably allochthonous units tectonically emplaced during the same late Early to early Middle Cambrian arc-continent collision event believed responsible for the emplacement of the Cleveland–Waratah association and related rocks in the Dundas Element, and the mafic-ultramafic complexes throughout western Tasmania.

The Barrington Chert is present in both the Dial Range and Fossey Mountains Troughs and is up to ~1000 m thick (Burns, 1964; Jennings, 1979). Internal features of the chert include fine lamination, flaggy bedding, interbedded chert breccia units, and irregular colour variations which are unrelated to bedding (Jennings, 1979). Intraformational slumping (Burns, 1957) and small-scale soft-sediment deformation (Jennings, 1979) suggest instability in the basin of deposition. Prior to the suggestion that it may be an allochthonous unit, the Barrington Chert was considered to be a possible correlate of the Success Creek Group in the Dundas Trough (Brown, 1989*b*).

Structurally overlying the Barrington Chert in the Dial Range–Fossey Mountains region is the Motton Spilite (Jennings, 1979), a lenticular unit up to ~500 m thick. This unit comprises pillowed and massive basaltic lavas with interbedded volcaniclastic sandstone, mudstone and breccia, the latter locally containing abundant chert detritus (Burns, 1964; Jennings, 1979). The lavas are fine- to medium-grained augite-bearing tholeiitic basalts. On the basis of limited chemical and petrographic evidence, the Motton Spilite has been correlated with basaltic lavas in the Crimson Creek Formation in the Dundas Element (see discussion in Brown, 1989*b*). However, a study of the chemistry of the clinopyroxene phenocrysts and the style of chemical and mineralogical alteration in the lavas has since indicated that the Motton Spilite has Ocean Floor Basalt affinity (Hashimoto *et al*., 1981). This suggests it is more akin to lavas in the Cleveland–Waratah association of the Dundas Element than to those in the Crimson Creek Formation (and equivalents in the Smithton Synclinorium), as the latter have Within Plate Basalt geochemical affinities (Brown, 1989*b*).

If the Barrington Chert and Motton Spilite are allochthonous units, their original ages of formation are essentially unknown, except that they pre-date the Radfords Creek Group which includes basal conglomerate containing clasts of basalt and chert (Burns, 1964).

Inferred allochthonous units are also present east and west of the Badger Head Inlier in the northeastern part of the Sheffield Element. In the east, the Badger Head Inlier is faulted against a thin sliver of oolitic chert, slate and impure sandstone which appears to have been contact metamorphosed against the Andersons Creek Ultramafic Complex to the east (Gee and Legge, 1974; Elliott *et al*., 1993). Chert in the sliver contains undated columnar stromatolites (Banks, p. 82 *in* Brown, 1989*b*), and has been assumed to be of Cambrian age (Elliott *et al*., 1993). However comparisons with the Barrington Chert, or with the Success Creek Group of the Dundas Element, also spring to mind. The Andersons Creek Ultramafic Complex is an elongate body of serpentinite, pyroxenite and gabbro considered to represent a tectonically re-emplaced layered complex (Gee and Legge, 1974; McClenaghan and Baillie, 1975). It is fault-bounded to the east against an east-dipping imbricate thrust stack which includes slices of Ordovician sandstone and Cambrian sedimentary rocks. The western margin of the Badger Head Inlier is thrust-faulted over a complexly-deformed sequence of marine sedimentary rocks and dolerite, named the Port Sorell Formation by Elliott *et al*. (1993), who considered it to represent a tectonic melange possibly developed within an east-dipping accretionary wedge. Elliott *et al*. (1993) considered the Badger Head Inlier to be part of a fault-bounded allochthonous slice which was brought to the present structural level during an inferred west-directed Cambrian thrusting event, which involved an imbricated accretionary complex containing dismembered ophiolite slices.

Middle to Late Cambrian fossiliferous sequences

The Middle Cambrian Cateena Group in the Dial Range Trough comprises about 1000 m of mudstone, lithic wacke, feldspathic sandstone, conglomerate, and minor felsic volcanic rocks (Burns, 1964; Jago and Brown, 1989). A distinctive conglomerate at the base contains angular pebbles of purple mudstone and rests with apparent unconformity on correlates of the Burnie Formation along the western border of the Forth Inlier (Burns, 1964). Fauna of the Cateena Group indicate ages ranging from Florian to Undillan (Jago and Brown, 1989).

The relationship between the Cateena Group and the late Middle to early Late Cambrian Radfords Creek Group in the Dial Range Trough is not clear, but they are thought to possibly form part of a continuous sedimentary succession (Jago and Brown, 1989). The Radfords Creek Group consists of several hundred metres of mudstone, lithic wacke and minor felsic volcanic rocks, with basalt-rich and chert-rich conglomerate at the base (Burns, 1964; Jago and Brown, 1989). The sequence structurally overlies the Barrington Chert, and a possible disconformity between the two sequences was suggested by Burns (1964), but the current trend of opinion is that the chert is tectonically emplaced. Fossils in the Radfords Creek Group indicate ages ranging from Boomerangian to Late Mindyallan (Jago and Brown, 1989), and the sequence is unconformably overlain by the Duncan Conglomerate, an equivalent of the Owen Group of the Dundas Trough.

A particularly distinctive feature of the Dial Range Trough is the occurrence of the Beecraft and Teatree Point Megabreccias at the western margin of the trough. Both units are about 150 m thick and consist of blocks of chert and other lithologies up to 120 m long, in a matrix of lithic wacke and conglomerate (Burns, 1964). The Beecraft Megabreccia rests with angular unconformity on the Burnie Formation on the north coast at Penguin. It has been suggested that the breccias were formed by gravity sliding of large masses of semi-indurated material from unstable flanks of the trough (Burns, 1964). The megabreccias may be older than, or partly coeval with, the Radfords Creek Group (Jago and Brown, 1989).

In the central part of the Dial Range Trough the Lobster Creek Volcanics form an 8×1 km body of massive andesitic plagioclase-pyroxene-hornblende porphyry. This unit was originally thought to be extrusive (Burns, 1964), but Jago *et al*. (1977)

suggested an intrusive origin on the basis of its massive nature, lack of intercalated sediments or pyroclastic rocks, associated marginal mineralisation, and an Early Ordovician radiometric age (re-calculated Rb-Sr 480 ± 18 Ma) obtained from a dyke of similar composition (Corbett and Solomon, 1989). Adams *et al*. (1985) also obtained an Ordovician Rb-Sr age (456 ± 22 Ma) for the main Lobster Creek mass, but considered it to be a minimum age, stating a preference for the age reported by Jago *et al*. (1977) because it was derived from a model 1 isochron for samples from a more restricted area.

The Cambrian stratigraphy of the Fossey Mountains Trough is still quite poorly known, although it is apparent that the proportion of felsic to intermediate volcanic and volcaniclastic rocks increases towards the southern margin of the Sheffield Element, suggesting an eastward extension of the Mt Read Volcanic belt from the Dundas Element. The volcanic sequences are mostly quartz-feldspar phyric, and this, together with known fossil horizons in the Middle to Late Cambrian range, suggests that the sequences are equivalents of the Dundas and Tyndall Groups (Corbett and Solomon, 1989). Andesitic and basaltic rocks, including pyroxene-phyric lava and breccia, occur in the "Beulah Formation" of Jennings *et al*. (1959) and Jennings (1979). The Cambrian sedimentary successions in the Fossey Mountains Trough include greywacke-mudstone sequences, which in the Native Track Tier area in the central western part of the trough contain rich trilobite faunas indicating an age in the Late Boomerangian or Early Mindyallan of the late Middle Cambrian (Baillie and Jago, 1985). A probably slightly older fauna, of Boomerangian age, was recovered from a siltstone unit within a sequence of felsic lavas and porphyries near Paradise in the eastern part of the more volcanic-rich southern part of the trough (Jago, 1989). Apart from these two occurrences, the only other biostratigraphic information is based on two poorly known faunas of probable Late Cambrian age from the central part of the Fossey Mountains Trough (Jago, 1979).

In the northeastern part of the Sheffield Element, a siltstone unit within the thrust stack east of the Andersons Creek Ultramafic Complex has yielded late Middle–early Late Cambrian fossils (Green, 1957, 1959; Gee and Legge, 1974).

Late Cambrian–Early Ordovician siliciclastic sequences

Siliciclastic conglomerate-sandstone sequences comparable to the Owen Group are widespread throughout the Sheffield Element. They form a prominent belt extending from the Black Bluff Range through Mt Roland and the Gog Range within the Fossey Mountains Trough, with a second belt extending from St Valentines Peak through Loyetea and Gunns Plains to the Dial Range. The age of the sequences is constrained by the youngest biostratigraphic ages from the underlying sequences (i.e. late Mindyallan or very early Late Cambrian), and the age of the uppermost fossiliferous marine sandstone (Bendigonian Stage of the Early Arenig) underlying the Gordon Group limestone.

The sequence in the Black Bluff–Moina area has recently been mapped at 1:25 000 scale as part of the Mt Read Volcanics Project (Vicary and Pemberton, 1988; Pemberton and Vicary, 1988, 1989), and descriptions are given in Pemberton, Vicary and Corbett, (1991) and Corbett (1990). The sequence can be related to that of the Owen Group in the Dundas Element, and includes correlates of the Jukes Conglomerate (basal lenses of volcaniclastic conglomerate and sandstone), Newton Creek Sandstone (grey micaceous sandstone and siltstone at the base of the sequence in some areas), Middle Owen Conglomerate (thick-bedded cobble-boulder conglomerate with minor sandstone lenses, disconformable at the base in places, wedging out and disappearing just east of Black Bluff, but probably re-appearing at Mt Roland), and Upper Owen sequence (see discussion, *Dundas Element*). Clasts in the conglomerates show compositions predominantly similar to the siliceous Precambrian metasedimentary rocks of the Tyennan Element, a provenance indication also supported by palaeocurrent evidence (Seymour, 1980; Banks and Baillie, 1989, p.198).

The Upper Owen sequence is much expanded in the Black Bluff area, and consists of three sandstone units, two of which have basal conglomerate resting on erosional disconformity surfaces. The lower sandstone ('Linda Sandstone correlate') is generally thin-bedded with minor siltstone intercalations, abundant bioturbation and rare marine fossils. An extensive sill-like body of dolerite, about 50 m thick, occurs within the upper part of this unit, and is thought to be possibly of early Ordovician age because it is altered, pre-dates the Devonian folding, and is probably genetically related to basalt at a slightly higher stratigraphic level (Pemberton, Vicary and Corbett, 1991). An erosional and transgressive unconformity, possibly equivalent to the Haulage Unconformity, separates the thin-bedded sandstone from a basal pebble conglomerate of the next sandstone unit, which is typically pink, coarse-grained and abundantly cross bedded. An abundance of chert clasts in this unit supports a correlation with the Pioneer Beds of the Queenstown area. Of particular interest is the occurrence within the upper part of this sandstone unit of flows of altered porphyritic basalt within a 'matrix' of volcaniclastic sandstone. This thin unit of basalt and volcaniclastic rocks has been mapped as far east as the Wilmot River and Mt Jacob (Pemberton and Vicary, 1989). The basalts are extremely altered to chlorite and hematite in most areas, and their primary chemistry is uncertain. However, a genetic relationship to the previously described dolerite seems likely. A basal pebble-cobble conglomerate resting on an erosional and transgressive disconformity forms the base of the uppermost sandstone unit, which is widespread and corresponds to the Moina Sandstone of Jennings (1958) and Jennings *et al*. (1959). This unit of grey, cross-bedded and strongly bioturbated sandstone with distinctive 'pipestem' burrows is 100–300 m thick and contains only poorly preserved marine fossils in most places. A richly fossiliferous sandstone at Caroline Creek, near Railton, is probably equivalent to the upper part of the formation and is of Early Arenig age (Bendigonian to Early Castlemanian — Banks and Baillie, 1989).

Earlier mapping of the siliciclastic sequences in the Sheffield–Middlesex area (e.g. Jennings, 1958; Jennings *et al*., 1959) used a bipartite system based of a lower "Roland Conglomerate" and an upper "Moina Sandstone". It is apparent, however, that much of the "Roland Conglomerate" as mapped corresponds to the basal conglomerate of the Moina Sandstone (plus elements of underlying units in some places), whereas the main conglomerate on Mt Roland, which is of the order of 250 m thick, is probably equivalent to the Middle Owen Conglomerate.

A lower unit of siliceous conglomerate in the Dial Range Trough (Duncan Conglomerate of Burns, 1964), about 500 m thick, is overlain by a sandstone unit up to 250 m thick which is correlated with the Moina Sandstone. The Duncan Conglomerate is unusual in that the clasts consist predominantly of chert, presumably derived from intra-basinal sources such as the Barrington Chert. Other clast types include quartzite derived from the Burnie Formation to the west, hematite and limonite (probably from orebodies in the Iron Cliffs area), and rare boulders of Cambrian lava and mudstone. A thin unit of interbedded mudstone and sandstone with siliceous conglomerate bands, referred to as the Gnomon Mudstone, occurs beneath the conglomerate in places (and may be an equivalent of the Newton Creek Sandstone).

In the central northern part of the Sheffield Element, the siliciclastic sequence contains a relatively thin (maximum 275 m) basal unit of chert breccia or siliceous conglomerate, the latter derived from the basement rocks of the Forth inlier to the northwest (Jennings, 1979; Scanlon, 1976). This basal unit is followed by the Caroline Creek Sandstone, a sequence of well-sorted siliceous sandstone and siltstone containing trace fossils (burrows both perpendicular and parallel to bedding) and marine macrofossils (Scanlon, 1976). The macrofossils indicate an Early Arenig (Bendigonian to Early Castlemanian) age for the sequence (Laurie, 1982; Jell and Stait, 1985; Banks and Baillie, 1989).

In the northeastern part of the Sheffield Element, the Cabbage Tree Formation comprises a siliciclastic conglomerate-sandstone sequence within the east-dipping imbricate thrust stack east of the Andersons Creek Ultramafic Complex. This sequence has yielded Early Ordovician fossils (Tremadoc–Upper Arenig; p. 193 in Banks and Baillie, 1989), and is correlated with the Owen and lower Denison Groups of western Tasmania. Chromite detritus in the Cabbage Tree Formation, and abundant chert clasts in the conglomerate units, suggest a westerly derivation from the Andersons Creek Ultramafic Complex and the cherts west of that complex (Banks and Baillie, 1989), a suggestion supported by limited palaeocurrent evidence (Hills, 1982).

Gordon and Eldon Groups

The Late Cambrian–Early Ordovician siliciclastic sequences in the Sheffield Element are conformably overlain by the fossiliferous, limestone-dominated Gordon Group. The region includes one of the main reference sections for the Gordon Group, at Mole Creek in the eastern part of the Fossey Mountains Trough, where the group is over 1200 m thick (Banks and Baillie, 1989). Based on the preliminary biostratigraphic system for the Tasmanian Ordovician of Banks and Burrett (1980), faunal assemblages between OT8 (Arenig) and OT19 (?Early Ashgill) have been identified in the Sheffield Element (Banks and Baillie, 1989). However the base of the Gordon Group is diachronous across the region, with carbonate deposition starting earlier in the eastern part of the region compared with the west. The earliest faunas are present east of the Badger Head Inlier, where the Flowery Gully Limestone contains Arenig conodonts near the base and Llanvirn or Llandeilo conodonts near the top (Kennedy *in* Banks and Baillie, 1989, p.213). Regionally it also appears that Gordon Group deposition began earlier in the Sheffield Element than in the Dundas Element.

Outcrops of Eldon Group correlates are scant in the Sheffield Element, occurring near St Valentines Peak close to the western limit of the region, near the southern margin of the region south of Mole Creek, and perhaps also at Flowery Gully east of the Badger Head Inlier near the northeastern margin of the region. The St Valentines Peak sequences are dominantly fine-grained sandstones correlated with the Florence Quartzite and Bell Shale, and marine macrofossil assemblages indicate Pragian and Early Emsian ages (Banks and Baillie, 1989). At Flowery Gully a thin unit of fine siltstone or slate containing Late Ordovician or Early Silurian brachiopods (Banks and Baillie, 1989) has been inferred to rest disconformably above the Flowery Gully Limestone (Hills *in* Banks and Baillie, 1989). The siltstone unit has been correlated with the Arndell Sandstone (Hills *in* Banks and Baillie, 1989), which is the uppermost unit of the Gordon Group in its type section in the Adamsfield–Jubilee Element, but the inferred disconformable lower contact suggests it may be an Eldon Group correlate (Elliott *et al*., 1993). Regionally it seems possible that equivalents of the lower four formations of the Eldon Group in its type area in the Dundas Element were not deposited in the Sheffield Element.

Also in the Flowery Gully area is a poorly exposed sequence of siltstone and mudstone containing mid-Early Devonian fossils, which has been correlated with the Devonian Mathinna Group east of the River Tamar (Hills, 1982; Baillie *et al. in* Banks and Baillie, 1989, p.236). Re-examination of this sequence (Powell *et al.*, 1993) showed that it has none of the typical turbiditic characteristics of the Mathinna Group east of the Tamar, and that it is more likely an equivalent of the Bell Shale of the Eldon Group of western Tasmania.

Devonian and earlier deformation

The regional orogenesis of late Early to early Middle Devonian age in western Tasmania is probably best known in the Sheffield Element, where the most precise dating of the deformation and the best knowledge of the structural sequence has been established. In the middle to western part of the Fossey Mountains Trough, intersecting fold trends have produced dome and basin structures and some overprinting relationships which have assisted analysis of the structural history. The structural style is of generally upright to steeply inclined, open to tight folds with horizontal to moderately plunging hingelines, and associated axial plane cleavage development and faulting. Many of the faults are steep thrusts associated with the development of inclined folds, particularly in the siliciclastic sequences. Williams (1976, 1979) proposed a subdivision of the Devonian structures in the Sheffield Element into three trend groups, two of which formed an early fold phase, and the other a late fold phase. This scheme was refined by Seymour (1980, and in Williams *et al.*, 1989) to show a sequence of four distinct phases of folding in the middle to western part of the Fossey Mountains Trough (see below).

While the existence of these four distinct groups of Devonian folds appears to be well established, their sequence of development is based on relatively few clear overprinting relationships. D_1 and D_4 are probably the most precisely determined, while the relative positions in the sequence of D_2 and D_3 are probably the least well established. Only D_3 and D_4 structures are of regional significance throughout western Tasmania (Williams, 1976, 1979), while D_1 appears to be confined to the Sheffield Element and the northernmost part of the Dundas Element, and D_2 appears to be a feature mostly of the northern part of the Dundas Element between the northeastern margin of the Tyennan Element and the Mt Bischoff Inlier (Seymour, 1980; Seymour in Williams *et al.*, 1989; P. R. Williams *in* Seymour, 1989).

The maximum age of the Devonian orogenesis is constrained by the biostratigraphic age of the youngest sequences affected, which in the Sheffield Element is Early Emsian (Banks and Baillie, 1989). Sequences of this age at St Valentines Peak, near the western margin of the region, appear to be affected by at least the Devonian D_3 and D_4 events, based on the mapping of Baillie *et al.* (1986). The minimum age is constrained by the Eugenana Beds near the northeastern margin of the Sheffield Element. These are undisturbed terrestrial cavern fillings which were deposited in cavities within a collapse breccia consisting of disoriented blocks of deformed Gordon Group limestone, and contain spores which indicate a latest Givetian age, i.e. in the latest part of the late Middle Devonian (Balme, 1960; Burns, 1964; McGregor and Playford, 1992). In this area the Eugenana Beds therefore post-date the deformation of the Gordon Group. The dominant Devonian structures in the Eugenana area are folds of NNW trend with associated axial plane cleavage dipping at about 60° east (Burns, 1964), and could be argued to represent either the Devonian D_3 or D_4 listed in the table above. However, regionally it appears possible to trace a progressive swing in D_4 fold trends from WNW-ESE in the southeastern corner of the Sheffield Element, through NW-SE to NNW-SSE trends at Eugenana (Williams *et al.*, 1989). Together with the westward vergence of the main Devonian structures at Eugenana, this appears to support assignment of the latter structures to the D_4 generation.

The most recent re-interpretation of the Sheffield Element is the structural study of Woodward *et al.* (1993), which focused particularly on the history and origin of major faulting, an aspect not fully addressed in earlier studies. The resulting interpretation shows deformation of thin-skinned style, with imbrication and stacking of thrust slices

DEVONIAN STRUCTURAL SEQUENCE (SEYMOUR, 1980)

Phase	*Dominant trend and vergence*	*Style*
D_4 (youngest)	NW–SE; vergence to SW	Open to tight, generally inclined folds with associated thrust faults and axial plane cleavage.
D_3	N–S; commonly no preferred vergence in this region.	Open to tight, upright folds with associated axial plane cleavage.
D_2	NE–SW; vergence to SE.	Open to tight, upright to inclined folds with associated axial plane cleavage.
D_1 (oldest)	E–W; no preferred vergence	Open upright folds, generally with no associated cleavage development.

of all pre-Eugenana Beds strata including the Precambrian basement, above a gently northeast-dipping detachment surface lying at depths of 5–10 km below the region. The model relies partly on a simplified regional stratigraphy resulting from re-interpretation of relationships shown on existing maps. Woodward *et al.* (1993) do not consider the Barrington Chert and Motton Spilite to be tectonically emplaced between the Cateena and Radford Creek Groups (contrary to some current opinions), and they apparently consider the Barrington Chert to be at least partly derived by silicification of a pelitic protolith in the Cateena Group.

The cross-section in Woodward *et al.* (1993) shows two major thrust slices of Precambrian basement (the main present outcrops of which are the Badger Head and Forth massifs), up to about 3 km thick, imbricated within the trough sequences. The interpreted geometry is believed to be a result of an early southwest-directed thrusting event of Cambrian age overprinted by further southwest-directed thrusting of Middle Devonian age. The proposed full deformation sequence for the region, including Cambrian and post-Devonian events, is as shown below (Woodward *et al.*, 1993, with corrections from Gray and Woodward, 1994).

The D_4 structures overprint all earlier fabrics, but are not clearly cross-cut by the post-Permian faults. However, all post-Permian structures are brittle, while the D_4 structures are more ductile, suggesting that the latter are earlier. According to Woodward *et al.* (1993), geometric restoration of their cross-section indicates that a total shortening in the order of 80 km occurred to produce the present geometry, with at least half of this displacement attributable to the Late Cambrian deformation.

Woodward *et al.* (1993) cite three lines of evidence for their suggested Late Cambrian thrusting event:–

- A pronounced angular unconformity occurs at the contact of the Late Cambrian–Early Ordovician siliciclastic conglomerate sequences with the underlying trough sequences at a number of localities in the region. Dramatic changes in Cambrian stratigraphic units beneath the unconformity across certain narrow zones of Devonian deformation also suggest that Cambrian movement on discrete faults juxtaposed Cambrian facies prior to deposition of the siliciclastic conglomerate sequences.

- It is proposed that the isolated blocks of Precambrian basement rocks within the Sheffield Element all overlie low-angle thrusts, and geophysical modelling suggests that these blocks are not rooted, but thin basement thrust sheets (Leaman *et al.*, 1973; Leaman, 1990*a*, 1992). Most of these basement blocks are presently juxtaposed against Late Cambrian and younger siliciclastic sequences on faults with at most hundreds of metres of Devonian displacement, but the basement rocks directly underlie the siliciclastic sequences at the southeastern margin of the Forth Massif, and in the eastern part of the Fossey Mountains Trough. Most of the displacement on faults which brought Precambrian basement rocks to the surface must therefore be Cambrian.

- In the Fossey Mountains Trough the Tyennan Element has always been considered the source area for quartzite clasts in the Roland Conglomerate, yet no conglomerate directly overlies the suggested source rocks, and the conglomerate is thickest in the northern Gog Range, some distance away from the nearest exposed rocks of the Tyennan Element proper. In the Dial Range Trough the clast composition of the Dial Conglomerate indicates sources in the Barrington Chert as well as in the Rocky Cape and Forth Massifs. Woodward *et al.* suggest that the siliciclastic conglomerate units may have

STRUCTURAL SEQUENCE (WOODWARD *et al.*, 1993)

Event	*Age*	*Characteristics*
D_5 (youngest)	Post-Permian	NW-trending dip-slip extensional faults, mostly down-to-the-east
D_4		Dextral strike-slip faults and shear zones of 160–170° trend, with local development of vertical cleavage of 170° trend.
D_3	Middle Devonian	ESE-vergent, NNE-trending thrust faults which re-deform SW-vergent thrusts and folds in Cambrian strata in the Dial Range area.
D_2	Middle Devonian	SW-vergent major thrust faults, folds and shear zones of 135° trend, and E-W trending folds, in the Fossey Mountains Trough area.
D_1 (oldest)	Late Cambrian	NW-trending, SW-vergent thrust faults causing at least two repetitions of the Cambrian trough sequence and its underlying basement on top of the Tyennan autochthonous basement.

shed from crystalline basement sources within thrust sheets emplaced *within* the basin prior to the middle Late Cambrian, and similarly that the chert component in the Dial Conglomerate shed from an emergent ramp-related anticline of Barrington Chert overlying a Cambrian thrust within the basin.

In a recent tectonic synthesis of western Tasmania, Berry (1994) differs from the model of Woodward *et al.* (1993) in regard to the timing of their D*1* event which emplaced the basement slices. Berry places this event in the late Early to early Middle Cambrian (525–510 Ma), based partly on the presence of garnet and biotite clastic grains in the Radfords Creek Group, which implies that the Forth Metamorphic Complex had been emplaced prior to the Middle Cambrian (Berry, pers. comm.). The synthesis of Berry (1994) also shows a N-S compressional event in the Fossey Mountains Trough in the late Middle to early Late Cambrian (during the 510–490 Ma Delamerian Orogeny), producing west-trending folds which were later to be tightened in the Devonian.

Devonian granitoid rocks

The major post-tectonic granitoid intrusive in the Sheffield Element is the Housetop Granite, which outcrops over a 120 square kilometre area near the central western margin of the region. The Housetop Granite generally consists of equigranular to sparsely porphyritic, medium-grained to coarse-grained biotite granite with minor variants, the most important of which is a fine-grained porphyritic granite with phenocrysts of quartz and feldspar (P. W. Baillie and P. G. Lennox *in* Seymour, 1989). The Housetop Granite does not appear to be composite, as despite reasonably detailed mapping, individual plutons have not been recognised within the body. The contacts are sharp and discordant, and the presence of several right-angled bends in the intrusive boundary suggests that emplacement was controlled to some extent by pre-existing joints (or faults), and that the mechanism of intrusion was probably by stoping (Baillie and Lennox, ibid.).

A provisional geophysical interpretation of the subsurface form of the Housetop Granite indicates that a large proportion of the roof of the body is exposed (Leaman and Richardson, 1989). The body has yielded Rb-Sr and K-Ar radiometric ages with a maximum range of 343–380 Ma including uncertainty limits (Williams *et al.*, 1989, after McDougall and Leggo, 1965).

The Dolcoath Granite, which intrudes the southern part of the Sheffield Element, has considerable shallow subsurface extent west of the small exposure near Cethana, according to gravity-based geophysical interpretation (Leaman and Richardson, 1989). The gravity interpretation has also indicated a large shallow subsurface extent for the Beulah Granite north and west of its small patchy outcrops near Beulah and Paradise, some 20 km ENE of the Dolcoath Granite outcrop, and which are described as fine-grained granite and microgranodiorite in Jennings (1979).

Granitoid-related mineralisation

The Housetop Granite is a mineralising granite, being directly associated with several Sn-W-magnetite skarn deposits (e.g. Kara) and some Pb-Zn-Ag, and Cu, vein deposits. The Dolcoath Granite is considered responsible for an unusual magnetite-fluorite-vesuvianite skarn with minor W and Sn, which occurs with other skarn types in basal Gordon Group carbonate rocks at Moina. Sn-W-Bi vein deposits (Shepherd and Murphy mine) are also associated with the Dolcoath Granite.

At the Tasmania mine at Beaconsfield, near the northeastern margin of the Sheffield Element, a major auriferous quartz reef filled a fault zone transgressing sandstone and quartzite units in the upper part of the early Ordovician Cabbage Tree Formation (Hicks and Sheppy, 1990). However, despite an apparent genetic relationship between Devonian granodiorites and some of the known gold deposits in the Northeast Tasmania Element, the remoteness of the Tasmania reef from known or inferred Devonian granitoids makes a genetic link difficult to prove (McClenaghan in Williams *et al.*, 1989, p.291). Russell (1994) favours a mineralisation model involving a deep geothermal system through which seawater and metamorphic waters circulated through a fracture system, and which may have involved a deep-seated magmatic plume.

Tyennan Element (DBS)

The term Tyennan Element refers to the substantial belt of metasedimentary and minor meta-igneous rocks and minor younger cover sequences extending north-south from central western Tasmania to the south coast, bounded to the west by the Dundas Element, to the north by the Sheffield Element, to the southeast by the Adamsfield/Jubilee Element, and overlapped to the east by the younger cover of the Tasmania Basin (fig. 1).

Metamorphic rocks

The extensively exposed metamorphic rocks of the Tyennan Element are subdivided into quartzite-chloritic pelite assemblages of lower greenschist facies and garnetiferous schist-quartzite assemblages of upper greenschist to eclogite facies (Turner, 1989*a*). Some transitional contacts are known between the two types of assemblage but many boundaries are thought to be major movement zones. P-T determinations indicate that disruption of the original orogenic pile (probably largely due to major thrusting during the syn-metamorphic D_2 deformation event) has juxtaposed rocks drawn from depths ranging from about 12 km to well in excess of 30 km (Turner, 1989*a*). The lower rank assemblage in most areas is derived from a suite of lithologies characterised by quartzarenite like that in the Rocky Cape Group and elsewhere in the relatively unmetamorphosed western part of the Rocky Cape Element. Evidence of the type of protolith from which the higher rank assemblage was derived has been obscured by the metamorphic overprint, but Turner (1989*a*) points out that the chemical composition of pelitic rocks in each metamorphic assemblage is similar, and pure quartzite is present in each, so they may have been derived from similar sedimentary suites. This interpretation is adopted herein.

The original age of deposition of the sedimentary protolith of the Tyennan Element metamorphic rocks is indicated by Rb-Sr model ages of phyllite near Strathgordon, which range from 1100–1150 Ma and are considered to probably correspond to the age of deposition (Råheim and Compston, 1977). This general age range is supported by a Pb-Pb isochron derived from metasiltstones sampled from the Lyell Highway, which indicated an age of 1300 ± 170 Ma (Gulson *et al*., 1988).

The quartzite-chloritic pelite assemblage is typified by the Strathgordon area, west of the Adamsfield/Jubilee Element in the southern part of the Tyennan Element (Boulter, 1978; Turner, 1989*a*). Sedimentary structures and textures, and gross lithological style indicate deposition in a tidally-dominated shallow-shelf sea. A common rock type is super-mature quartzarenite (quartzite), which displays sedimentary structures indicative of marine deposition, e.g. <1 m scale herringbone cross-stratification and current patterns of bipolar modal and polymodal types. The sand grain shape characteristics are like those generated by extensive abrasion in a desert environment, and suggest a marine transgression over an aeolian setting (Boulter, 1978; Turner, 1989*a*). The more pelitic lithologies (now schist/phyllite) may have been deposited in tidal flat or deltaic environments. Minor tholeiitic dolerite dykes pre-date the metamorphism. Only greenschist facies mineral assemblages have been recorded, the common pelite assemblage being quartz + tourmaline ± chlorite, although one area shows incipiently crystallised garnet. The rocks show a polyphase deformation history, and peak metamorphic conditions of 400 ± 50°C and 300 ± 100 MPa occurred just before or during D_2 (Råheim, 1977). Deformation phases D_1 to D_5 were essentially coaxial, with sub-horizontal fold hinges which show a progressive regional swing in trend from N-S to E-W through the area. D_1 folds were originally recumbent isoclinal fold nappes with inverted limbs up to 5 km in length, resulting in common parallelism of bedding and S_1. All D_1 folds face east or northeast, and this sense of over-riding from the west was continued in D_2.

Most work on the higher rank metamorphic rocks of the Tyennan Element has focused on the Franklin Metamorphic Complex in the Lyell Highway–Franklin River area in the central part of the Element. Garnetiferous rocks are present throughout the Franklin Metamorphic Complex, and on Raglan Range the rocks comprise about equal proportions of massive to schistose quartzite and coarse-grained, commonly knotted schist (Turner, 1989*a*). The dominant minerals in the schist are quartz, muscovite, albite and almandine garnet, and biotite is common. Primary chlorite is widespread, garnet is commonly altered to secondary chlorite, and uncommon clusters of small kyanite grains are partially retrogressed to muscovite. The quartzite units are thick, commonly discontinuous slabs which in some cases are detached fold cores, and no sedimentary structures have been recorded in them. Minor amphibolite occurs as either massive boudins, or as schistose tabular bodies concordant with the main foliation. The schists of the Franklin Metamorphic Complex are polydeformed, but folds related to S_1 can rarely be identified, whereas common D_2 fold closures range from open to isoclinal (Turner, 1989*a*). Turner (1971) concluded that a single prograde metamorphic event had occurred in the Franklin Metamorphic Complex, commencing near the beginning of crenulation associated with D_2 and persisting until after maximum cleavage development. Little metamorphic crystallisation appears to have accompanied D_1. The close association of peak metamorphism with D_2 is consistent with the relative timing of metamorphism and deformation in the lower rank metamorphic rocks of the Strathgordon area.

The highest grade rocks in the Franklin Metamorphic Complex occur in the Lyell Highway–Collingwood area (Råheim, 1976; Kamperman, 1984), where metasedimentary rocks consist of garnet-mica schist, mica schist and garnet-mica-kyanite gneiss containing migmatite veinlets. Boudins of eclogite and garnet amphibolite (?after basaltic lavas) are present, the dominant eclogite mineral assemblage being garnet-omphacitic clinopyroxene-amphibole-phengite-quartz. The dominant assemblage in the enclosing schists is garnet-phengite-quartz-biotite ± albite ± kyanite ± tourmaline (M. Kamperman *in* Turner, 1989*a*). Peak metamorphism may have accompanied mylonitisation (D_2) which produced an LS fabric in the metasedimentary rocks with a generally southeastward sense of tectonic transport (R. F. Berry *in* Turner, 1989*a*). Peak metamorphic conditions in the eclogite of 715–730°C at 1560–1700 MPa were calculated by Kamperman (1984) from various mineral geothermometers and geobarometers. A consistent result of 698 ± 28°C at 1520 ± 105 MPa was derived from equilibrium thermodynamics calculations by Goscombe (1990). Råheim (1976), Kamperman (1984) and Goscombe (1990) all inferred that the metasediments adjacent to the eclogite had experienced the same prograde metamorphic conditions as the eclogite. This indicates that the eclogite and adjacent rocks have experienced depths of burial in excess of 30 km and perhaps as much as 50 km.

Until recently the accepted age of the main metamorphism in both the lower rank and higher rank assemblages in the Tyennan Element was based on a preliminary radiometric evaluation of the Strathgordon and Lyell Highway areas by Råheim and Compston (1977). This study used Rb-Sr geochronology on phengite-whole rock pairs and whole-rock isochrons, and in the Strathgordon area suggested an age of ca. 780 Ma for peak metamorphism, and that isotopic disturbance due to S_4 development occurred between 620 and 540 Ma (see Turner, 1989*a*). K-Ar slate and phyllite ages from near Strathgordon, and in the lower rank rocks elsewhere in the Tyennan Element, are also in the 540–620 Ma range (Adams *et al*., 1985), but are commonly syn-S_3 in their respective areas (Turner, 1989*a*). In the higher rank assemblage, interpretation of Rb-Sr data from the Lyell Highway–Collingwood River area by Råheim and Compston (1977) indicated consistency with results from the lower rank assemblages, with peak metamorphism (D_{1-2}) at ca. 780 Ma and a later, lower P-T event (?D_3) in the interval 540–620 Ma (see Turner, 1989*a*). The ca. 780 Ma structural and metamorphic event in the Tyennan Element was considered to represent the metamorphic peak of the Frenchman Orogeny of Spry (1962).

Opinion on the age of the main metamorphism in the Rocky Cape and Tyennan Elements has recently been completely revised, following re-evaluation of the complete set of existing whole-rock and mineral radiometric data, in conjunction with new mineral ages including U-Pb zircon data (Turner *et al*., 1992; 1994). This re-evaluation has generally involved rejecting whole-rock K-Ar and Rb-Sr dates in favour of mineral ages, on the basis that pelites in the Rocky Cape Element commonly contain not one, but several generations of mica in various proportions, and metasedimentary rocks in the Tyennan Element commonly show the effects of several episodes of mineral growth (Turner *et al*., 1994). Consideration of K-Ar, Ar-Ar and Rb-Sr metamorphic mineral ages alone indicates consistent peak metamorphic ages between the major regions, the approximate age ranges being 494–510 Ma in the Arthur Metamorphic Complex of the Rocky Cape Element, 496–515 Ma in the Tyennan Element, and 493–510 Ma in metasedimentary rocks of the Forth Inlier in the Sheffield Element (Turner *et al*., 1992, 1994). The new view of the age of the main metamorphism in the Tyennan Element appears to be confirmed by a recent U-Pb zircon age of 502 ± 8 Ma from the eclogite in the Franklin Metamorphic Complex (Black, 1994; Turner *et al*., 1994).

The new interpretation places the age of the main metamorphism and D_{1-2} deformation in the Tyennan Element either synchronous with, or closely following, the proposed arc-continent collision event which is now believed to have been responsible for the emplacement of the ultramafic complexes and numerous other major tectonic effects in western Tasmania (Berry and Crawford, 1988; Berry, 1994). In the Tyennan Element (and also in the Forth Inlier in the Sheffield Element), mylonitic fabrics and the common occurrence of faults which are subparallel to the axial surfaces of tight to isoclinal folds, and which commonly form the boundaries of major lithological units, indicate that widespread translation occurred in the D_{1-2} period (Turner, 1989*a*). Interlayering of metamorphic complexes drawn from different levels in the orogenic pile apparently occurred during this time through the stacking of thrust sheets, whereby rocks with depths of burial varying from 12 km to greater than 30 km were brought into close proximity (Turner, 1989*a*).

?Neoproterozoic sequences

The metasedimentary rocks in the Jane River area in the central part of the Tyennan Element are overlain by the little-studied Jane Dolomite, which consists of great thicknesses of dolomite with thin basal siliciclastic rocks and associated units of diamictite (Wells, 1955; Spry and Zimmerman, 1959; Gee, 1968; Hall *et al*., 1969*a, b*). The Jane Dolomite is regarded as a possible correlate of the Success Creek Group, Black River Dolomite and Weld River Group (Brown, 1989*b*). The contact of the Jane Dolomite with the underlying rocks may be an unconformity (Hall *et al*., 1969*a, b*; Gee, 1968), but Wells (1955) regarded the contact between the Jane

Dolomite and the underlying Scotchfire Metamorphics as conformable. The diamictite is described by Spry and Zimmerman (1959) as massive and unbedded, with angular fragments up to ~1 m in size (most being in the size range 5–10 cm) consisting mostly of dolomite but with some of unmetamorphosed mudstone, in a crystalline dolomitic matrix. Some clasts are composed of fine-grained quartz-muscovite schist, with the schistosity disoriented from clast to clast, and rare clasts consist of fine-grained slate, siltstone, and coarse even-grained quartzite.

In the Algonkian Rivulet–upper Maxwell River area just to the south of the Jane River area, correlates of the Jane Dolomite consist of thin-bedded dolomite, massive dolomite and dolomitic siltstone (Dixon, 1992). The dolomite sequence overlies older Precambrian rocks with low-angle unconformity. According to Dixon (1992) the massive dolomite contains grainstone, low-angle crossbedding, and possible stromatolitic laminations, while minor soft-sediment deformation occurs in the bedded dolomite.

The accepted provisional correlation of the Jane Dolomite with the Success Creek Group and correlates places its likely age of deposition well before the recently proposed ca. 505 Ma age for the main metamorphism and D_{1-2} deformation in the metasedimentary rocks on which it rests. This may pose a problem for the interpretations of Turner *et al*. (1992; 1994) which are followed herein, as the Jane Dolomite may be relatively unmetamorphosed, some published sources state that it rests unconformably on older rocks, and it contains schist fragments with disoriented schistosity. On the other hand, the underlying Scotchfire Metamorphics are of lower greenschist facies, consisting of schistose fine-grained sandstone or siltstone and dark grey mudstone (Turner, 1990*c*), and it may be difficult to establish a metamorphic rank lower than this for the Jane Dolomite which petrographically consists of a fine-grained aggregate of idioblastic dolomite rhombs with quartz, sericite and a little pyrite (Spry and Zimmerman, 1959).

?Middle Cambrian intrusive rocks

Near the northern boundary with the Sheffield Element, metamorphic rocks of the Tyennan Element are intruded by several small plug-like bodies of granitoid, collectively known as the Dove Granite (Jennings, 1963). These intrusive rocks are noted for their lithological variability, the presence of roof pendants and complex outcrop patterns, and Jennings (1963) considered that only the top of the granite is exposed and that many of the lithologies observed represent marginal phases. Exposures are commonly deeply weathered, but many samples have been described as granodiorites (Jennings, 1963), while other lithologies present include grey biotite granite, aplite and granite porphyry (Leaman and Richardson, 1989).

Thirteen K-Ar radiometric age determinations from the Dove Granite, including samples from all three outcrop areas, show a long total range (including uncertainties) of 448–519 Ma (re-calculated data after McDougall and Leggo, 1965). However, the Dove Granite is considered to be genetically related to, and so probably coeval with, Middle Cambrian granitoids in the Dundas Element, whose ages are well constrained.

Ordovician–Early Devonian sedimentary successions

Following the major ca. 505 Ma tectonometamorphic event, the Tyennan Element was inundated during much of the Ordovician to Early Devonian, when correlates of the marine carbonate and siliciclastic successions of the Gordon and Eldon Groups were deposited. These sequences are preserved in a number of outliers. Basal siliciclastic sequences underlying the Gordon Group sequence are relatively thin compared with equivalent sequences (Denison Group correlates) in the Dundas Element, Sheffield Element and Adamsfield–Jubilee Element, and there are probably one or more depositional hiatuses present in the overlying Gordon–Eldon Group sequences.

The most complete preserved sequence is in the Gordon–Olga–Denison Rivers area near the western margin of the southern half of the Tyennan Element. Here the Gordon Group includes a number of quartz sandstone units and is underlain by quartz sandstone, and includes faunal assemblages ranging from possibly OT8 (mid-Arenig) to OT18 (Late Caradoc) (Banks and Baillie, 1989). The preserved Eldon Group succession of fossiliferous marine sandstone, siltstone and limestone was deposited over a time-span comparable to that in the Dundas Element, and ranges in age from mid-Llandovery to Early Emsian, with a probable hiatus occupying much of the Ludlow.

At Bubs Hill, near the western margin of the central part of the Tyennan Element, a somewhat abbreviated (300 m thick) Gordon Group carbonate section overlies a very thin quartz sandstone unit which unconformably overlies Precambrian basement rocks. The Gordon Group sequence contains faunal assemblages ranging from OT15 (Early Caradoc) to OT19 (Caradoc-Ashgill boundary), but there is some thickness of section below the lowest fossiliferous horizon (Banks and Baillie, 1989). The carbonate sequence is overlain disconformably by quartzite, which is probably a correlative of the Crotty Quartzite, the basal formation of the Eldon Group.

The other main occurrence of strata of Ordovician–Early Devonian age in the Tyennan Element is in the Loddon River area in the centre of the Element, where they occupy a broad open NW-trending syncline partly faulted against metamorphic basement rocks. A basal siliceous conglomerate up to 180 m thick rests with angular unconformity on the Precambrian rocks, and is overlain by fine-grained well-sorted quartz sandstone with rare shelly fossils, pebble bands, intense bioturbation including vertical burrows, and herringbone cross-bedding which suggests tidal deposition (Banks and Baillie, 1989; Allen, 1983). The overlying Gordon Group limestone section is thin and contains a probable OT12 (Gisbornian) fauna (Allen, 1983). The limestone is overlain by ~340 m of cross-bedded quartz sandstone and ~200 m of slate, both of which contain marine macrofossils, and which are probable correlates of the Crotty Quartzite and Amber Slate of the Eldon Group (Banks and Baillie, 1989; Allen, 1983). Above this the section is faulted, but fossils recovered from other sandstones in the syncline indicate the presence of Wenlock-age rocks, and also the probable presence of correlates of the Florence Sandstone (Banks and Baillie, 1989).

Devonian deformation

The most significant Devonian structural effect on the Tyennan Element appears to have been during the later Devonian fold phase (Zeehan/Gormanston trend of Williams, 1979; Williams *et al.*, 1989; equivalent to D_4 of Seymour, 1980). During this event the Tyennan Element apparently yielded in a NW-trending narrow zone on a line roughly between Queenstown and the northern end of the Adamsfield–Jubilee Element, and behaved as two blocks. Within the zone separating these blocks, open upright NW-trending folds in the Precambrian rocks geometrically match similar structures in the overlying Ordovician to Devonian sedimentary sequences, and are therefore inferred to be Devonian structures (Williams *et al.*, 1989; Spry and Gee, 1964). In the Frenchmans Cap area, within the zone of yielding, unfolding of a Devonian fold indicates that prior to the deposition of the Gordon Group correlate, bedding and foliations in the Precambrian metasedimentary rocks were flat-lying, and recumbent isoclinal folds were present (Spry and Gee, 1964).

There are almost certainly other structural effects of the Devonian orogenesis within the Tyennan Element. For example, it seems possible that the latest structures in the Precambrian metamorphic rocks in the Strathgordon area are Devonian. These comprise D_5 minor folds and associated crenulation cleavage, and two later generations of minor conjugate kink sets, which have all been dispersed by the regional swing in structural trend from N-S to E-W through the area (Boulter, 1978, and in Turner, 1989). If this line of reasoning is correct then it would imply that the regional swing was also caused by a Devonian event, perhaps the folding of N-S trend which created the Florentine Synclinorium, and which probably correlates with the early fold phase (West Coast Range / Valentines Peak trend) of Williams (1979) and Williams *et al*. (1989). Furthermore, folds of this trend are apparently present in the Ordovician to Devonian strata in the Gordon–Olga Rivers area near the southwestern margin of the Tyennan Element, and have probably also affected the basement rocks in this area. Overall, it seems likely that the earlier N-trending Devonian fold phase may have had a more significant affect on the Tyennan Element (in particular on its southern half) than has generally been acknowledged.

Devonian granitoid rocks

Intruding the Precambrian metasedimentary rocks of the Tyennan Element near its northwestern margin is the Granite Tor Granite, which consists of coarse-grained biotite-muscovite granite with megacrysts of K-feldspar (McDougall and Leggo, 1965; McClenaghan *in* Williams *et al*., 1989). Interpretations of regional gravity data by Leaman (1986) showed that the Granite Tor Granite is very nearly connected, at relatively shallow depths, with the Heemskirk Granite on the west coast by a large buried granite spine oriented ENE-WSW beneath the Dundas Trough. Further gravity interpretation by Leaman and Richardson (1989) has shown that the exposed part of the Granite Tor Granite is only a small part of a large body which also extends some 35 km east and northeast of the present outcrop limits at depths of 2 km and less beneath the Precambrian metamorphic rocks. Minor Sn, W and Cu vein deposits are hosted by Proterozoic rocks overlying the eastern subsurface extension of the pluton (Leaman and Richardson, 1989).

The full range of K-Ar age determinations from the Granite Tor Granite is 344–375 Ma, including uncertainties (re-calculated data after McDougall and Leggo, 1965).

Two small granitoid outcrops on the south coast of the Tyennan Element show intrusive relationships with metasedimentary rocks. The bodies differ slightly in lithological characteristics, the Cox Bight Granite consisting of coarse-grained to sparsely porphyritic light-coloured biotite granite with minor muscovite, while the South West Cape Granite is a coarse-grained foliated biotite granite with phenocrysts of feldspar and biotite (McClenaghan, 1989). The bodies also show differing total ranges in isotopic ages, ca. 366–386 Ma for the Cox Bight Granite, and ca. 309–329 Ma for the South West Cape Granite (McClenaghan, 1989). However regional gravity interpretations indicate that these two outcrops are small emergents of a large body of granitoid which underlies significant portions of the southern parts of the Tyennan and Adamsfield–Jubilee Elements (Leaman and Richardson, 1992; see 2 km granite contour on Plate 2 herein).

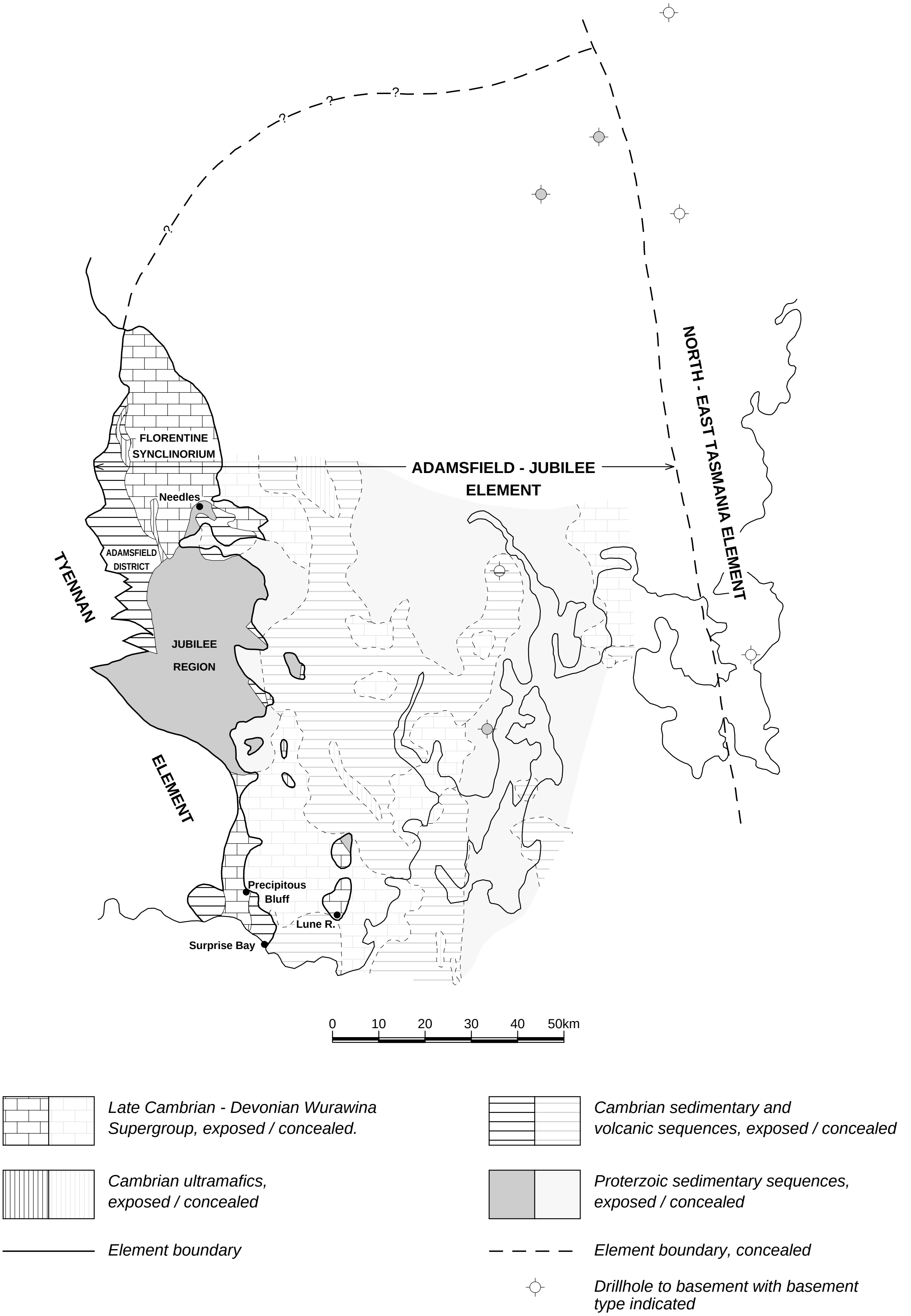

Figure 2.

Adamsfield–Jubilee Element, including inferred distribution of rock types beneath the Tasmania Basin in southeast Tasmania, from Leaman (1990b).

Adamsfield–Jubilee Element (CRC)

The Adamsfield–Jubilee Element is defined as the area of deformed but relatively unmetamorphosed Proterozoic and lower Palaeozoic rocks lying to the east of the Tyennan Element. It crops out in central southern Tasmania and southwards as a narrow discontinuous strip as far as the south coast (fig. 1). A much greater subsurface extent to the east — under rocks of the Tasmania Basin — is inferred on the basis of small inliers, drill-hole intersections, xenoliths in igneous rocks, and a regional gravity/magnetics interpretation (fig. 2) (Leaman, 1990*b*). The main exposed part of the Element (fig. 2) consists of a large southern area of mainly Proterozoic rocks (the Jubilee region); a northwestern area of mainly Cambrian or inferred Cambrian age (the Adamsfield district); and to the north, a large synclinorium (the Florentine Synclinorium) containing a succession of Late Cambrian to Early Devonian age, the Wurawina Supergroup.

The relationship of the Proterozoic sequences of the Adamsfield–Jubilee Element with those of the Tyennan Element is poorly known. An unconformity has been inferred on the basis of the generally lower metamorphic grade in the Adamsfield–Jubilee Element (Spry, 1962; Godfrey, 1970), but the broad lithological similarity and observed overlap in tectono-metamorphic grade suggest the relationship may be a metamorphic transition affecting essentially coeval sequences (Calver, 1989*a*).

A gravity/magnetics interpretation of southeast Tasmania suggests that the Tyennan Element (or similar lithologies) forms basement to the Adamsfield–Jubilee Element, at a general depth of about 5 km (Leaman, 1990*b*).

Late Proterozoic successions

The Proterozoic Clark Group and correlative sequences consists of a shallow-marine orthoquartzite unit, 1 to 2 km thick, which is underlain and overlain by similar thicknesses of pelitic rocks with minor siltstone and carbonate, the younger units locally with stromatolites and evaporite indicators. The base of the succession is unknown and the age is poorly constrained. The Clark Group may overlie rocks of the Tyennan Element unconformably or it may be broadly coeval with them. A transitional relationship is suggested by a similar range of lithologies to those in the Tyennan Element (and to the Rocky Cape Group of the Rocky Cape Element), and overlap in tectono-metamorphic rank with the Tyennan Element (Calver, 1989*a*; Calver *et al.*, 1990).

There is a gentle angular unconformity, locally a paraconformity, at the base of the Weld River Group that does not correspond to any major folding or cleavage-forming event in the older sequences (Calver, 1989*a*). The basal conglomerate and sandstone unit of the Weld River Group is lenticular and up to about 500 m thick. Then follows a 2 to 3 km thickness of shallow marine, oolitic and catagraphic dolostone. In faulted contact with this is a unit, 1–3 km thick, of dolostone with interlayered diamictite, mudstone and sandstone (Turner *et al.*, 1985; Calver, 1989*b*; Calver *et al.*, 1990). Dropstone laminites of probable glacial origin are associated with the diamictites and also, locally, with the basal siliciclastic rocks of the Weld River Group (Calver, 1989*b*; unpubl.).

The glacigene rocks suggest a Cryogenian age for the Weld River Group, and a broad lithostratigraphic correlation with the Success Creek Group of the Dundas Trough and the lower Togari Group of the Rocky Cape Element has been proposed (Calver, 1989*b*). However, a preliminary carbon isotope chemostratigraphic study offers some evidence for correlation of the Weld River Group with the Smithton Dolomite (upper Togari Group) of the Rocky Cape Element and hence a younger, Ediacarian (Neoproterozoic III) age (Calver, unpubl.).

The regional gravity/magnetic interpretation of Leaman (1990*b*) shows dense, non-magnetic rock forming basement to the Tasmania Basin, at least to depths of 3–5 km, in a broad north-south strip through Bruny Island to east of Hobart, and west of Hobart (fig. 2). These rocks are inferred to be Proterozoic, dolomitic sedimentary successions similar to those of the exposed Adamsfield–Jubilee Element (Leaman, 1990*b*).

Deformation of Proterozoic rocks

A deformation (D_1) resulting in northwest-plunging, upright to overturned (northeast-facing) major folds and associated axial-plane cleavage preceded deposition of the Middle Cambrian successions. S_1 decreases in intensity in a general northeasterly direction across the Jubilee region, reflecting a decline in metamorphic grade from lower greenschist facies to unmetamorphosed. Chlorite and ?andalusite porphyroblasts predate D_1 (Calver *et al.*, 1990). Duncan (1976) reported west-verging, isoclinal F_1 folds in the southern Jubilee region. Locally there are upright F_2 folds, coaxial with F_1, and an associated crenulation cleavage. D_1 and D_2 predate a ?Middle Cambrian conglomerate (possible Trial Ridge beds correlate) at Mt Bowes (Calver *et al.*, 1990).

Cambrian successions

The Ragged Basin Complex is an extensively disrupted (broken) formation in fault contact with other Cambrian and Proterozoic units, and consists of lithic sandstone, chert, red mudstone and minor mafic volcanic and shallow intrusive rocks (Turner, 1989*b*). The sandstone is of metamorphic and volcaniclastic derivation. Serpentinised ultramafic rocks with faulted margins consistently occur within, or in contact with, the Ragged Basin Complex. The ultramafic rocks are compositionally closely related to those of the Dundas Element; they formed as cumulates in shallow crustal magma chambers, and are of 'Alaskan-type' rather than ophiolitic in origin (Brown, 1986). Limited lode mining of osmiridium alloys has taken place in the Adamsfield Ultramafic Complex (Bottrill, 1989).

The Ragged Basin Complex is a lithologic correlate of the 'Cleveland–Waratah association' (Turner, 1989*b*), but the succession is unfossiliferous and local age constraints are poor. The Ragged Basin Complex, like the Cleveland–Waratah association, is interpreted as allochthonous, and was introduced, together with the ultramafic rocks, by westward overthrusting in the early Middle Cambrian (Crawford and Berry, 1992; Corbett, 1994).

Quartzitic conglomerate, including coarse alluvial fans and submarine fans, and turbiditic quartzwacke comprise the Trial Ridge beds and correlates (including the Island Road Formation and the Boyd River Formation; Turner, 1989*b*) of the Adamsfield district. The Trial Ridge beds and the Island Road Formation contain agnostoids of *Lejopyge laevigata* Zone (Boomerangian) age (Jago, 1979; Jago *et al*., 1989). The sequences overlie, with inferred unconformity, Proterozoic rocks of the Tyennan Element and are largely of metasedimentary (Tyennan Element) provenance, with rare ultramafic detritus (Corbett, 1970; Turner, 1990*a*, p.50). There are several structurally separated, turbiditic quartzwacke-lithicwacke-conglomerate successions east and south of the Adamsfield district as far as the south coast (Plate 2; fig. 2). These are locally of dolomitic (Weld River Group), metamorphic and mafic-volcaniclastic provenance (e.g. Calver, 1989*b*). No fossils are known from these successions and they are here tentatively correlated with the Middle Cambrian fossiliferous rocks on broad lithostratigraphic grounds. At one locality there is an inferred unconformity upon deformed (D_1 + D_2) Weld River Group (Calver *et al*., 1990). On the south coast, quartzitic conglomerate unconformably overlying Tyennan rocks passes up into a turbiditic lithicwacke succession; ultramafic detritus is present (Findlay, in press).

The gravity/magnetics interpretation of Leaman (1990*b*) shows a thick (several kilometres) trough of dense, magnetic rock — inferred to be a Cambrian mafic volcanics-dominated succession — extending northwest-southeast beneath the cover rocks of the Tasmania Basin (fig. 2). A smaller trough extends northwards to Hobart, where a drillhole intersected altered ?Cambrian volcanic rocks beneath the Tasmania Basin rocks (Everard, 1976).

Late Middle to early Late Cambrian deformation

A significant period of folding and faulting preceded deposition of the Wurawina Supergroup, and is shown in simplified form on Plate 1. Two folding episodes; an earlier west to northwest-trending phase and a later, open, northeast-trending phase; are seen in the Adamsfield district (Turner, 1989*b*; 1990*b*). A major NW-trending fold at The Needles is of probably similar age, and has been tightened during Devonian deformation (Brown *et al*., 1989; Calver, unpubl.). In the Adamsfield district, many faults dip east and facing directions are predominantly east to northeast (Turner, 1990*b*; Turner *et al*., 1985; Brown *et al*., 1982). Folds in the Jubilee region face north and Proterozoic and Cambrian rocks are structurally interleaved through faulting. An interpretation involving an imbricate series of east and north-dipping thrust slices, with transport from the east and north in the Adamsfield district and Jubilee region respectively, is consistent with the broad structural pattern. Other types of fault may be present. Thrusting appears to have largely post-dated folding in the Adamsfield district (Turner, 1989*b*) but it predated folding at The Needles (Calver, unpubl.).

Wurawina Supergroup

A thick, essentially conformable, Late Cambrian to Devonian succession, the Wurawina Supergroup, unconformably overlies older rocks. The Supergroup is divided into the Denison, Gordon and Tiger Range Groups. On the Denison Range the Denison Group consists of (from the base):–

- the Singing Creek Formation — 700 m of predominantly quartzwacke turbidite deposited as a submarine fan complex, with Idamean faunas;
- the Great Dome Sandstone — 500 m of quartz sandstone of shallow-marine to fluvial palaeoenvironments with late Cambrian fossils;
- the Reeds Conglomerate — up to 1560 m of fluvial to shallow-marine conglomerate and sandstone, composed largely of quartzitic Tyennan detritus;
- and finally the Squirrel Creek Formation (a correlate of the Florentine Valley Formation), shallow-marine sandstone and calcareous siltstone, ca. 700 m thick, with a rich early Ordovician fauna (Corbett and Banks, 1974; Corbett, 1975*a*; Banks and Burrett, 1980; Stait and Laurie, 1980; Jago, 1987).

The Denison Group onlaps higher basement to the east, where lateral equivalents of the uppermost Reeds Conglomerate unconformably overlie older rocks north of The Needles (Turner *et al*., 1985).

Palaeoplacer deposits of osmiridium are present in basal Wurawina Supergroup rocks near Adamsfield (Ford, 1981).

The Gordon Group comprises predominantly shallow-water platform carbonates conformably and transitionally overlying the Denison Group, and (in the Jubilee Element) consists of (from the base):–

- the Karmberg Limestone, about 500 m thick, of nodular, silty and cherty, subtidal limestone;
- the Cashions Creek Limestone, 150 m thick, of oncolitic calcarenite probably deposited as offshore shoals or bars;
- the Benjamin Limestone, 1200 m of predominantly peritidal, dolomitic micrite and calcarenite;
- and the Arndell Sandstone, 250 m of shallow marine siltstone and sandstone (Corbett and Banks, 1974; Baillie, 1979; Banks and Baillie, 1989).

The Florentine Valley Mudstone contains fossil assemblages (OT1–OT7) that can be dated as Middle Tremadocian (OT3) to Chewtonian or Early Castlemanian (OT7). The Karmberg Limestone contains faunas (OT7 and 8) of Mesial to Late Arenigian (OT7) and Whiterockian (OT8) age. The Cashions Creek Limestone is Late Whiterockian (Chazyan). The Benjamin Limestone includes faunas of Gisbornian (OT12–14) to Bolindian (OT19) age. The Arndell Sandstone contains faunas of probably Ashgillian (OT20) to Early Llandovery age (Baillie, 1979; Banks and Burrett, 1980).

Lithological correlatives of the Karmberg, Cashions Creek and Benjamin Limestones can be recognised at isolated localities as far south as Lune River (e.g. Summons, 1981). However west of Lune River, at Precipitous Bluff, a biostratigraphic correlate of the Benjamin Limestone consists of platform-edge coralline calcarenite; while on the south coast at Surprise Bay, a correlate (the Shoemaker beds) consists of deep-water (ca. 300 m), graptolitic shale and carbonate turbidite. A distinct platform edge trending east-southeast through Precipitous Bluff thus delimits the widespread shallow-water platform interior facies of the Gordon Group (Burrett *et al.*, 1984).

The Tiger Range Group — a correlate of the Eldon Group of western Tasmania — consists of shallow marine siliciclastic rocks conformably overlying the Gordon Group. Constituent formations are (from the base):–

- the Gell Quartzite (130 m thick);
- the Richea Siltstone (220 m, with a Late Llandovery age fauna and a higher horizon with Middle or Late Wenlock age fossils);
- the Currawong Quartzite (150 m, with a late Silurian fauna);
- and the ?Lower Devonian McLeod Creek Formation, at least 400 m of poorly fossiliferous fine-grained sandstone and shale.

The top of the succession has been removed by erosion (Baillie, 1979; Baillie, 1989*a*).

Devonian deformation

In the Devonian, the Wurawina Supergroup was folded into a large, gently north-plunging synclinorium (Florentine Synclinorium), with a related subvertical cleavage of north to NNW trend in pelitic lithologies. Locally developed, weak north to northwest-trending crenulation cleavages in the Proterozoic and Cambrian rocks of the Adamsfield–Jubilee Element may be Devonian in age (Turner, 1989*b*). Conodont alteration indices in the Florentine Synclinorium suggest maximum burial temperatures of 110–200°C (Burrett, 1984). The effects of the Devonian Orogeny appear to have been weaker in southern Tasmania, as folding in the Wurawina Supergroup around Ida Bay is very open and conodont alteration indices suggest maximum burial temperatures of less than 100°C (which may have been attained during the Late Cretaceous rather than the Devonian: Sharples and Klootwijk, 1981; and see later section).

Tectonic Synthesis of Western Tasmania

The range of new data and interpretations which have become available over the last few years has also led to attempts at a revised overall synthesis of the tectonic history of western Tasmania, and one of the more comprehensive recent compilations is that by Berry (1994) which is summarised and commented on below:–

Event 1 Late Proterozoic sedimentation

- Shallow water sedimentation on the eastern margin of the East Antarctic shield.

Event 2 Penguin Orogeny ca. 750 Ma

- Polyphase (D_1–D_4) deformation on King Island, with syn-kinematic granitoid intrusion (760 ± 12 Ma).
- Early open upright folding of Rocky Cape Group.
- Early recumbent folding in Burnie and Oonah Formations.

Event 3 Passive margin formation ca. 600 Ma

- Two-stage rift phase: early shallow water sedimentation followed by a broad zone of extensive tholeiitic volcanism with a nominal age of 600 Ma to early Cambrian.

Event 4 Arc-continent collision late Early to early Middle Cambrian (525–510 Ma)

- Arc-continent collision with obduction resulting in tectonic emplacement of allochthonous elements including:–
  - mafic/ultramafic complexes
  - Forth Metamorphic Complex
  - Badger Head Metamorphic Complex
  - eclogites in the Tyennan Massif
  - Bowry Formation in the Arthur Lineament
- Major phase of thrusting to the south.

Event 5 Delamerian Orogeny 510–490 Ma

- Middle Middle Cambrian E-W extension, eruption of Mt Read Volcanics.
- Late Middle to early Late Cambrian N-S compression producing E-W trending folds (Fossey Mountains Trough, Adamsfield, Bathurst Harbour area).
- Late Cambrian E-W compression:
  - Earlier extensional faults re-activated as reverse faults (Henty Fault).
  - Major reverse faults and upright open N-S trending folds formed in western Tasmania.
  - Major uplift of the Tyennan block and deposition of syn-orogenic sediments (Owen Group).

Event 6 Sag phase Ordovician–Silurian

- Shallow water marine platform sedimentation.

Event 7 Depositional hiatus Silurian

- Recognised as a deformation phase in eastern Victoria and further north but only represented as a hiatus in deposition in Tasmania.

Event 8 Deposition Early Devonian

- Renewal of shelf sedimentation.

Event 9 Orogenesis Middle Devonian

- Fold geometry controlled by existing Cambrian fold trends which were tightened during the Devonian (e.g. tightening of Cambrian E-W folds in the Fossey Mountains Trough, tightening of Cambrian N-S folds in Dundas Trough with development of NNW-trending Devonian cleavage).
- NNE-trending Devonian folds in northern Dundas Trough controlled by re-activation of Henty Fault.
- Most western Tasmanian granites intruded syn- to post-kinematic with respect to the NNW-trending cleavage and associated folding and faulting.
- Subsequent NE-SW to NNE-SSW compression produced WNW-trending folds and thrusts in the Dundas Trough, and NW-trending thrusts and associated folds in the Sheffield Element. In the Dundas Trough an associated phase of brittle wrench faults was dominated by NNE-striking sinistral movement on the Henty Fault.

Comments arising from this synthesis are as follows:–

- Berry uses the term "Penguin Orogeny" for his ca. 750 Ma Event 2. However, the Penguin Orogeny has always been considered responsible for the main deformation and metamorphism in the Arthur Metamorphic Complex (Gee, 1967), and this event is now dated at ca. 500 Ma (Turner, 1993*b*) and is known to have post-dated the Ahrberg Group, and the Togari Group in the Smithton Synclinorium. Thus, Turner *et al.* (1994) consider the Penguin Orogeny to be equivalent to the Delamerian Orogeny (Event 5 of Berry, 1994).

- The placement of the early recumbent folding in the Burnie and Oonah Formations in Event 2 is apparently inconsistent with the various K-Ar age data pertaining to the age of deposition of these units. However, the K-Ar data may be minimum ages, and assuming depositional ages only ~10% older than those suggested by the K-Ar data produces a logical correlation of the early recumbent folding with the main tectono-metamorphic event on King Island.

- The age of Event 4 (arc-continent collision, 525–510 Ma) appears to be inconsistent with the 510 ± 6 Ma U-Pb age of tonalite in the Heazlewood Ultramafic Complex, assuming that this represents the age of crystallisation of the whole complex, as the crystallisation age must pre-date the tectonic emplacement. Turner *et al.* (1994) and Corbett (1994) see Events 4 and 5 of Berry (1994) as a continuum beginning at ca. 510 Ma, rather than two discrete events.

- The identification of the Bowry Formation in the Arthur Metamorphic Complex as an allochthonous element emplaced during Event 4 is inconsistent with the interpretations of Turner *et al.* (1991, 1992), in which the Bowry Formation is considered to be a metamorphic equivalent of known autochthonous sequences in the Rocky Cape Element.

- The existence of a depositional hiatus some time in the Silurian in Tasmania (Event 7) is still a matter of argument, but the apparent lack of sequences of Ludlow age on the time-space diagram (Plate 1) could indicate a hiatus.

Northeast Tasmania Element (CRC)

The Northeast Tasmania Element is characterised by lower Palaeozoic turbiditic rocks (the Mathinna Group) and an absence of known Proterozoic rocks, contrasting with the post-Cambrian shelf deposits and extensive Proterozoic rocks to the west. The western boundary of the Northeast Tasmania Element is not exposed but the different lower Palaeozoic successions are separated by only a 20 km wide cover of younger rocks in the Tamar Valley (fig. 1). There is also a contrast in structural style across the Tamar Valley, with Devonian folds to the west (Sheffield Element) verging southwest while those in the Northeast Tasmania Element verge northeast (Williams, 1979). A major crustal fracture, the Tamar Fracture System, has been postulated (Williams, 1979) to trend southeast along the axis of the Tamar Valley and then SSE under the Tasmania Basin. Both sinistral (Williams, 1979) and dextral (Baillie, 1985) wrench movements have been suggested. However, there is no compelling geophysical evidence for such a major crustal fracture (Leaman, 1994). A major gravity gradient lies not at the postulated site of the fracture but some 40 km west of the Tamar Valley, in part following the line of the Tiers Fault (Plate 2; Leaman, 1994). Some recent interpretations favour a shallow thrust contact, with the Northeast Tasmania Element thrusted over essentially western Tasmania-style basement (Powell and Baillie, 1992; Keele *et al.*, 1994; Roache, 1994). A broad magnetic anomaly west of Bridport (Plate 2) may derive from Cambrian mafic and ultramafic rocks underlying Palaeozoic rocks of the northeast Tasmania element at a depth of several kilometres (Roache, 1994).

Mathinna Group

The Mathinna Group, whose base is unknown, consists of (from oldest to youngest):–

- quartzose turbiditic sandstone (Stony Head Sandstone, ca. 1 km thick);

- black pelite (Turquoise Bluff Slate, ca. 1–2 km);

- then an increasingly sand-rich succession of quartzose turbidites of sublithic (Bellingham Formation, ca. 2 km), then feldspatholithic composition (Sidling Sandstone, ca. 2 km) (Powell and Baillie, 1992; Powell *et al.*, 1993; Keele, 1994).

Age-diagnostic fossils are extremely rare. Late Arenigian graptolites occur in the Turquoise Bluff Slate (Banks and Smith, 1968). Graptolites of late Ludlow age occur at Golden Ridge (Rickards *et al.*, 1993) in a sequence that may belong to the Turquoise Bluff Slate (Keele, 1994). The Sidling Sandstone contains early Devonian plant remains, and graptolites and shelly fossils of Pragian age (Rickards and Banks, 1979; Alberti *in* Banks and Baillie, 1989, p.236). Despite the sparse age control the succession is thought to be conformable, as in the probably-related Melbourne Trough (Powell *et al.*, 1993). There is a regional younging of the sequence from west to east.

Early Devonian deformation

Late Early Devonian deformation (D_1) produced NNW-trending, upright to locally east-facing overturned folds (Williams, 1978; Turner, 1980) accompanied by low-grade regional metamorphism. D_1 preceded the St Marys Porphyrite, a dacitic ash-flow pile dated at 388 ± 1 Ma (Turner *et al.*, 1986); and preceded the emplacement of related granodiorite plutons as old as 395 Ma (Cocker, 1982).

Devonian granitoids

The Devonian granitoids (395–368 Ma) include three large composite batholiths (the Scottsdale, Blue Tier and Eddystone Batholiths), shown by geophysical modelling to be all part of a large northerly-trending elongate subsurface mass that underlies most of the Northeast Tasmania Element (Leaman and Richardson, 1992). A major north-trending gravity gradient (Plate 2) reflects the steep western boundary of the less-dense granites and adamellites (Leaman and Richardson, 1992). Emplacement was at high crustal level (6 km), and contact metamorphic aureoles are narrow (Cocker, 1982; McClenaghan, 1985). The St Marys Porphyrite, on the east coast, is extrusive (Turner *et al.*, 1986). There are four granitoid types; granodiorite (I-type), biotite adamellite/granite, biotite-garnet adamellite (S-type), and alkali-feldspar granite (McClenaghan, 1985; McClenaghan *in* Williams *et al.*, 1989).

In the relatively well-studied Blue Tier Batholith the main granodiorite plutons are the oldest, and are intruded by adamellite (Mt Pearson Pluton), followed by granodioritic dykes that are comagmatic with, and in part feeders to, the St Marys Porphyrite. A large number of K-Ar and Rb-Sr dates indicate these phases are 388–398 Ma but their age differences are not resolvable radiometrically (McClenaghan and Higgins, 1993). A second phase of adamellite intrusion occurred about 15 Ma later, followed by alkali-feldspar granites (364–374 Ma: Cocker, 1982; McClenaghan and Higgins, 1993). Generally, field relationships and radiometric dating suggest a similar succession of phases elsewhere in northeast Tasmania (Gee and Groves, 1971; McClenaghan, 1985). The alkali-feldspar granites were derived from adamellite magma by crystal fractionation (McClenaghan, 1985). Dolerite dykes intrude all major granite types but probably overlap in age with quartz-feldspar porphyry dykes thought to be genetically related to the granitoids (McClenaghan, 1985).

The proportion of I-type granitoids decreases to the east, and unequivocal S-type granites are restricted to the Eddystone Batholith in the far northeast. $^{87}Sr/^{86}Sr$ systematics suggest protolith ages of 800–1700 Ma for S-type, and 1250–1400 Ma for I-type granitoids (Cocker, 1982).

Devonian mineralisation

Auriferous quartz reefs are widespread in the Mathinna Group. Many deposits, including the largest known (the New Golden Gate deposit at Mathinna) lie within a 70 × 6 km NNW-trending corridor — the Mathinna-Alberton Gold Lineament — lying between the Scottsdale and Blue Tier Batholiths (Keele, 1994). Gold-bearing quartz veins within this zone were deposited by fluids of deep-seated regional metamorphic origin (Taheri and Bottrill, 1994), and veins were genetically associated with wrench faulting associated with D_2 (see below) (Keele, 1994). Some other gold deposits (e.g. Lisle, Golconda) appear to have a close genetic association with hornblende granodiorites (Klominsky and Groves, 1970).

The marginal, altered phase of the Mt Pearson biotite adamellite-granite pluton is genetically associated with the zoned Scamander field, with W-Mo, Sn-Cu, and Ag-Pb-Zn vein deposits arrayed successively further away from the pluton (Groves, 1972).

Disseminated cassiterite deposits are found within greisenised alkali-feldspar granite in the Blue Tier Batholith (Groves, 1977), and dilatational quartz-wolframite vein deposits occur in Mathinna Group country rock over altered granite cupolas (e.g. Aberfoyle, Storeys Creek mines). The absence of skarn and replacement-type Devonian deposits in the Northeast Tasmania Element reflects the absence of carbonates in the country rocks (Collins and Williams, 1986).

Mid-Devonian to Carboniferous deformation

There is locally evidence for at least two phases of deformation during and after granitoid emplacement (e.g. Turner *in* McClenaghan *et al.*, 1982). Granodiorites and granite/adamellites in the Blue Tier and Scottsdale Batholiths have a predominantly NNW-trending tectonic foliation, and similarly orientated crenulation cleavages are locally present in the Mathinna Group. Upright NNW-trending F_2 folds are seen in the west (Bridport–Georgetown area; Powell and Baillie, 1992). Powell and Baillie (1992) proposed that WSW-directed transport of the Mathinna Group over a shallowly east-dipping sole thrust was associated with these folds, and that this thrust is related to a system of southwest-verging imbricate thrusts in the Beaconsfield area of the Sheffield Element. Keele (1994) recorded north to northeast-trending D_2 folds and fracture cleavage in the Mathinna area, and associated north to WNW-trending dextral wrench faults that were conduits for Au-bearing fluids.

A large north-trending fold is present in the St Marys Porphyrite, but a north-trending tectonic foliation in the nearby Piccaninny Creek granite belongs to a younger event (Turner *et al.*, 1986).

Late Devonian to Carboniferous mega-kinking involved rotation of domains up to 9 km wide during overall NNW-SSE bulk shortening of 4–5% at shallow (~4 km) crustal levels (Goscombe *et al.*, 1994). This event post-dates granitoid emplacement and predates the late Carboniferous–Triassic Parmeener Supergroup. Mega-kinking in northeast Tasmania is correlated with mid-Carboniferous mega-kinking in the Lachlan Fold Belt (Powell *et al.*, 1985; Goscombe *et al.*, 1994).

Tasmania Basin (CRC)

The Tasmania Basin contains a succession of predominantly flat-lying sedimentary rocks of Late Carboniferous to Late Triassic age known as the Parmeener Supergroup, and thick sheets and sills of Jurassic dolerite that presently occupy most of the outcrop area of the basin. The total maximum thickness of the succession is ca. 1.5 km, and the Tasmania Basin covers most of central and eastern Tasmania and overlaps most of the older Elements (fig. 1). The present basin limits are erosional, not depositional, and the original basin was probably considerably larger. Proterozoic metamorphic rocks of the Tyennan Element and granitic rocks of the Northeast Tasmania Element are basement highs, and the main depocentre was situated along the axis of the postulated Tamar Fracture System. The thin and incomplete succession in far northeast Tasmania suggests proximity to a basin margin in this area.

Provincialism of the cold-climate Late Carboniferous to Permian Gondwanan biotas makes difficult any correlation with the international standard, so placement of units in Plate 1 must be regarded as approximate. Correlation within Tasmania, however, is well established, with a detailed local biostratigraphic framework based on marine macroinvertebrates in the Lower Parmeener Supergroup (Clarke and Banks, 1975; Clarke and Farmer, 1976; Clarke, 1989) and on palynomorphs (e.g. Truswell, 1978; Truswell *in* Calver *et al.*, 1984; Forsyth, 1989*a*). Stage names of the Late Carboniferous to Permian Rekunian Series, defined within Tasmania (Clarke and Farmer, 1976), are indicated on Plate 1.

Lower Parmeener Supergroup

The Lower Parmeener Supergroup, consisting largely of glacigene and shallow-water glaciomarine rocks, rests on a landscape unconformity with a relief of about 1000 metres.

In most areas the succession begins with a unit of tillite, diamictite and minor rhythmitic claystone that reaches a maximum thickness of 500 m near Wynyard (Wynyard Tillite) and 580 m at Cygnet (Truro Tillite) (Clarke and Banks, 1975; Hand, 1993). Ice flowed eastwards from a source west of Tasmania. Tillite is absent from the major highs of the northern Tyennan Element and parts of the Northeast Tasmania Element (Hand, 1993). This part of the succession is associated with Stage 1 and early Stage 2 microfloras (Truswell, 1978) of Late Carboniferous age (Kemp *et al.*, 1977; Balme, 1980). (Alternatively, Stage 2 may be partly or wholly Permian; e.g. Archbold, 1982).

The basal glacigene rocks are succeeded by a sequence of carbonaceous, pyritic siltstone (the Woody Island Siltstone in southern Tasmania; the Quamby Mudstone in the north) passing up into richly fossiliferous siltstone, sandstone and minor limestone (the Bundella Formation and correlates, including the Golden Valley Group in the north). Thickest developments are found along the axis of the main depocentre (>400 m) but elsewhere the thickness is typically 150–180 metres. Near the base of this sequence a bed of oil shale, usually about 2 m thick, is widespread in the north and in a small sub-basin at Douglas River in the east (Calver *et al.*, 1984). The oil shale ('tasmanite') is immature in these areas, with a maturity corresponding to a vitrinite reflectance of 0.5%. The carbonaceous siltstone is typically around 1% in total organic carbon (Domack *et al.*, 1993), and passes laterally into littoral and sub-littoral sandstone and conglomerate where the succession onlaps basement highs in eastern and northern Tasmania. Ice-rafted dropstones are common in the Bundella Formation and correlates. The fauna of this succession define the Tamarian Stage, and may be broadly correlated with the Allandale and Rutherford Formations of New South Wales (Clarke, 1989).

The sequence so far described is nearly everywhere succeeded by a thin (20–40 m) unit of fluvial to paralic origin (the 'lower freshwater sequence', including the Faulkner Group and its lateral equivalents, the Liffey Group and the Mersey Coal Measures). Cross-bedded quartzarenite is characteristic, but siltstone and carbonaceous mudstone is dominant in the southeast near what is inferred to have been the seaward margin of the coastal plain. A thin marine intercalation is present in southeast and central Tasmania. Lateral equivalents of the 'lower freshwater sequence' are wholly marine in the far southeast (Farmer, 1985). Thin (<1 m) coal seams are present in the north and northeast, around the landward margin of the basin, with vitrinite reflectances of 0.31–0.58% having been measured. The 'lower freshwater sequence' yields a Substage 3b microflora (and is thus significantly older than the Greta Coal Measures of NSW). The lower part of the sequence is older (Substage 3a = late Tamarian) at Douglas River (Truswell *in* Calver *et al.*, 1984).

Where the 'lower freshwater sequence' onlaps basement in northeast Tasmania, it locally contains small palaeoplacers of tin and gold (Twelvetrees, 1907; Reid and Henderson, 1929; Finucane, 1932).

In southeast and east Tasmania the 'lower freshwater sequence' is succeeded by marine calcareous siltstone (Nassau Formation and correlates), then bioclastic (bryozoal-crinoidal) limestone (Berriedale Limestone and correlates), totalling about 100 m in maximum thickness. Rocks of this age (mid to late Bernacchian) are absent from most of northern and western Tasmania, which were probably emergent at this time (Clarke and Banks, 1975). Thin metabentonite layers (distal ashfalls?) are locally present in the Berriedale Limestone and correlates. Lonestones are present and cold-water conditions persisted throughout deposition of the carbonate rocks (Rao and Green, 1982).

Gentle uplift in southern Tasmania resulted in local erosion down to the level of the Bundella Mudstone, and a more widespread disconformity at the top of the Berriedale Limestone, before renewed marine sedimentation of the lower Lymingtonian Stage (Farmer, 1985). These rocks (Malbina Formation, Deep Bay Formation and correlates) are dominantly fossiliferous siltstone and poorly-sorted sandstone in which ice-rafted lonestones are common. Thicknesses reach a maximum of 180 m along the axis of the old Tamar Fracture System but thin rapidly to the west and northeast, the latter area being marked by shallow-water glauconitic sandstone, in places only a few metres thick.

The upper Lymingtonian — embracing the uppermost Malbina Formation, the Risdon Sandstone and the Abels Bay Formation — is characterised by a predominance of bioturbated, unfossiliferous siltstone of probable brackish-estuarine environment. Minor fossiliferous intervals that represent brief, fully-marine incursions are restricted to southern Tasmania. Invertebrate affinities suggest a Kazanian age (Clarke, 1987). There are two or more thin (a few metres) coarse-grained units, probably representing brief regressions (Risdon Sandstone, Blackwood Conglomerate). Felsic volcanic ash is locally present in the upper part of the succession. This succession is thickest (180 m) in the southeast, and thins to the northeast and northwest. Basement highs, small in area, in far eastern Tasmania (Friendly Beaches, Maria Island) were finally buried at this time. At the top of this succession there is a transition over several metres into the Upper Parmeener Supergroup.

Upper Parmeener Supergroup

The Upper Parmeener Supergroup, a non-marine succession of Late Permian to Late Triassic age, has been divided into four lithological units (Forsyth, 1989*a*).

Unit 1 corresponds to the Late Permian Cygnet Coal Measures (*sensu* Farmer, 1985) and correlates. The unit consists of well-sorted, cross-bedded sandstone — feldspathic in southern Tasmania but tending to be more quartz-rich and micaceous in the north — and carbonaceous siltstone and mudstone. Thin coal seams are present in the far southeast (Cygnet–Bruny Island) and in the northwest (Mt Ossa). Palaeocurrents indicate low-sinuosity, east-flowing rivers. The unit is thickest in the west (108 m) but in most other areas is around 50 m thick, but wedges out to the northeast at least partly because of erosion prior to the deposition of Unit 2. The age of the base of the unit is poorly constrained, but a microflora at the top of the unit in the southern Midlands has been assigned to the upper *P. microcorpus* zone, indicative of a latest Chhidruan (to early Greisbachian?) age (Forsyth, 1984).

Unit 2 is 200–300 m thick but thinner in the northeast, and consists predominantly of well-sorted quartzarenite. There is a widespread lutite-dominated interval, 20–60 m thick, at the top of unit 2, and 35 m of lutite in the middle of the unit at Hobart. Pebbly horizons are common in the northwest. The sequence was deposited from low-sinuosity rivers flowing east or southeast. The unit is widely distributed and may have originally extended across western Tasmania, but wedges out against high basement in the northeast. A microflora near the middle of the unit is assigned a Greisbachian to mid-Smithian age, while microfloras and macrofloras near the top are mid-Smithian to pre-Anisian (Forsyth, 1989*a*).

Unit 3 begins with an impersistent quartz granule sandstone, conglomeratic in places and around 5 m thick, deposited by NNW-flowing palaeocurrents. This is overlain by interbedded quartz sandstone, lithic sandstone and lutite. Microflora suggest an earliest Middle Triassic age. Then follows an interval, about 80 m thick in the southern Midlands, of interbedded lithic sandstone and lutite. Lithic grains are mostly of fine-grained felsic volcanic type. Macro and microflora suggest a late Anisian age. Unit 3 concludes with an interval, 100 m thick in the Midlands but thinner elsewhere, of quartz sandstone interbedded with carbonaceous lutite and minor lithic and feldspathic sandstone. Thin coal seams are locally present. In far eastern Tasmania, this interval overlaps all older units of the Upper Parmeener Supergroup to rest directly on eroded Lower Parmeener Supergroup. Near St Marys, the lower of two alkali-olivine basalt flows, each up to 30 m thick, has been dated at 233 ± 5 Ma (whole-rock K-Ar minimum age; Calver and Castleden, 1981). Microflora suggest that this interval is Ladinian, or late Anisian to Ladinian.

Unit 4 is predominantly lithic sandstone with minor lutite and coal, with the thickest preserved sections being in the northeast (maximum ca. 350 m). Unit 4 contains all of Tasmania's economic coal reserves, these being mostly in the northeast (Bacon, 1991), where eight or more seams or groups of seams are present. The base of Unit 4 is transitional on Unit 3 and is diachronous, being slightly older in the northeast than elsewhere. The lithic sandstone is largely of intermediate to felsic volcanic provenance, and there are rare, thin (<1 m) felsic tuff horizons high in Unit 4 (Bacon and Everard, 1981). There are also rare conglomeratic horizons at similar levels which include common rhyolitic clasts. A calc-alkaline volcanic source, probably to the east of Tasmania, is indicated. Unit 4 is characterised, almost throughout, by Carnian microfloras. The youngest preserved part of Unit 4, with Norian microfloras, is a lutite-dominated interval about 100 m thick near the Douglas River. An ashfall tuff at the top of the Carnian interval contains biotite dated at 214 ± 1 Ma (Bacon and Green, 1984), in close agreement with the Carnian–Norian boundary age of 215 Ma proposed by Webb (1981) but significantly different from the boundary age of 220 Ma given in both Odin and Letolle (1982) and in the timescale used in Plate 1 (Jones, 1994).

Post-Carboniferous deformation, western Tasmania

An outlier of Late Carboniferous tillite near Zeehan (the Zeehan Tillite) displays gentle northwest-trending folds and an associated, pervasive, subvertical axial-surface cleavage (Goscombe, 1991). The minimum age of this apparently areally-restricted deformation event is unconstrained.

Jurassic mudstone

Mudstone underlying basalt in the Lune River area of southern Tasmania contains fossil wood and leaves which suggest a 'mid-Mesozoic' age (Tidwell *et al*., 1987). The mudstone is probably Jurassic as the basalt is, on strong petrographic and geochemical evidence, comagmatic with extensive Middle Jurassic dolerite intrusions (see below).

Jurassic dolerite

A large volume of tholeiitic dolerite was intruded into the Tasmanian crust in the Middle Jurassic, mainly as sheets and sills in the flat-lying sediments of the Tasmania Basin. These sheets and sills are typically 400–500 m thick, and Jurassic dolerite is currently exposed over most of the area of the Tasmania Basin.

Sills, unroofed by erosion, are underlain by rocks as young as Norian (see above). A younger cover of unknown thickness, now entirely removed by erosion, can be inferred. Attempts to constrain the thickness of the cover by inferring burial depths are inconclusive (Hergt *et al*., 1989).

Intrusion of the dolerite was probably related to tensional stresses that heralded the eventual rifting of Australia and Antarctica in the mid-Cretaceous (Veevers and Eittreim, 1988). Studies of the chilled contacts show that the quartz-tholeiite magma was remarkably homogeneous on emplacement, but marked differentiation, controlled by fractional crystallisation, occurred within the thicker sheets and larger dyke-like bodies (Edwards, 1942; McDougall, 1962, 1964). Geochemical and isotopic data suggest that the magma is derived from the continental crust rather than the upper mantle, whence such large volumes of magma might be expected to have been derived (Hergt *et al*., 1989).

An average of widespread whole-rock and feldspar K-Ar ages is 174.5 ± 8.0 Ma (Schmidt and McDougall, 1977; recalculated according to Steiger and Jäger, 1977). The resolution of the K-Ar system is such that it is only possible to state that emplacement probably occurred over an interval of less than 20 million years (Schmidt and McDougall, 1977).

Limited areas of basalt in a graben at Lune River in southern Tasmania are the only known extrusive occurrences of this phase of magmatism. The basalt is associated with sediments containing plant fossils of probable Jurassic age, and is comagmatic with the dolerite on petrographic, geochemical and isotopic evidence (Hergt *et al*., 1989).

Late Mesozoic–Cainozoic history (CRC)

Introduction

Continental extension (rifting) of Australia–Antarctica began in the mid-Jurassic (160 Ma) and separation occurred in the mid-Cretaceous (95 ± 5 Ma) (Veevers and Eittreim, 1988). Rifting along the eastern margin of Tasmania probably began in the mid-Cretaceous, with the opening of the Tasman Sea starting in the Late Cretaceous (80 Ma) (Weissel and Hayes, 1977).

A number of extensional sedimentary basins associated with continental breakup — the Bass, Sorell, Otway, Durroon and Gippsland Basins — lie in offshore Tasmanian waters, and contain thick (several kilometre) fills of late Mesozoic–Cainozoic age. Onshore extensions of the Bass and Sorell Basins, such as the Tamar Graben and the Macquarie Harbour Graben, contain thinner successions of continental derivation, and there are widespread, onshore surficial Cainozoic sediments and mafic volcanic rocks. The geological history of the southern Bass Basin, the Durroon Basin and the Sorell Basin is included, together with the onshore deposits, in Plate 1.

Apatite fission-track analysis in northeastern Tasmania indicates a thermal event and/or uplift at 90–110 Ma (O'Sullivan, 1994) and uplift in the Durroon Basin at about 95 Ma (Duddy, 1992), probably related to Australia–Antarctica rifting. Apatite fission track dates of 70–80 Ma in northeastern Tasmania coincide with the Tasman rifting event (O'Sullivan, 1994).

Widespread late Mesozoic to Tertiary normal faulting, chiefly of meridional to northwest trends, is also a consequence of continental rifting. Major faulting had essentially ceased in the Durroon and Bass Basins by the middle Eocene, with minor reactivation in the late Tertiary, but major faulting continued into the Oligocene in the Sorell Basin. Tertiary basalts in general post-date faulting in onshore Tasmania.

Latest Jurassic–Cretaceous igneous activity

On King Island a biotite lamprophyre dyke has yielded a latest Jurassic age of 143 Ma (McDougall and Leggo, 1965). A similar dyke contains xenoliths of two pyroxene-garnet granulite, which indicate a metamorphic Precambrian basement and a possible link with granulite-facies basement rocks exposed in eastern Antarctica (Brown, 1989*b*; Waldron *et al*., 1993).

Volcaniclastic sandstone in the Early Cretaceous Otway Group of the Otway, Bass and Durroon Basins (see below) are derived from coeval vulcanism of alkaline intermediate character. Detrital minerals in outcropping Otway Group rocks in Victoria have yielded dates between 126 and 103 Ma (Gleadow and Duddy, 1980). In the Bass Basin, there is seismic evidence for widespread vulcanism at the level of the mid-Cretaceous breakup unconformity and in the lower parts of the overlying Eastern View Coal Measures (Williamson *et al*., 1987).

Cretaceous alkaline igneous rocks of variable composition occur in the southeast (Port Cygnet–Oyster Cove area) and in the far northeast (Cape Portland). The Port Cygnet rocks occur as a series of sheet-like bodies a few tens of metres thick and as numerous, widely-dispersed smaller sills, dykes and pipes (Farmer, 1981; 1985). They are principally intruded into rocks of the Lower Parmeener Supergroup, and are most abundant within the lower stratigraphic units of the sequence. Over-saturated syenite porphyry is the predominant rock type and forms the major sheet-like bodies. Also present are dykes of undersaturated sanidine porphyry, garnet trachyte, hornblende porphyry and other types (Farmer, 1985; Ford, 1989). Locally there are xenoliths of quartzite, marble, quartz, and amphibolite (Ford, 1989). Geochemistry of the garnet trachyte suggests an origin by assimilation of mineralised carbonate into a sanidine porphyry-type melt. $^{87}Sr/^{86}Sr$ and REE systematics suggest that the Port Cygnet rocks were formed by partial melting and mixing of an undersaturated alkali basaltic parent with amphibolite (Ford, 1983). Disseminated gold in the Cygnet district is associated with the alkaline intrusive rocks. A large magnetic anomaly is centred on Cygnet (Plate 2), and a combined regional magnetics/gravity interpretation suggests that a large syenite laccolith, 750 m thick, lies close to the surface at Cygnet (Leaman, 1990*b*). Four K-Ar ages of mineral separates give good agreement and indicate an age of emplacement at 100.5 ± 3 Ma (Evernden and Richards, 1962; McDougall and Leggo, 1965).

Near Cape Portland an appinitic suite comprising andesite, lamprophyre and porphyrite occurs as small flows, dykes and irregular plug-like intrusions (Jennings and Sutherland, 1969). These rocks were dated at 101.3–102.3 ± 2.6 Ma (McDougall and Green, 1982). At nearby Musselroe Bay, 'basalt' and lamprophyre intruding Lower Parmeener Supergroup rocks have been dated at 98.7 ± 0.8 Ma (Baillie, 1984).

Cretaceous thermal events

Organic matter in palynological preparations from the Parmeener Supergroup throughout the Cygnet–Oyster Cove–Woodbridge district is totally carbonised, and this is attributed to regional heating associated with the syenite intrusions of the Port Cygnet district (Farmer, 1985). However, 30 km southwest of Port Cygnet, at Ida Bay, conodont colour alteration in the Ordovician Gordon Group indicates maximum burial temperatures of less than 100°C (Burrett, *in* Sharples and Klootwijk, 1981). Notwithstanding these low burial temperatures, the Gordon Group was here remagnetised by a Late Cretaceous regional heating event that persisted for a period in the order of 10 million years (Sharples and Klootwijk, 1981). Apatite fission track analysis in the Durroon 1 well, in the Durroon Basin, suggests high geothermal gradients (55°C/km) in the early Late Cretaceous (Duddy, 1992).

Bass Basin

The intracratonic Bass Basin contains up to 12 km of ?Jurassic to recent sediments. A poorly known, early rift-fill succession of immature lithic sandstone is correlated with the Otway Group, and is considered to range from possibly Late Jurassic to Early Cretaceous age (Williamson *et al.*, 1987). A widespread middle Cretaceous unconformity correlates with the Australia–Antarctica breakup. This is succeeded by the Eastern View Coal Measures, a correlate of the hydrocarbon-producing Latrobe Group of the Gippsland Basin. The Eastern View Coal Measures comprise up to 8 km of sandstone, conglomerate shale and coal of alluvial fan, lacustrine and flood-plain origin. A late Paleocene–early Eocene unconformity (seismic sequence boundary) occurs near the top. The succession is sandier (more proximal) in the southern Bass Basin, than in the north. A further sequence boundary of late Eocene age defines the base of the overlying Demons Bluff Formation, a restricted marine shale, which then grades up into marl and limestone of the open-marine, Oligocene to Holocene, Torquay Group (Williamson *et al.*, 1987). Major fault movement in the Bass Basin largely ceased by the mid-Eocene, with minor reactivation up to the Holocene.

Durroon Basin

The Durroon Basin (Edgerley and Taylor, 1990; Baillie and Pickering, 1991) lies adjacent and southeast of the closely-related Bass Basin (fig. 3). The Durroon 1 well (Esso, 1973) intersected a sandier (more proximal) Cretaceous–Cainozoic succession than is present in the Bass Basin (Williamson *et al.*, 1987). Late Cretaceous rocks at the onshore southeast extremity of the Durroon Basin are mainly conglomerate (Boobyalla 1 and 2 wells; Moore *et al.*, 1984).

The Otway Group is over 1200 m thick in Durroon 1 (base not reached), and is a rift-fill succession of immature volcaniclastic lithic sandstone, siltstone, rare conglomerate and thin coal seams. Palynological evidence suggests that the Otway Group is here Aptian–Albian (Esso, 1973; Morgan, 1991). Apatite fission track analysis suggests high geothermal gradients (up to 55°C/km) were experienced at 95–100 Ma, followed by Cenomanian uplift and erosion of about 900 m of Otway Group, followed by a lower geothermal gradient (27°C/km) to the present day (Duddy, 1992). A thin band of undated, altered olivine basalt succeeds the unconformity (= Southern Ocean breakup unconformity), followed by 300 m of carbonaceous shale (lacustrine Durroon Mudstone) of Turonian–Coniacian age (Morgan, 1991). An angular unconformity (Tasman Sea breakup unconformity) is then followed by a relatively thin correlate (790 m) of the Eastern View Coal Measures, of non-marine, coarse-grained quartz sandstone and minor shale. This unit is Campanian to middle Eocene, with significant breaks at the top of the Cretaceous and the top of the Palaeocene. Then follows the middle to late Eocene Demons Bluff Formation (here consisting of sandstone, 40 m thick); and finally 540 m of the Oligocene–Holocene Torquay Group, of sandstone, shale, then shallow-marine calcarenite. Normal faulting, of predominantly northwest trend, occurred from Early Cretaceous to mid-Eocene. Graben development in the Durroon Basin was asymmetrical, with fault blocks tilted southwest.

Palynocorrelates of the Durroon Mudstone and lower Eastern View Coal Measures in the onshore proximal extremity of the basin, intersected in Boobyalla 1 and 2, consist of at least 500 m of conglomerate, mainly derived from Jurassic dolerite, passing laterally into a shale and siltstone-dominated succession a few kilometres away from the basin-margin fault (Moore *et al.*, 1984).

Tamar Graben

The Tamar Graben is an onshore southern extension of the Bass Basin. The oldest known sediments are 100 m of latest Cretaceous and Palaeocene mudstone, sandstone and conglomerate intersected in a well in the southern part of the graben. At Bell Bay, at the northern (seaward) end, this older succession is absent (probably because of erosion) and basement is overlain by a younger (latest Paleocene–middle Eocene) succession of sandstone, mudstone and minor lignite. Middle to late Eocene sequences are restricted to south of White Hills (basalt and conglomerate). At Bell Bay, an Eocene hiatus is overlain by 60 m of Oligocene carbonaceous silt and sand. There are also post-basalt, siliceous gravels of probable late Pliocene age (Forsyth, 1989*b*).

The Longford Sub-basin is a southern extension of the Tamar Graben, and contains up to 800 m of clay, sand and gravel, with basalt towards the top of the succession. The sediments are largely Eocene. Palaeocene sediments are present in the deepest part of the basin (Matthews, 1983).

Devonport–Port Sorell Sub-basin

The fill of the Devonport–Port Sorell Sub-basin (Cromer, 1989; Burns, 1964) begins with 250 m of carbonaceous mudstone, sandstone and minor conglomerate with Palaeocene and early Eocene microfloral assemblages (Harford beds); overlain by

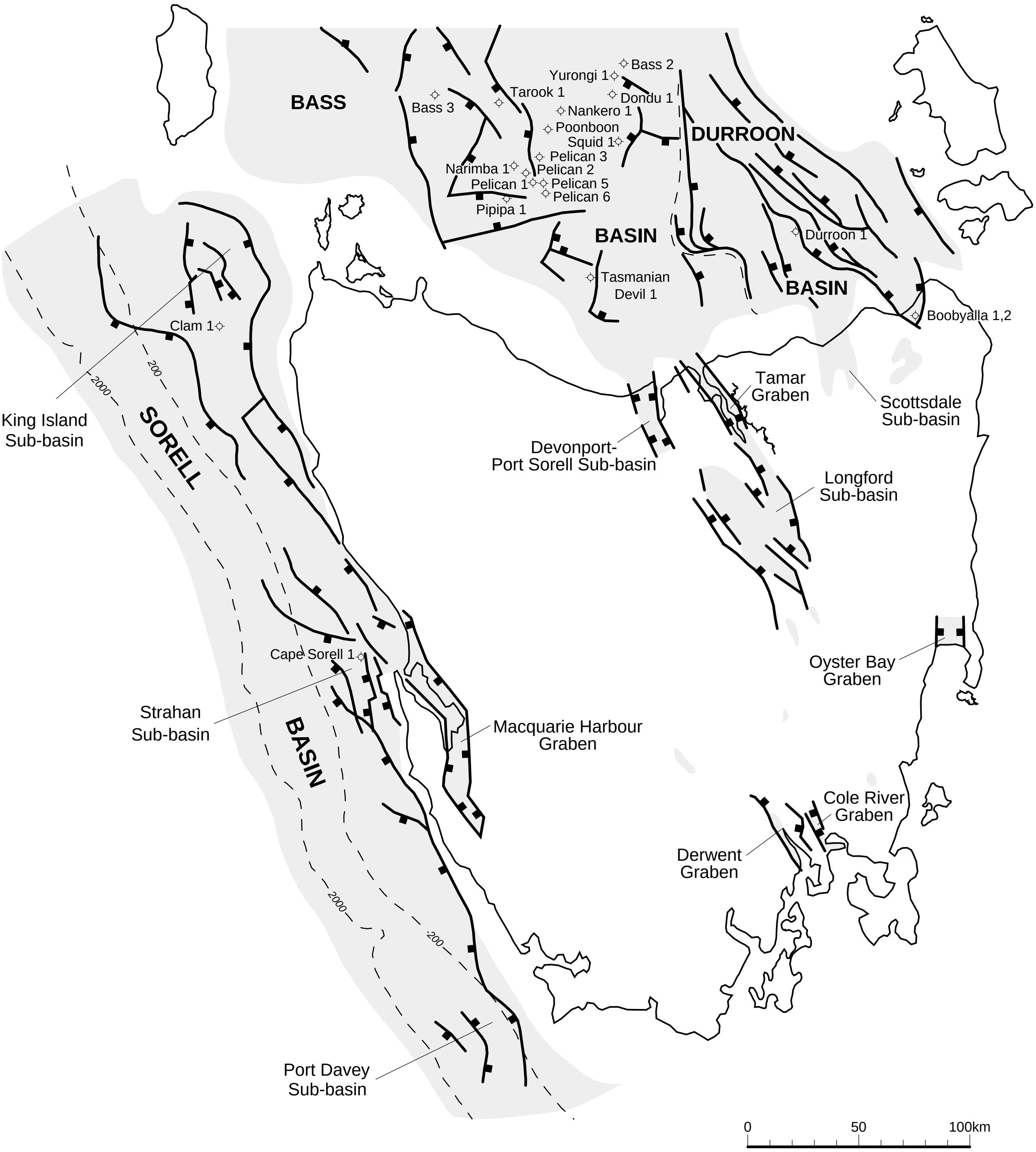

Figure 3.

Main areas of late Mesozoic–Cainozoic sediments in and around Tasmania, and major faults active in basin formation. From Baillie (1989b); Baillie and Pickering (1991); Forsyth (pers. comm.).

175 m of repetitious thin flows of alkali olivine basalt (Thirlstane Basalt, dated at 38 Ma — Cromer, 1980); 75 m of weakly consolidated sandstone of late Oligocene age (Wesley Vale Sand); and finally 50 m of basalt (Moriarty Basalt — 25.9 Ma; Baillie, 1986).

Sorell Basin

Sediments of the Sorell Basin cover the western continental margin of Tasmania (Willcox *et al*., 1989), and on the continental shelf there are local depocentres with over 4 km of sediment thickness; the King Island, Sandy Cape, Strahan and Port Davey Sub-basins. These sub-basins developed as trans-tensional features, with an element of left-lateral strike-slip movement, in the Early Cretaceous (Willcox *et al*., 1989).

The King Island Sub-basin is a half-graben that deepens eastward to a high-angle normal fault downthrown to the north and west (Willcox *et al*., 1989). In Clam 1, drilled on the shallower western flank of the half-graben, a Rocky Cape Group lithocorrelate (Proterozoic) is overlain unconformably by 190 m of conglomeratic red beds of problematic, probably Cretaceous, age. Then follows 347 m of quartz sandstone and mudstone with a mid-Upper Cretaceous microflora, correlated with the Sherbrook Group of the Otway Basin; conglomerate, sandstone and mudstone of Late Cretaceous to early Eocene age correlated with the Wangerrip Group (402 m); 126 m of probably-marine quartz sandstone; then 285 m of mudstone, marl and limestone of late Oligocene to Miocene age (Heytesbury Group correlate) (Lunt, 1969; Baillie and Hudspeth, 1989). In the King Island Sub-basin, gentle warping, uplift and erosion in the late Miocene resulted in a low-angle unconformity at the base of the Pliocene (Jones and Holdgate, 1980).

In Sorell 1, drilled just offshore of a large basin-margin fault in the Strahan Sub-basin, the section begins with 700 m of non-marine to marginal marine conglomerate, sandstone and shale of Maastrichtian to lower Palaeocene age, followed by a thick (2500 m) succession of interbedded sandstone and shale, of Palaeocene to late Oligocene age — this succession is marginal marine, deepening to sublittoral palaeoenvironment. The bulk of this succession is Palaeocene to early Eocene, with the uppermost 100 m being Oligocene. Faulting, in part related to sinistral strike-slip movement, terminates at the mid-Oligocene unconformity (Willcox *et al*., 1989). The uppermost unit is 250 m of limestone of late Oligocene to Miocene age (a correlate of the Heytesbury Group) (Hughes *et al*., 1983; Baillie and Hudspeth, 1989; Anon., 1989).

The Macquarie Harbour Graben is a shallow onshore extension of the Sorell Sub-basin, containing a sedimentary fill that is about 500 m thick immediately west of Strahan (Baillie and Corbett, 1985). Basal conglomerates, in places composed largely of Jurassic dolerite boulders, are overlain by Eocene sediments (Cox *in* Baillie and Corbett, 1985). The early Eocene succession includes marginal-marine interbedded sandstone, siltstone and minor coal overlain by fluvial cross-bedded sandstone (Baillie *et al*., 1986). Sandstone, siltstone and claystone around Strahan have Eocene microfloras and are disconformably overlain by conglomerate and sandstone of Plio-Pleistocene age (Baillie and Corbett, 1985).

Derwent Graben

The oldest sediments in the Derwent Graben — plant-bearing siltstone near Hobart — may be Palaeocene or early Eocene in age (Harris, 1968; Baillie and Leaman, 1989). Sediments of late Eocene to Oligocene age underlie basalt at Plenty and Sandy Bay (Forsyth *in* Baillie and Leaman, 1989; Gill, 1962); the basalts at Sandy Bay are dated at 26.5 ± 0.3 Ma (Sutherland and Wellman, 1986).

Onshore marine sediments

High sea level stands in the early early Miocene (early Longfordian) resulted in the deposition of a veneer of calcarenite and sandstone up to 30 m above present sea level along the coastal fringes of northwest Tasmania and King Island, and again in northwest Tasmania up to 100 m above present sea level in the late early to early middle Miocene (late Longfordian–Batesfordian) (Quilty, 1972). Late Oligocene limestone in the northeast subcrops 30 m below present sea level in Musselroe Bay; other preserved marine Tertiary rocks are just above sea level and are early Miocene in age (Quilty, 1972).

Marine carbonates of late Pliocene age are present a few metres above sea level on Flinders Island (Quilty, 1985).

Other onshore Tertiary sediments

There are widespread, patchy developments of terrestrial sediments, in many places preserved beneath basalt flows, that are almost all latest Eocene or younger (Forsyth, 1989*c*). These are, for the most part, not shown on Plate 1.

Fluvial and lacustrine gravel, sand and clay fill shallow depressions in northeast Tasmania, and are locally important sources of placer cassiterite. Three broad age groupings of tin-bearing deposits are recognised (Morrison, 1989):–

- probably-Eocene deposits filling narrow channels north and west of Mt Cameron;
- late Oligocene braidplain deposits in the Ringarooma Valley, that have sourced most of the recovered tin; and
- deposits in the terraces of the present Ringarooma and Musselroe Rivers, that post-date the extrusion of widespread mid-Miocene basalts.

Tertiary sediments are up to 225 m thick in the Scottsdale Sub-basin, and are late Oligocene to early Miocene age (Hill and McPhail, 1983; Harris, 1968; Brown *in* McClenaghan *et al*., 1982; Moore, 1989).

Chromite-rich Tertiary sand derived from the Andersons Creek Ultramafic Complex has been mined near Beaconsfield (Summons *et al*., 1981).

Cainozoic volcanic rocks

Cainozoic basalt is widely distributed across northern and eastern Tasmania and in the Bass Basin, but is absent from southwest Tasmania. Some 36 radiometric dates are available, with ages ranging from Palaeocene to late Miocene (58–8 Ma); all are whole-rock K-Ar dates unless otherwise stated. Silica-undersaturated to saturated types are temporally widespread, but oversaturated basalts are confined to the mid-Tertiary (21–31 Ma), suggesting that relatively high degrees of partial melting of mantle source rocks corresponding to the peak of volcanism were Oligocene to early Miocene (Sutherland and Wellman, 1986). The oldest basalts lie to the east and the youngest basalts (16–8 Ma) young to the west along the north coast.

The oldest known rock is an alkali olivine basalt at Bream Creek (far south of Northeast Tasmania Element) dated at 58.5 ± 0.7 Ma (Palaeocene) (Baillie, 1987). In the north of the Northeast Tasmania Element, the erosional remnants of alkaline basalt cap hills above 500 masl in the Weldborough area; three dates cluster closely about 47 Ma (Sutherland and Wellman, 1986). Locally, 150 m of agglomerate and tuff underlie 230 m of flows (McClenaghan *in* McClenaghan *et al*., 1982). Partly dissected flows in the Ringarooma Valley give late early Miocene ages of 16.0 to 16.4 Ma (three dates: Brown *in* McClenaghan *et al*., 1982; Sutherland and Wellman, 1986). Locally these rocks overlie the early Miocene terrestrial gravels.

In the northern Northeast Tasmania Element, a variable suite of flows — ranging in age from probably middle Eocene to at least early Miocene — filled valleys draining northward into Bass Strait (Sutherland and Wellman, 1986). Alkali olivine basalt is overlain by quartz tholeiite, alkaline basalt and olivine nephelinite. A tholeiite has been dated at 30.7 ± 0.4 Ma; the flows are thought to range in age from probably middle Eocene to at least early Miocene (Sutherland and Wellman, 1986).

In the central-western part of the Northeast Element, an olivine tholeiite has been dated at 25.6 ± 0.2 Ma but widespread, undated tholeiitic lavas are thought to be younger (early to middle Miocene) by Sutherland and Wellman (1986) but may be older (middle to late Eocene) (Everard *in* Matthews and Everard, 1995).

In the northwest (Rocky Cape Element) the lavas overtopped the valleys to form an extensive lava plain, which has an upper surface at 100 m on the coast at Wynyard rising to 750 m altitude 70 km inland. The succession includes late Eocene and early Oligocene lake sediments in the Waratah area (northern Dundas Trough) (Brown and Forsyth, 1984). The basalts in this area include late Oligocene or older alkaline rocks (26.3 ± 0.3 Ma K-Ar minimum age), early Miocene olivine tholeiite, and middle to upper Miocene basanites, the latter with dates of 13.3 ± 0.2 Ma minimum age at Table Cape (Sutherland and Wellman, 1986) and 12.5 ± 0.2 Ma and 8.5 ± 0.1 Ma at Stanley (Brown, 1989*a*). At Mt Cameron West, olivine basalt lava dated at 14.5 ± 0.2 Ma and 15.5 ± 0.2 Ma overlies early early Miocene limestone (Seymour and Baillie, 1992).

Two dates from plugs of olivine melilite nephelinite in the eastern and southern Central Plateau area (approximately projected northern boundary of Tyennan and Jubilee Elements) are 35.4 ± 0.4 Ma and 30.1 ± 0.4 Ma (Sutherland and Wellman, 1986). Lavas ranging from olivine nephelinite to quartz tholeiite were erupted onto the Central Plateau (which includes areas of the Tasmania Basin, Jurassic dolerite, and northern Tyennan Element), and descended south into the Derwent drainage system, comprising sequences locally up to 400 m in thickness. Five dates are late Oligocene–early Miocene (22.4–24.2 ± 0.6 Ma) (Sutherland *et al*., 1973; Sutherland and Wellman, 1986). An olivine nephelinite plug and a nepheline hawaiite flow at the eastern edge of the Central Plateau have yielded dates of 24.9 ± 0.2 Ma and 24.2 ± 0.2 Ma respectively (Sutherland, 1989). Nearby, in the western Midlands, an alkaline flow and plug are dated at 36.3 ± 0.5 Ma and 27.6 ± 0.4 Ma; and two hawaiite flows are 25.0 ± 0.3 and 24.3 ± 0.3 Ma (Sutherland and Wellman, 1986).

In the southeast (Tasmania Basin and Adamsfield–Jubilee Element) an alkali basalt (near Campania) is dated at 24.2 ± 1 Ma (minimum age); this is overlain by widespread olivine tholeiites that are thought to be latest Oligocene to early Miocene (Sutherland and Wellman, 1986). An olivine basalt near Hobart is dated at 23.0 ± 0.5 Ma (Tedford *et al*., 1975). Nearby (Sandy Bay) sediments of probably Yallournian (Oligocene) age are overlain by alkaline volcanic rocks dated at 26.5 ± 0.3 Ma (Sutherland and Wellman, 1986).

Quaternary

Surficial Quaternary sediments, including glacigenic, slope, coastal, aeolian, fluvial and cave deposits, are widespread in Tasmania (see Colhoun, 1989). For reasons of scale no attempt has been made to show these on Plate 1.

Quaternary placer deposits include alluvial cassiterite and gold in northeast Tasmania, and osmiridium and chromite in western and southern Tasmania (Threader *in* Colhoun, 1989).

Directions for Further Work

A significant part of the rationale for making this compilation prior to the main interpretative phase of the TASGO Project was to highlight problems which need further work, particularly those which may be addressed within components of the TASGO Project itself. The new data from the TASGO deep seismic sounding, geochronology, and aeromagnetic coverage, is expected to generate new or revised interpretations. Some of the significant problems remaining to be addressed are:–

- A number of Early Palaeozoic rock units have been interpreted as allochthonous, or are at least suspected to be so. The allochthon model seems attractive for the ultramafic/mafic complexes, due to comparison with the tectonics of similar rocks in Oman, and to geochemical links between the ultramafic rocks and some of the mafic volcanic associations. Furthermore, the emplacement of an ultramafic/mafic sheet across all of Tasmania would necessitate tectonics involving large lateral translations. However, can it be assumed that the ?allochthonous sedimentary sequences (Cleveland–Waratah association, Ragged Basin Complex, etc.) were derived from an original location in the same primary tectonic environment as the ultramafic/mafic complexes, or did they come from another environment at some intermediate distance? In other words, what is their degree of allochthoneity? Are some of (or parts of) the older sequences/blocks (e.g. the ?Mesoproterozoic of the Tyennan Element) also allochthonous?

- Corroboration is needed of the ages and correlation of the Neoproterozoic units, and of the various ?allochthonous sequences. Techniques which may assist include isotope chemostratigraphy of carbonate rocks, zircon geochronology of volcanic rocks, and trace element lithogeochemistry.

- Correlation of Mesoproterozoic units between the Rocky Cape, Tyennan and Adamsfield–Jubilee Elements is presently conjectural. Dating of detrital zircons will give a provenance signature and maximum age for each sequence.

- It now seems likely that if any fundamental tectonic suture exists between the Rocky Cape and Dundas Elements, such a boundary lies within the Arthur Lineament. What is the nature and origin of any such suture?

- The depositional age of the Bowry Formation of the Arthur Metamorphic Complex is still questionable. At the time of finalisation of this report, new zircon geochronology of a highly deformed granite intruding mafic schist of the Bowry Formation has indicated a magmatic crystallisation age of 777 ± 7 Ma (Turner and Black, in prep.). Is the Bowry Formation a sliver of old basement caught up in the Arthur Metamorphic Complex?

- What is the stratotectonic significance of the old radiometric ages obtained from inherited zircon grains in a number of disparate host rock units in western Tasmania?

- What is the nature of the contact between the Oonah Formation and the Success Creek Group north and south of the Pieman River in the Dundas Element? If a major unconformity is present, what is its tectonic significance?

- Further work is needed on the possible existence of a structural and metamorphic hiatus at the base of the Jane Dolomite. If such a break exists it represents evidence for an intra-Proterozoic tectono-metamorphic event — whereas most effects of the event once known as the Frenchman Orogeny in the Tyennan Element have been re-interpreted as ca. 500 Ma in age.

- What is the full extent of the Adamsfield–Jubilee and Tyennan Elements beneath the younger cover rocks of the Tasmania Basin?

- What is the nature of the western boundary of the Northeast Tasmania Element? Does the concept of a 'Tamar Fracture' still have any relevance?

- Can zones of mineralisation be related to major structures at middle crustal levels?

- The chronology of late Mesozoic–Cainozoic uplift and erosion needs elucidation. More fission-track work is needed, particularly in southern Tasmania.

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