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Controls on Devonian–Carboniferous magmatism in Tasmania, based on inherited zircon age patterns, Sr, Nd and Pb isotopes, and major and trace element geochemistry

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Devonian–Carboniferous granites are widespread in Tasmania. In eastern Tasmania, Devonian granites intrude Ordovician–Early Devonian quartz-rich turbidites of the Mathinna Supergroup. The earliest (~400 Ma) I-type granodiorites may be arc-related. Following the Tabberabberan Orogeny (~389 Ma), more felsic and, finally, strongly fractionated I- and S-type granites were emplaced until ~373 Ma. In contrast, western Tasmania granites intrude a more diverse terrane of predominantly marine shelf successions, with depositional ages as old as Late Mesoproterozoic. They are mostly felsic and fractionated I- and S-types emplaced from ~374–351 Ma, possibly in response to post-collisional crustal extension following juxtaposition of the eastern and western Tasmanian terranes. Granites from the two terranes are readily distinguishable by the age spectra of their inherited zircon, which are noticeably similar to those of the detrital zircon from sedimentary successions in their respective terranes. Furthermore, within each terrane, both I and S-types yield similar inheritance patterns. This suggests a pivotal role for the sedimentary successions in the petrogenesis of both types. Western Tasmanian granites are also enriched in ~1600 Ma zircon, which is essentially unrepresented in the exposed supracrustal succession. Subtle differences between the inheritance and detrital age spectra in eastern Tasmania probably relate to unrepresentative sampling of the supracrustal rocks. Nd, Sr and Pb isotopic characteristics of the granites are consistent with their derivation by mixing of magmas derived from the mantle, possibly the lower crust, and from supracrustal rocks. Systematic isotopic trends in some eastern Tasmanian I-types, particularly in the Scottsdale Batholith, correlate well with major and trace element geochemistry and age. The isotopes are inconsistent with simple restite unmixing or crystal fractionation in a closed magma chamber, and indicate progressive contamination by the Mathinna Supergroup, or similar rocks. The isotopic characteristics of late, strongly fractionated granites, although sometimes obscured by hydrothermal alteration, are also consistent with concurrent assimilation-fractional crystallisation processes. Together with the close association of some strongly fractionated I- and S-types, this suggests that such granites were generated directly in the lower crust, and were not derived from unfractionated parental granite magmas.

KEY WORDS: Carboniferous, Devonian, granite genesis, inherited zircon, Nd isotopes, Pb isotopes, SHRIMP, Sr isotopes, Tasmania, U–Pb ages.

INTRODUCTION

The petrogenesis of eastern Australian granites, particularly in the Lachlan Orogen, has been keenly debated for over three decades. Perhaps the crux of the controversy is the relative roles of mantle, lower crust and upper crust, and the extent to which the granites represent addition of juvenile material to the crust.

One school of thought, first outlined by Chappell & White (1974), proposes that the granites were derived by the partial melting of concealed Proterozoic basement.

Mantle-derived magmas may have acted as a heat source, but contributed little or no material to the granites, which simply redistributed material more-or-less vertically, within the crust (e.g. Chappell 1997). The granites are interpreted as ‘imaging their source rocks’ (Chappell 1979), in that a first order classification can be made into I-types and S-types, derived from meta-igneous (or infracrustal) and metasedimentary (or supracrustal) protoliths, respectively. Plutons can be grouped into suites, each with certain common characteristics and thought to be derived from a common

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protolith. Variations within each suite are primarily attributed to differential separation of melt and entrained crystals (or restite) within a liquid-crystal mush, producing linear trends on appropriate variation diagrams.

An alternative model, largely suggested by isotopic data, derives the granites by two-component mixing between a basaltic, mantle-derived component and a partial melt of the exposed sedimentary rocks (e.g. McCulloch & Chappell 1982; Gray 1984). McCulloch & Chappell (1982), however, argued that this cannot satisfy major and trace element constraints. Collins (1996, 1998) developed a more complex mixing model with an additional component, partial melts derived from a Neoproterozoic–Cambrian lower crust (represented at the surface by the mafic volcanic belts of the Lachlan Orogen). He claimed that this three-component model avoids the perceived problems of the two component model, but this has been contested by proponents of the restite model (Chappell *et al.* 1999, 2000).

Studies of the Devonian–Carboniferous granites of Tasmania have the potential to resolve, or at least illuminate, this controversy, because in the east and west of the island, they intrude contrasting terranes with different histories (Figure 1).

GEOLOGICAL BACKGROUND

The Early–Late Devonian granites of eastern Tasmania intrude an Early Ordovician to Early Devonian quartz-rich turbidite succession, the Mathinna Supergroup, which is the southward extension of similar successions in the Lachlan Orogen of southeastern mainland Australia.

In contrast, the slightly younger (Late Devonian to Early Carboniferous) granites of western Tasmania intrude a separate, older and more diverse terrane, termed the Taswegia terrane by Stump *et al.* (1986). Probably the oldest unit is the Surprise Bay Formation in western King Island, consisting of amphibolite-grade metasedimentary rocks and minor amphibolites; its youngest detrital zircons are $\sim 1350 \pm 90$ Ma (Black *et al.* 2004). It has yielded a 1297 ± 18 Ma U–Th–Pb monazite age, interpreted as recording an early Grenvillean metamorphic event (Berry *et al.* 2005) and was deformed and intruded by syntectonic granitoids at 760–748 Ma (Wickham Orogeny, Black *et al.* 1997; Turner *et al.* 1998). It is uncertain whether, or to what degree, these events are represented in mainland Tasmania.

The other western Tasmanian successions are largely composed of mature, shallow-water marine shelf deposits (Rocky Cape Group and correlates in the Tyennan and Jubilee Regions). These have similar detrital zircon age spectra to the Surprise Bay Formation. Their youngest detrital zircon ages, also mostly ~ 1400 Ma, define a maximum possible depositional age, although the Jacob Quartzite, near the top of the Rocky Cape Group, contains a population at ~ 1010 Ma (Black *et al.* 2004). Proterozoic turbidites are also present in western Tasmania, and have detrital zircon patterns similar to the Rocky Cape Group, with maximum

depositional ages ranging from ~ 1200 Ma (Oonah Formation) to ~ 1070 Ma (Badger Head Group) (Black *et al.* 2004). It is uncertain whether the turbidites are contemporaneous with the shelf successions or derived from them.

The minimum age of all these successions is defined by the Middle Cryogenian (~ 750 Ma) age of the base of overlying Togari Group and its correlates (Calver 1998). These successions consist predominantly of shallow water siliciclastics, dolostone, mafic volcanics and volcanoclastics, accompanied by rare felsic lavas and shallow intrusions dated at ~ 580 Ma (Calver *et al.* 2004).

An arc–continent collision in the Early Cambrian (Tyennan Orogeny; Berry & Crawford 1988) was followed by widespread felsic volcanism (Mount Read Volcanics) and subvolcanic intrusions, which are dated at about 505 Ma to 493 Ma (Perkins & Walshe 1993; Black *et al.* 1997, 2005; Turner *et al.* 1998). Subsequently, Late Cambrian to Early Devonian shelf successions (Wurawina Supergroup) were widely deposited throughout western Tasmania.

Both eastern and western Tasmania were deformed during the Tabberabberan Orogeny at ~ 389 Ma (Black *et al.* 2005), which probably represents the time of juxtaposition of the two terranes. The terrane boundary, possibly a series of thrusts, is obscured by the Carboniferous–Triassic Parmeener Supergroup and Jurassic dolerite, but small inliers and deep drill holes indicate that it extends approximately between Port Sorell in the north to the Tasman Peninsula in the south (Figure 1).

The Devonian–Carboniferous granites have narrow contact aureoles and were emplaced at high crustal levels, commonly in distinct batholiths (Figures 1–3). They have been grouped into about twenty suites by McClenaghan (2006), to which the reader is referred for a comprehensive description. Compared with the Lachlan Orogen, there is a high proportion of evolved, strongly fractionated bodies in Tasmania (particularly in the west), but mafic S-types, A-types and ‘high temperature’ I-types appear to be completely absent. Their crystallisation ages recently have been determined by SHRIMP U–Pb zircon dating (Black *et al.* 2005).

In eastern Tasmania, the oldest granites (~ 400 Ma) are typically foliated I-type hornblende-biotite granodiorites containing abundant mafic enclaves. At about 385 Ma, shortly after the Tabberabberan Orogeny, abundant felsic S-type granites began to accompany the I-types. Strongly fractionated, alkali-feldspar granites, also including both I- and S-types, were the last to be emplaced in eastern Tasmania, between about 378–373 Ma.

In western Tasmania, the granites are slightly younger, all postdating the Tabberabberan Orogeny, and occur in isolated intrusions or small two-pluton batholiths (Figures 1, 3). Felsic ($> 70\%$ SiO₂), and mostly moderately to strongly fractionated, bodies predominate in mainland Tasmania. The oldest are the I-type Housetop and Wombat Flat Granites (374–373 Ma) in the northwest, but S-types had appeared by 370 Ma in the southwest and by 362 Ma in the northwest. The youngest granites (~ 351 Ma) are relatively mafic unfractionated I-types on King Island.

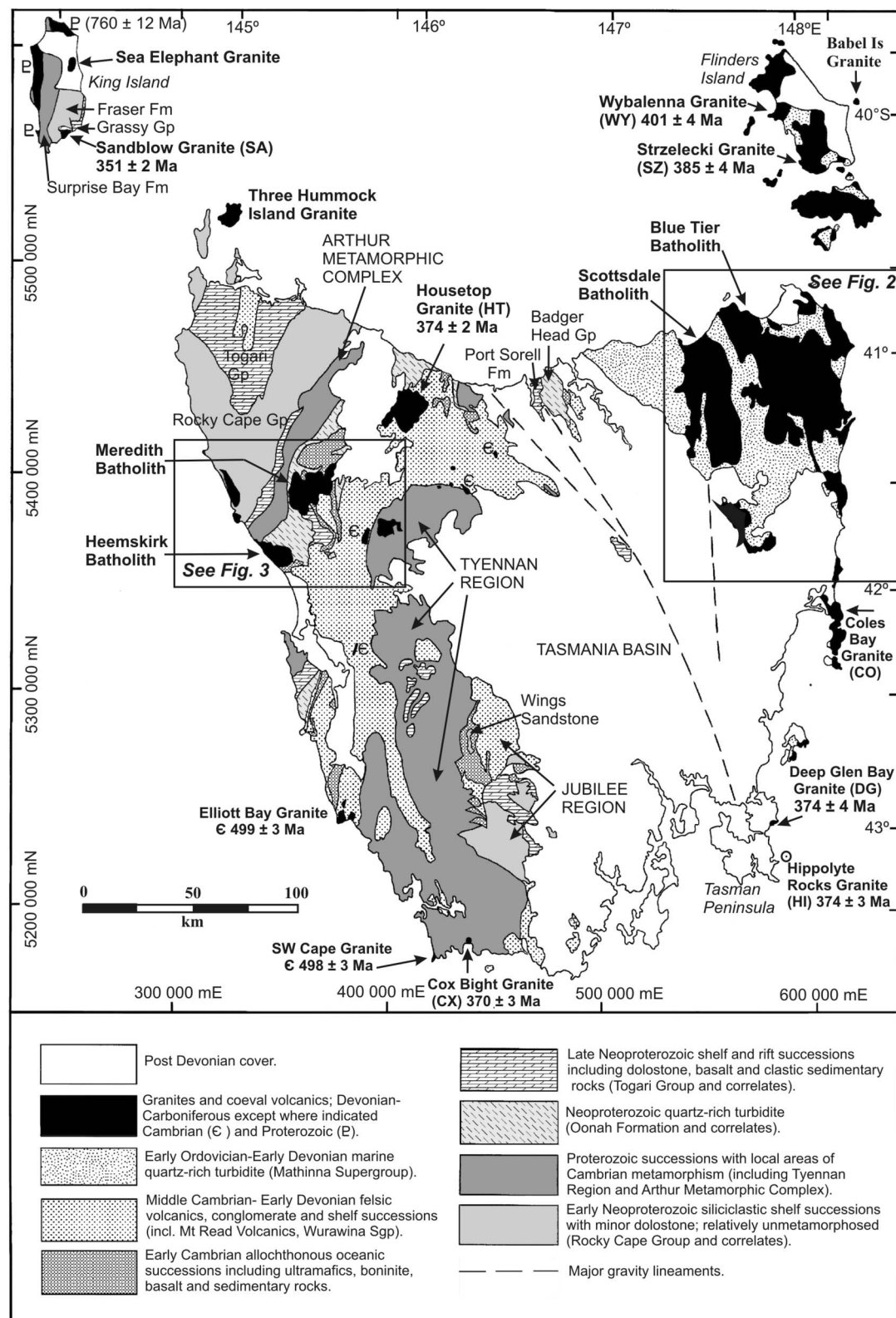


Figure 1 Simplified geology of Tasmania; showing the gross distribution of Devonian and older rock types, and Devonian–Carboniferous, Cambrian and Proterozoic granites (crystallisation ages of the granites from Black *et al.* 2005).

SAMPLE SELECTION

We have determined the age of inherited zircon grains from, and the Sr, Nd and Pb isotopic compositions of, the

23 samples of Tasmanian granites and a mafic enclave which were dated using the SHRIMP U–Pb method by Black *et al.* (2005) (Figures 1–3). The major and trace

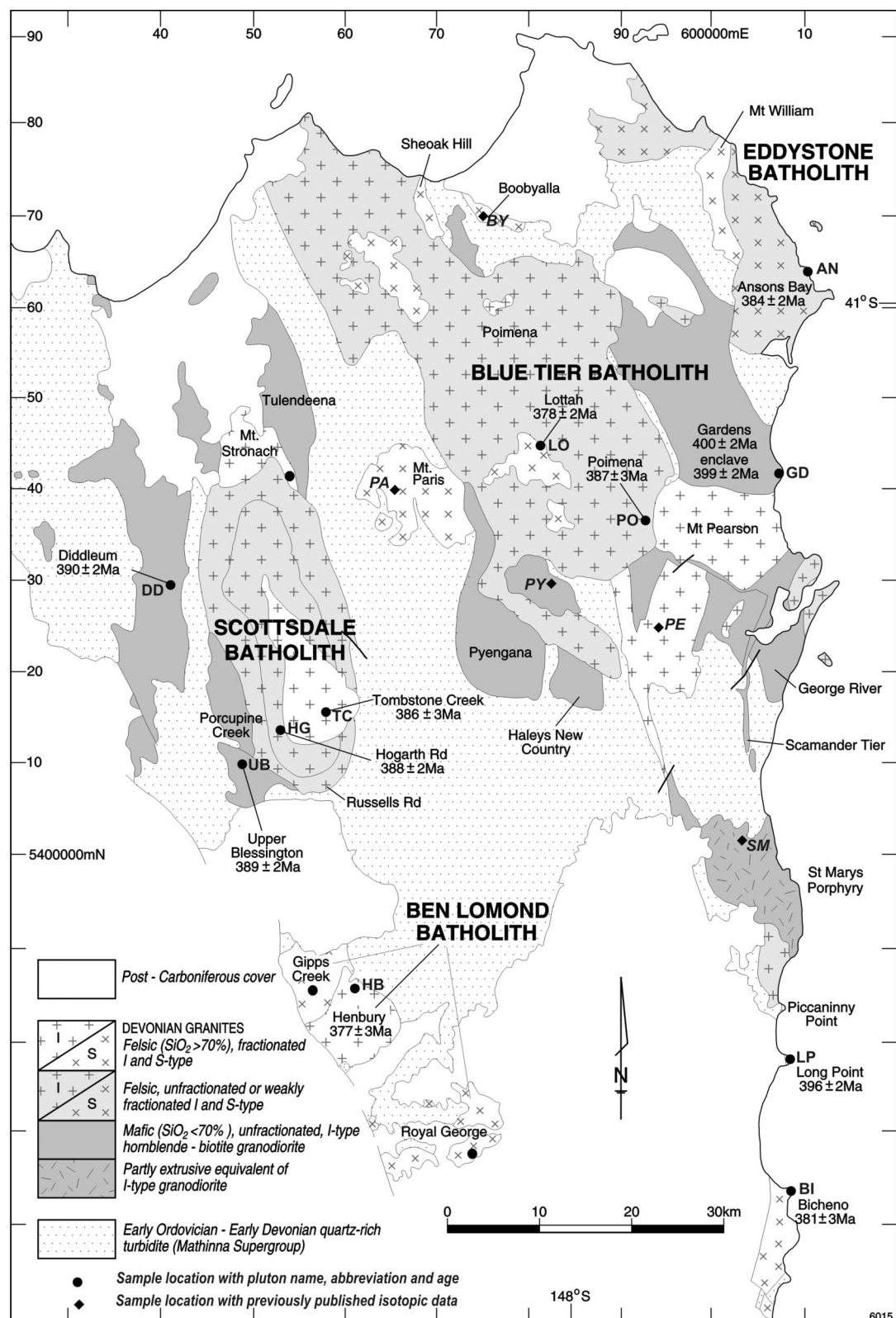


Figure 2 Simplified geology of northeast Tasmania showing Devonian granite batholiths and plutons, and sample locations (crystallisation ages from Black *et al.* 2005).

element geochemistry of the samples (Table 1) is illustrated by plots of Sr against Rb and Alumina Saturation Index (ASI, molar $\text{Al}_2\text{O}_3/(\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO})$) against SiO_2 (Figure 4). To test the nature and

degree of any role of the sedimentary country rocks in the petrogenesis of the granites, we also report isotopic data from nine of the seventeen Proterozoic and early Paleozoic sandstones for which detrital zircon ages were

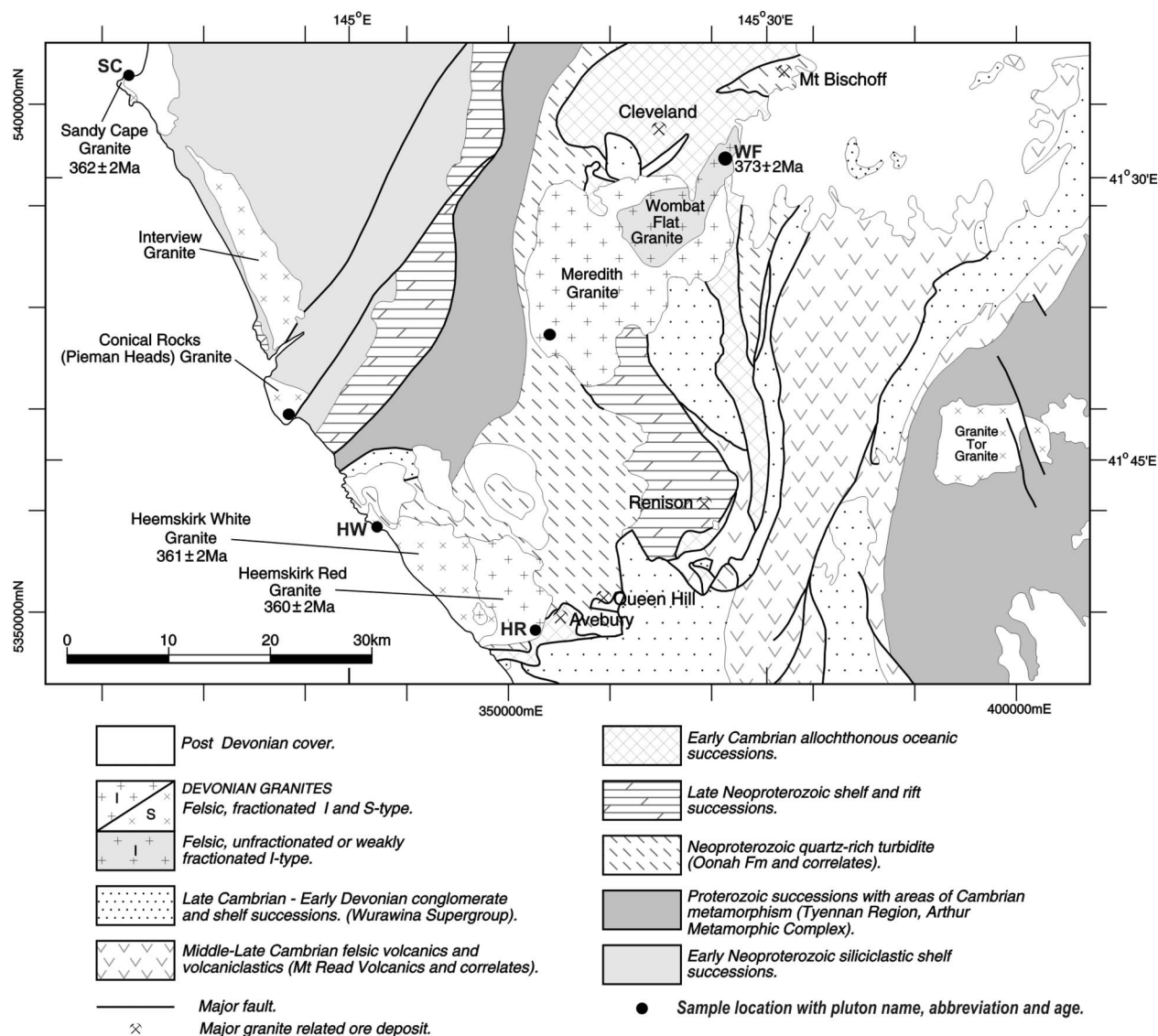


Figure 3 Simplified geology of part of western Tasmania showing Devonian granite plutons and sample locations (crystallisation ages from Black *et al.* 2005).

obtained by Black *et al.* (2004). The remainder are mainly orthoquartzites (>97.5% SiO₂), which we consider were too infertile to be likely source components for granites, and unsuitable for isotopic analysis due to the high and uncertain parent/daughter ratios, and the consequent difficulty of making accurate age corrections.

INHERITED ZIRCON AGE PROFILES OF THE TASMANIAN GRANITES

Analytical considerations and statistical methodology

Analytical procedures are described in Appendix 1. Inherited age spectra have been derived dominantly from zircon cores within prismatically zoned zircon that crystallised from the host magma. Those cores are thought to represent zircon that was caught up

early in the history of the magma and, therefore, are most likely to reveal information on its source region. In contrast, about 5% of the ages have been derived from grains without core-rim relationships, and which commonly have pock-marked surfaces. Although those grains are interpreted as late-stage xenocrysts, they yield very similar age spectra to the zircon cores (albeit from a much smaller sampling base).

At least one day was spent on the search for inherited zircon in each sample. Although inheritance was found in all 23 of the studied Devonian–Carboniferous granites, the yields were mostly poor or very poor. In addition, many of the cores were either too small to accommodate the ~25 µm SHRIMP analytical spot or, more commonly, contained too many imperfections (such as cracks, inclusions or metamict areas) to allow siting it solely on pure zircon. In many instances apparent cores, visible on optical and cathodoluminescence images, yielded the same age as their host granite,

Table 1 Chemical analyses of granite and host rock samples.

Eastern Tasmanian granites																	
Unit	GD	EG	LP	PO	LO	DD	UB	HG	TC	AN	HB	BC	HI	DG	SZ	SZ (f)	WY
Type	IMu		IMu	IFu	SFh	IMu	IMu	IFu	IFu	SFu	IFh	SFf	SFf	SFh	SFu	SFf	IMu
SiO ₂	67.77	57.71	66.66	72.48	76.22	63.97	62.64	70.26	74.01	75.48	76.12	76.28	77.46	74.39	71.09	76.12	67.95
TiO ₂	0.47	0.74	0.57	0.29	0.08	0.73	1.01	0.48	0.21	0.29	0.08	0.14	0.19	0.06	0.44	0.15	0.47
Al ₂ O ₃	14.71	16.13	13.76	14.07	12.78	15.55	15.67	14.23	13.63	12.61	12.39	12.43	12.02	12.43	14.81	12.63	14.17
Fe ₂ O ₃	0.55	1.16	0.81	0.56	0.27	1.40	1.02	0.72	0.20	0.33	0.66	0.40	0.64	0.40	0.38	0.15	3.01
FeO	2.93	5.86	3.46	1.42	0.96	3.72	4.92	2.66	1.40	1.86	0.46	1.14	0.52	0.82	2.64	1.30	1.22
MnO	0.07	0.16	0.08	0.05	0.03	0.09	0.11	0.06	0.05	0.03	0.02	0.02	0.02	0.03	0.06	0.01	0.09
MgO	2.11	4.59	3.31	0.60	0.07	2.97	2.43	1.20	0.43	0.33	0.05	0.19	0.17	0.07	0.93	0.13	2.23
CaO	3.97	6.53	4.10	1.82	0.45	4.97	4.76	2.39	1.41	1.41	0.61	0.64	0.51	0.43	2.82	0.81	3.63
Na ₂ O	2.56	3.22	2.26	3.03	2.94	3.43	3.61	3.58	3.55	2.21	3.29	2.56	2.79	3.05	3.35	2.45	2.48
K ₂ O	3.64	2.21	3.44	4.83	5.08	2.25	1.96	3.46	4.45	4.99	4.81	4.99	5.05	5.04	2.35	5.32	3.71
P ₂ O ₅	0.11	0.18	0.11	0.12	0.08	0.19	0.30	0.13	0.06	0.17	0.02	0.18	0.09	0.05	0.15	0.06	0.15
SO ₃ tot	0.04	0.09	0.07			0.07	0.09	0.05	0.07		0.06	0.08	0.07	0.07	0.06	0.00	0.10
rest	0.17	0.20	0.21	0.14	0.01	0.21	0.21	0.19	0.12	0.12	0.13	0.12	0.10	0.09	0.14	0.10	0.21
O = S, F	-0.02	-0.04	-0.03			-0.03	-0.03	-0.03	-0.04		-0.27	-0.11	-0.11	-0.16	-0.10	0.00	-0.04
LOI	0.63	0.64	0.84	0.57	0.95	0.08	0.78	0.34	0.47	0.14	1.38	0.89	0.46	3.14	0.82	0.60	0.68
Total	99.70	99.37	99.64	99.98	99.92	99.61	99.48	99.71	100.02	99.97	99.78	99.96	99.97	99.90	99.95	99.83	100.05
ASI	0.957	0.825	0.924	1.041	1.146	0.909	0.937	1.017	1.031	1.087	1.056	1.153	1.095	1.105	1.123	1.123	0.965
Li	35.3	45.2		56	129					50.5							
Be	2.2	2.8	2.4	3.7	7.9	2.3	2	3.5	4.9	3.4	7.8	5.9	5.2	9.1	5		2.6
F	481	853	650	821	3123	634	753	732	681	1757	6312	2302	2409	3552	2124		642
S				70	50					60			268				
Sc	13	30	18	7	5	16	17	10	5	5	7	3	5	5	8	<9	16
V	84	173	99	24	2	114	108	47	15	11	<3	6	8	<3	45	<5	90
Cr	56	147	167	8	<1	66	41	50	4	3	<2	10	<2	7	3	11	200
Ni	16	37	37.9	7	7	29.4	22	51.2	11	9	10	11	13	10	13	<5	29
Cu	10	21	18	3	3	25	25	12	7	6	5	10	5	6	13	6	12
Zn	49.6	94.2	50.9	45	45	63.1	95.2	53.7	33.7	52	31.6	56	38	21.4	65.5	38	63.2
Ga	15.5	18.7	16.1	18.2	26.1	18.2	18.7	17.5	16	20.4	267	19.5	19.4	22.7	18.4	19	15.9
Ge	1.6	2.2	1.8	1.8	2.3	1.6	1.4	1.7	2	1.6	2.5	1.8	1.9	2.3	1.6		1.9
As	2.2	1.1	2.1	0.4	<0.2	1.9	1.6	1.2	<0.5	1.2	<0.5	11.2	1.6	<0.5	1.1	<20	6.2
Rb	155.3	128.6	151.5	259.9	477.4	90.7	72.6	146.2	253.8	238.9	720	379.8	383.2	482	185.1	320	174
Sr	251.3	238.1	235.7	131	14.9	328.8	286.9	192.4	95.7	94.8	13.8	51	51.5	14.7	227.7	51	255.5
Y	16	35.2	26.1	32.5	50.9	23.2	22.2	25.4	34	52.3	201	30	69.3	74.4	21.1	26	32.4
Zr	104	118	144	141	83	173	313	177	125	212	121	120	120	74	193	90	145
Nb	7.6	11.2	9.6	11.6	13.9	12	16.1	14.4	15.9	12.9	32.5	13.1	11.2	12.4	14.1	9	10.7
Mo	1.5	4.3	2.4	0.2	0.8	3.4	1.7	8.1	1.6	0.4	3.7	6.6	0.9	3.5	1.7	<5	5.8
Ag	0.05	0.05	0.05	<0.5	<0.5	0.05	0.08	0.06	0.03	<0.5	0.04	0.04	0.04	0.02	0.03		0.02
Cd	0.05	0.18	<0.05			<0.05	0.13	0.13	0.15		<0.1	0.12	<0.1	0.1	<0.1	<9	<0.1
Sn	3.5	4.3	3.1	8	13	3.1	1.8	3.7	7.6	5	2.8	15.9	11.2	14.9	10.8		7.1
Sb	0.1	0.2	0.2	<0.2	<0.2	0.2	0.1	0.1	<0.1	<0.2	<0.1	<0.1	<0.1	0.1	0.1		<0.1

(continued)

(continued)

Table 1 (Continued).

Unit	Eastern Tasmanian granites																		
	GD	EG	LP	PO	LO	DD	UB	HG	TC	AN	HB	BC	HI	DG	SZ	SZ (f)	WY		
Cs	9.16	9.04	7.04	15.5	29.5	5.11	4.81	7.5	18.19	8.9	34.34	16.53	8.43	9.86	15.49		11.23		
Ba	475	359	493	413	59	546	442	545	346	546	31	234	82	25	320	220	579		
La	9.9	17.9	26.1	29.4	14.8	34.5	31.5	33.4	28.7	44.0	58.4	23.0	28.6	16.0	38.6	<20	37.0		
Ce	23.5	47.5	57.2	62.3	37.7	68.6	68.8	69.3	59.2	97.0	140.4	51.7	69.1	41.3	72.0	33	74.7		
Pr	2.98	6.5	6.4	7.3	4.7	6.96	7.6	7.3	6.35	11.9	15.96	5.55	7.38	4.76	7.07		7.92		
Nd	12.4	27.4	25.0	26.6	16.9	25.6	29.1	26.9	22.8	45.0	57.3	19.8	30.8	16.9	24.9	<20	29.48		
Sm	2.92	6.29	5.42	5.5	5.2	4.79	5.58	5.19	5.03	10.2	16.05	5.13	7.72	5.49	4.65		5.99		
Eu	0.85	0.96	0.99	0.80	0.02	1.15	1.59	0.90	0.52	1.20	0.12	0.49	0.23	0.09	0.98		1.02		
Gd	2.94	6.22	5.2	5	5.8	4.61	5.24	4.86	4.72	10.7	18.81	5.34	7.62	6.61	4.32		5.37		
Tb	0.5	1.06	0.81	0.9	1.4	0.71	0.81	0.79	0.85	1.9	4.53	1.11	1.78	1.7	0.76		0.98		
Dy	2.73	5.68	4.46	4.7	7.7	3.75	4.12	4.1	4.69	9.2	26.64	5.53	11.62	10.65	3.71		4.98		
Ho	0.57	1.19	0.92	1	1.6	0.79	0.82	0.82	0.97	1.7	5.89	0.89	2.6	2.29	0.66		1		
Er	1.76	3.66	2.53	2.9	4.8	2.26	2.35	2.37	2.93	4.3	18.37	2	7.48	7.35	1.66		2.95		
Yb	1.88	3.8	2.63	2.9	5.4	2.16	2.21	2.26	3.11	3.4	19.54	1.24	7.99	7.99	1.46		2.88		
Lu	0.3	0.58	0.42	0.4	0.7	0.36	0.35	0.36	0.47	0.4	2.72	0.16	1.25	1.15	0.22		0.41		
Hf			4.6			5.5		5	3.9		4.9	3.7	4.2	2.9	4.9		4.4		
Ta	1.4	1.4	1.1	2	3	1.1	1.4	1.4	2.9	1.5	6.6	1.8	2.3	3.1	1.8		1.1		
Pb	22.1	15.1	16.3	30.3	24	13.9	14.6	21.5	43.9	31.5	40.5	26.4	25.6	24	11	24	18.8		
Bi	0.3	0.2	0.1	0.2	0.1	0.1	0.2	0.4	1.4	0.2	0.9	1	0.2	0.4	0.2	<5	0.6		
Th	10.8	9	16.2	23	15.5	14.7	11.1	20.7	30	22.2	53.1	20.5	22.3	22.1	15.2	10	22.2		
U	6.78	5.98	4.27	10	15.1	2.71	2.14	3.81	15.58	2.9	23.4	9.65	6.09	6.52	5.76	<10	7.75		
Unit	Western Tasmanian granites										Western Tasmanian sedimentary successions							Amphibolites	
	HT	WF	SC	HW	HR	SA	CX	sh	qe	uo	oh	bh	ps	ff	su	bf	fa	aa	
Type	IFu	IFu	SFf	SFf	IFf	IMu	SFf												
SiO ₂	72.71	70.28	72.15	75.18	74.43	68.15	75.10	74.97	88.82	83.17	79.84	83.14	90.89	70.84	78.35	89.08	48.12	49.8	
TiO ₂	0.40	0.49	0.26	0.23	0.31	0.54	0.22	0.60	0.42	0.55	0.61	0.39	0.37	0.68	0.32	0.29	1.23	2.13	
Al ₂ O ₃	13.26	13.96	14.60	12.88	12.53	14.35	13.83	10.38	6.75	7.99	9.92	5.27	5.13	14.15	10.50	3.71	14.69	13.23	
Fe ₂ O ₃	0.99	0.90	0.41	0.46	0.77	1.45	0.26	0.82	0.30	0.79	0.98	0.16	0.01	1.44	1.79	2.16	2.65	3.28	
FeO	1.24	2.02	1.19	1.26	1.34	1.80	1.00	3.31	0.24	0.97	2.31	3.17	0.26	3.13	1.53	0.99	8.99	11.02	
MnO	0.03	0.06	0.03	0.04	0.04	0.06	0.06	0.06	0.00	0.02	0.03	0.10	0.00	0.07	0.06	0.02	0.20	0.18	
MgO	0.49	0.74	0.42	0.30	0.31	1.59	0.39	1.85	0.38	0.73	1.06	1.12	0.30	1.36	0.85	0.99	8.95	4.85	
CaO	1.54	1.84	0.82	0.73	0.58	2.70	1.13	0.00	0.01	0.04	0.01	0.00	0.01	0.19	1.15	0.05	9.76	9.01	
Na ₂ O	2.89	2.63	2.96	2.56	2.78	3.04	3.58	1.39	0.03	0.02	0.08	0.04	0.12	1.70	2.77	1.72	2.50	2.93	
K ₂ O	4.94	4.77	4.83	4.90	5.14	4.07	3.86	1.80	1.64	2.11	2.26	1.27	1.10	2.94	1.55	0.03	0.33	0.22	
P ₂ O ₅	0.07	0.15	0.24	0.10	0.06	0.27	0.07	0.17	0.02	0.05	0.06	0.06	0.01	0.09	0.05	0.00	0.12	0.19	
SO ₃ tot									0.06				0.09		0.06				
rest	0.15	0.16	0.10	0.07	0.08	0.26	0.10	0.12	0.07	0.10	0.12	0.11	0.11	0.16	0.14	0.09	0.06	0.19	
O = S, F																			
LOI	1.25	1.90	1.97	1.18	1.57	1.64	0.42	4.33	1.22	2.91	2.84	4.96	1.67	3.17	0.96	0.66	1.66	2.76	
Total	99.96	99.90	99.98	99.89	99.94	99.92	100.02	99.80	99.95	99.45	100.12	99.79	100.04	99.92	100.04	99.79	99.26	99.79	

continued

(continued)

Table 1 (Continued).

Unit	Western Tasmanian granites							Mathinna Sgp		Western Tasmanian sedimentary successions							Amphibolites	
	HT	WF	SC	HW	HR	SA	CX	sh	qe	uo	oh	bh	ps	ft	su	bf	fa	aa
ASI	1.028	1.088	1.260	1.188	1.120	1.002	1.142	2.451	3.687	3.344	3.821	3.658	3.648	2.237	1.261	1.256	0.661	0.617
Li	32.5	35	187	107.5	56	68.5		38		20.5	21.5	18		47		4		
Be	3	3.5	11.6	8.9	3.4	6.2	10.6	1.6	1.3	1.3	1.6	0.9	0.9	2	3.2	3	0.6	1
F	1268	1081	2072	1961	2060	966	579	848	557	680	600	539	408	940	730		900	400
S	60	270	60	70	80	250		70		1360	100	380		330		90	4340	420
Sc	6	8	6	6	7	9	7	8	8	8	8	3	6	13	8	6	42	41
V	26	32	14	10	14	59	14	59	42	41	41	29	29	62	22	49	305	419
Cr	1	3	4	2	<1	39	6	55	47	31	37	85	32	59	14	22	393	30
Ni	6	5	5	5	4	22	14	21	10	13	16.5	84	15	24	16	2	137	34
Cu	3	10	3	2	2	10	5	14	7	11	13	9	7	41	10	8	88	264
Zn	19	60	42	23	35	56	35	62	7.5	20	20	19	3.5	62	81.3	10	102	102
Ga	16.1	17.1	20.8	17.4	17.9	18.3	18	13.1	7.9	10.5	12.1	7.2	6.1	18.9	11.6	6	14.6	20.4
Ge	1.6	2	3.1	2.6	2.1	1.9	2.7	1.7	1.7	1.9	1.9	1.9	1.6	2.3	1.9	2	1.4	1.7
As	1.1	1.1	8.4	1.8	1.3	1.6	1.6	3.5	<0.5	4.5	2.8	2.3	<0.5	16.1	0.8	1	<0.2	2.1
Rb	229.8	228.7	416.2	362.6	320.5	211	442.7	70.2	82	81.4	89.1	62.6	42.8	130.6	107	0.6	5.9	72.1
Sr	87.9	150.1	51.9	33.4	27.5	472.5	58.7	37.4	33.8	33.9	10.8	14.3	8.8	53.1	141.5	3.8	72.1	266.7
Y	52.7	42.7	23.1	37.2	67.4	26.5	73.1	25.4	23.1	31.5	24	22.5	26.1	24.8	35.9	20	28.5	37.5
Zr	198	223	105	113	197	197	86	297	399	333	337	417	479	274	276	222	73	150
Nb	16.5	17.9	23.9	17.7	20.1	30.8	15.4	11.3	9.3	13.3	12.2	8.2	8.6	14.6	9.8	5	10.9	10.3
Mo	0.8	2	0.7	1.8	1	0.8	1.1	0.2	0.9	0.4	0.6	12.8	5	0.6	2	2	0.6	0.7
Ag	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	0.03	<0.5	0.03	<0.5	<0.5	<0.5	0.06	<0.5	0.03	<1	<0.5	<0.5
Cd						<0.1	<0.1		<0.1									
Sn	4	3	18	3	5	3	19.8	3	3	2	2.5	1	2.5	3	3.1	<2	0.5	1.5
Sb	0.2	<0.2	<0.2	<0.2	0.4	<0.2	0.2	0.6	1.4	0.8	0.3	0.4	0.4	0.6	0.2		<0.1	0.3
Cs	4.4	13.6	49.5	24.1	15.7	15.4	30.41	5.6	2.81	6.7	4.1	3.8	1.52	10.9	8.46	<3	<0.1	<0.1
Ba	490	606	204	167	141	889	160	418	250	460	283	245	155	631	334	49	114	50
La	97.9	54.3	25.3	28.7	63.5	71.7	9.3	36.5	30.2	37.4	40.2	30.2	33.6	34.5	29.9	12	8.6	11.8
Ce	188.2	115.9	57.9	65.0	136.1	143.2	20.7	75.7	61.5	79.1	88.5	64.0	70.9	77.8	72.9	34	19.4	29.5
Pr	18	12.5	6.8	7.5	14.4	15	2.38	9	6.62	9.2	9.5	7.4	7.59	8.7	8.04	3	2.4	4
Nd	57.1	44.7	25.3	27.4	50.9	52.4	11.1	32.9	24.79	33.9	35	27.3	28.34	31.2	30.68	13	11.2	18.8
Sm	9.2	8.6	5.8	6.3	10.3	8.4	4.11	6.4	4.94	6.3	6.4	5.1	5.41	5.8	6.62		3.1	5.2
Eu	0.90	1.30	0.60	0.40	0.60	1.70	0.29	1.20	0.86	1.00	1.10	0.90	0.92	1.00	1.25		1.10	1.80
Gd	7.8	7.7	5.3	6	10.4	6.3	6.75	5.8	4.22	5.4	5.2	4.5	4.42	4.8	6.25		4.3	6.7
Tb	1.3	1.3	0.9	1.1	1.8	0.9	1.69	0.9	0.72	0.9	0.8	0.7	0.74	0.8	1.11		0.8	1.1
Dy	7	6.8	4	5.9	9.7	4.3	11.18	4.5	3.71	4.9	4.5	3.7	3.97	4.3	6.02		4.8	6.7
Ho	1.5	1.4	0.7	1.2	2	0.8	2.49	0.9	0.75	1	1	0.7	0.82	0.9	1.34		1.1	1.5
Er	4.5	4.1	1.9	3.5	6.1	2.3	7.26	2.5	2.1	3	2.8	2.2	2.44	2.7	4.09		3.3	4.3
Yb	4.4	3.9	1.7	3.6	5.9	2.3	7.74	2.3	2.17	3	2.8	2.2	2.47	3.1	4.11		3.1	3.4
Lu	0.7	0.6	0.2	0.5	0.9	0.3	1.27	0.3	0.33	0.5	0.4	0.3	0.39	0.5	0.63		0.5	0.5
Hf							3.8		10.6		6		12.4		7.6	6	0.7	1.2
Ta	2	2	4	3	2.5	3	4.9	1	0.8	1.5	1.5	1	0.8	1.5	1.3	<2	1	1.1
Pb	9.5	16	30	23	22.5	37.5	37.9	15.5	28.6	4.5	6	2.5	3	19.5	22.3	2	2	8

(continued)

Table 1 (Continued).

Unit	Western Tasmanian granites						Mathinna Sgp						Western Tasmanian sedimentary successions						Amphibolites	
	Western Tasmanian granites			Mathinna Sgp			Western Tasmanian sedimentary successions			Amphibolites										
	HT	WF	SC	HW	HR	SA	CX	sh	qe	uo	oh	bh	ps	ff	su	bf	fa	aa		
Bi	<0.1	0.5	0.3	<0.1	0.2	0.1	2.5	0.4	<0.1	0.2	0.2	0.1	0.1	0.2	0.3	<2	<0.1	<0.1		
Th	59.2	25	15.9	21.6	52.3	38.8	14.2	15.9	12.5	11.8	13.7	7.1	8.3	13.2	8.3	4	1.2	2.2		
U	8	8.2	6.2	18.2	11.9	15.1	7.2	2.5	1.76	2.9	3.2	2.2	2.01	4.2	1.94		0.5	0.5		

Abbreviations and sample numbers as in Table 2. Grid coordinates can be found in Black *et al.* (2004, 2005) and Turner *et al.* (1998). Sample SZ (f) is the main felsic phase of the Strzelecki Granite, same location as sample SZ.

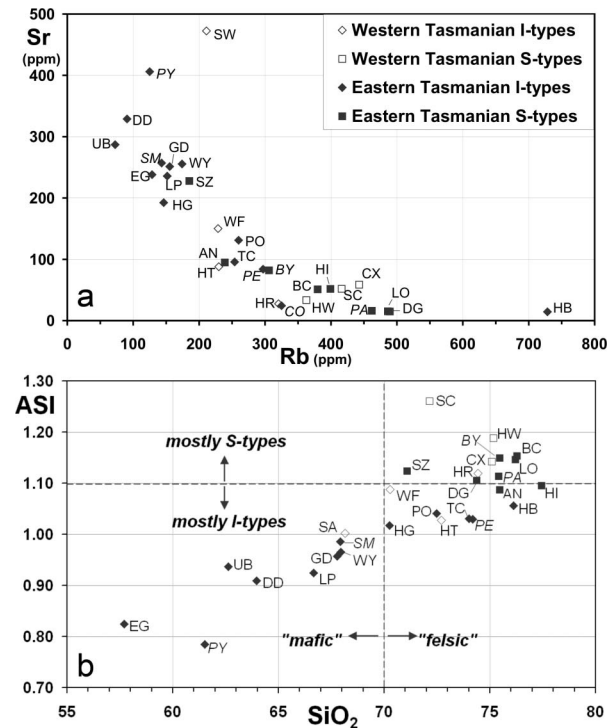


Figure 4 (a) Plot of Sr against Rb; for the granite samples studied (labels as in Table 2) and for those with previously published Sr–Nd isotope data (labels as in Table 3). (b) Plot of Alumina Saturation Index (ASI, molar $\text{Al}_2\text{O}_3/(\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO})$) against SiO_2 .

and have therefore been interpreted as early crystallites from that magma.

The paucity of datable inherited zircon has made it particularly difficult to interpret the significance of some age spectra; for example, only two inherited grains could be dated from the Henbury Granite. In an effort to increase confidence in interpretation, data from related intrusions have been assessed both individually, in combination, and on a regional basis.

Individual ages are represented in two categories on the cumulative probability diagrams (Figures 5–9). Those that are no more than 10% discordant are more heavily shaded than those that are more than 10% discordant. Because the latter analyses reflect a higher degree of radiogenic Pb loss, they are intrinsically less authoritative than more concordant analyses. If, however, Pb loss occurred only in recent times, $^{207}\text{Pb}/^{206}\text{Pb}$ ages (used here to derive ages greater than 1100 Ma, with $^{206}\text{Pb}/^{238}\text{U}$ ages being used to establish younger ages) will not have been significantly changed. Even if relatively ancient Pb loss has occurred, the inclusion of less concordant data will allow a qualitative appraisal of age components. For example, less concordant data are important in demonstrating a general presence of Archean zircon; if they were excluded, evidence of this would be a considerably less convincing.

In addition to a simple visual assessment, we have also used the method of Sircombe & Hazelton (2004) to obtain a statistically based comparison of the age profiles. The kernel functional estimates obtained by

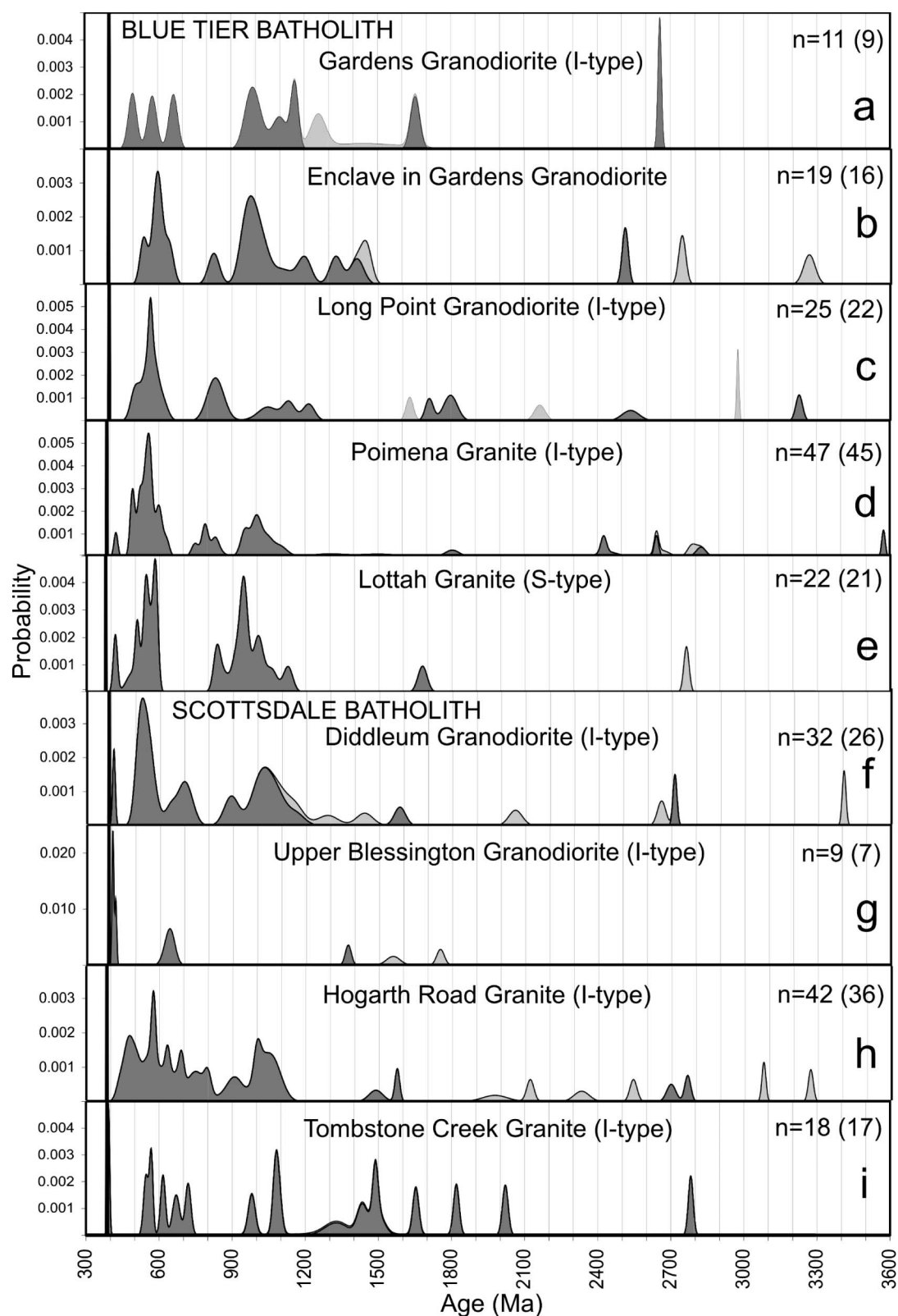


Figure 5 Probability age distribution diagrams for inherited zircon from granites in the Blue Tier Batholith and the Scottsdale Batholith. The total number of individual ages used in the construction is indicated, with the number of analyses that are less than 10% discordant given in brackets. The discordant analyses are represented by paler shading. The black bars at the left hand side show the magmatic ages.

this method for each set of age data compensate for different degrees of measurement error by applying different levels of smoothing. The dissimilarity between

these estimates provides a measure of the difference between the age spectra. A Monte Carlo permutation test has also been applied to test for equivalence

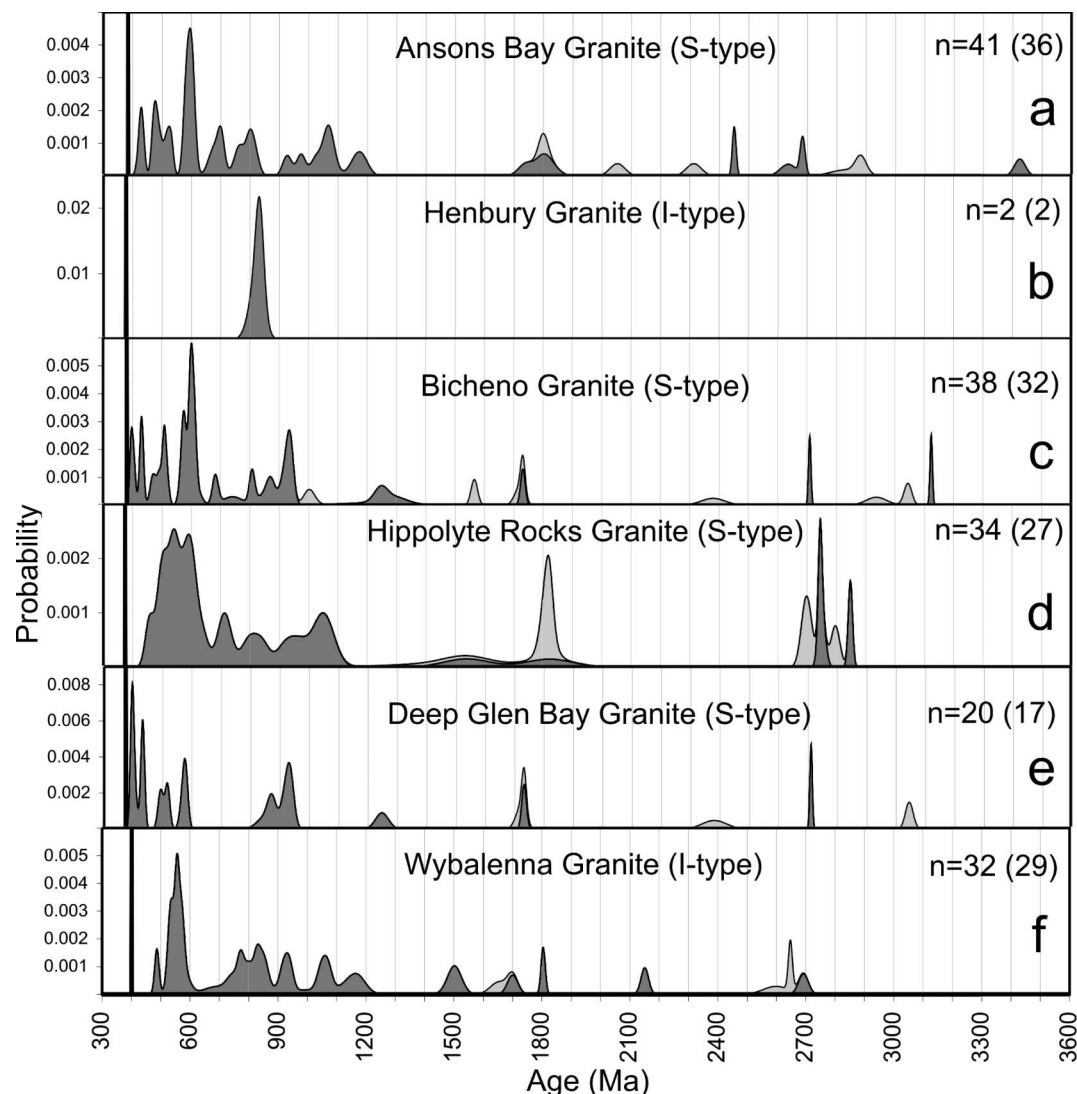


Figure 6 Probability age distribution diagrams for inherited zircon from other granites in eastern Tasmania.

between the age spectra, by grouping pairs of age data and resampling the age distributions.

Eastern Tasmania

Inheritance age spectra have been determined from 15 granites and one enclave from eastern Tasmania (Figures 5, 6). Partly because of the difficulty in interpreting a limited number of analyses, we cannot conclude, on purely visual grounds, that any of the 16 individual age patterns are significantly different from each other. On that basis, combined spectra for the Blue Tier Batholith (five samples) and Scottsdale Batholith (four samples) have been compiled (Figures 7a, b), but these fail to distinguish any significant differences between the two batholiths. The same can be said for the other seven dated granites from eastern Tasmania (Figure 6).

The similarity of detrital zircon age spectra in eastern Tasmanian granites is also demonstrated by kernel functional estimation (Sircombe & Hazelton 2004). The upper right half of Figure 10 is composed of

P values for testing equality of age distributions. Shaded fields represent P values of greater than 0.05, which indicate that the null hypothesis of equal age distributions between the pair of samples cannot be rejected at the 5% significance level. The plethora of shaded fields for all eastern granites (the rows between Gardens and Tombstone Creek) on Figure 10 testifies to the similarity of their inherited zircon age populations. The lower left half of Figure 10 quantifies the difference between samples, which is also illustrated graphically by Figure 11.

A more fundamental question is whether the zircon age spectra discriminate between granites of I- and S-type, which have been considered fundamentally different (e.g. Chappell & White 1992). Figure 7 (c, d) demonstrates a remarkable similarity between the age spectra of the two types, both in terms of populations present and their relative proportions. This conclusion is supported by Figure 12, which shows that the P value for a comparison between I- and S-type granites in eastern Tasmania is 0.10, a value that is not significant at the 95% confidence level.

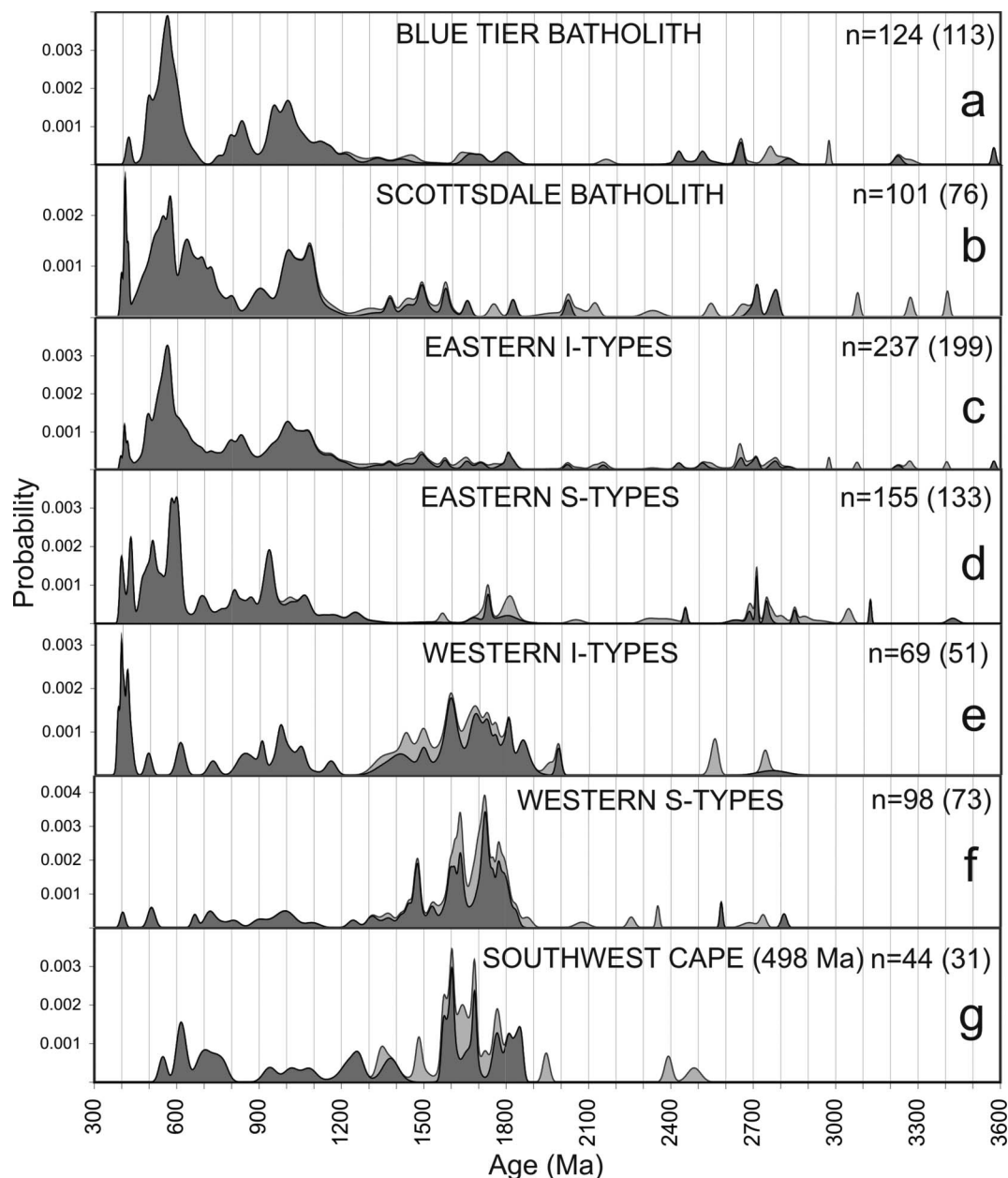


Figure 7 Comparison of combined inherited zircon age distributions for granites from different geological entities and regions of Tasmania. (a) Blue Tier Batholith. (b) Scottsdale Batholith. (c) I-type granites from eastern Tasmania. (d) S-type granites from eastern Tasmania. (e) I-type granites from western Tasmania. (f) S-type granites from western Tasmania. (g) ~498 Ma South West Cape Granite, southwestern Tasmania.

The combined inherited zircon age spectrum of eastern Tasmanian granites (Figure 9a) resembles that of detrital zircon from their Mathinna Supergroup country rocks (Black *et al.* 2004; Figure 9b). Both are dominated by ~500–600 Ma zircon, which is also prominent in the Lachlan Orogen turbidites (Ireland *et al.* 1998), although the latter also contains a much larger 1000–1200 Ma component. This suggests that either anatexis of the Mathinna Supergroup played an important role in the production of the granites, or that they were derived from a common source.

Kernel function estimates (Figure 12) and the derived dendrogram (Figure 11, inset) quantify the

relative similarity of the Mathinna Group and eastern Tasmanian granite spectra, although they cannot be said to be the same at the 95% confidence level. To assess differences, the critical (youngest) parts of their age spectra have been expanded (Figure 9f, g). Compared with the Mathinna Supergroup, the granites appear to be relatively enriched in ~975–1075 Ma zircon, and possibly also in ~800 Ma zircon. These could be discrete features, or might form part of a general enrichment in pre-600 Ma zircon, particularly up to about 1100 Ma. However, particularly as the Mathinna Supergroup spectrum is based on only two samples (Black *et al.* 2004), these results cannot be confidently

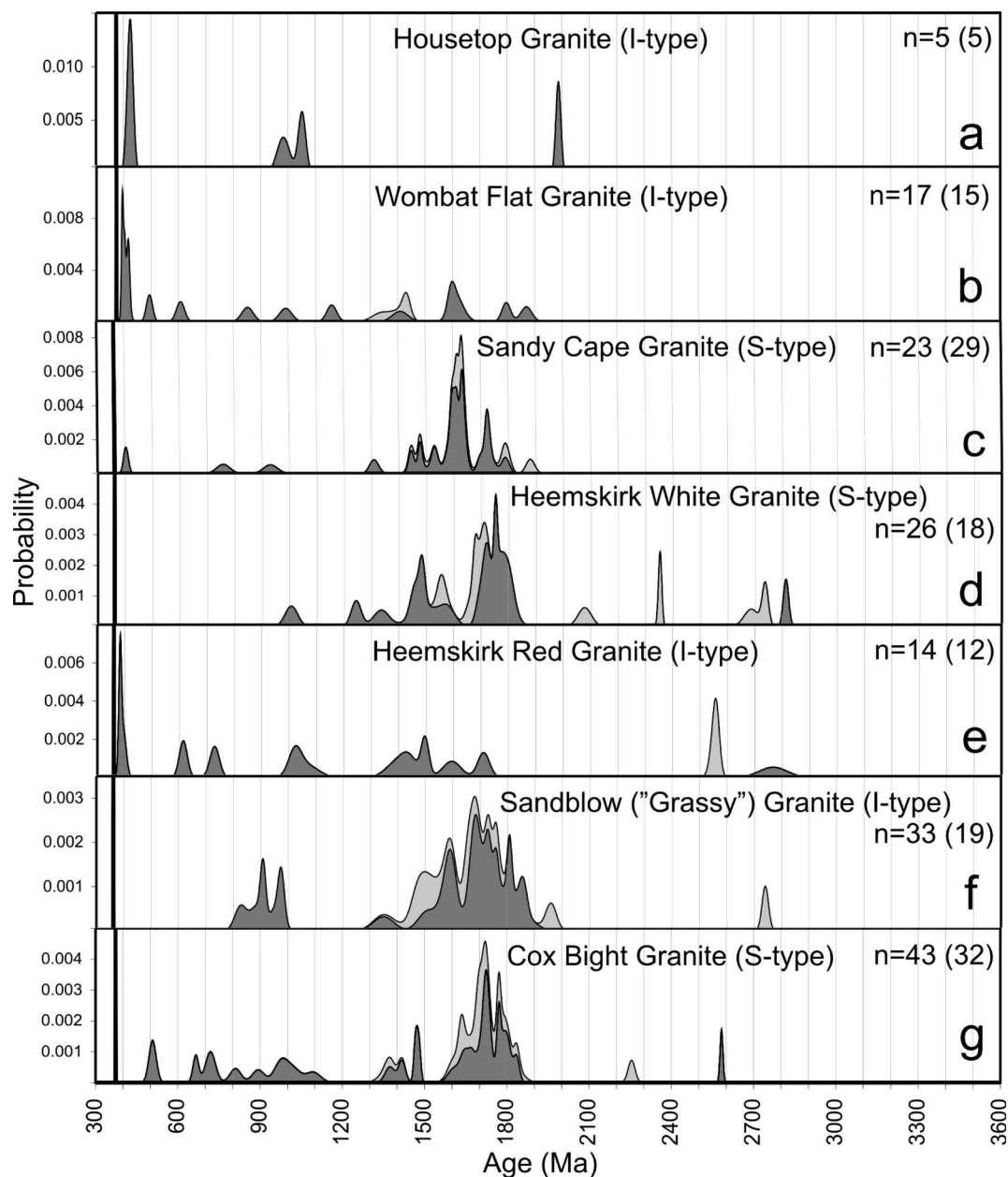


Figure 8 Probability age distribution diagrams for inherited zircon from granites in western Tasmania.

used to define any old, deep-crustal source for the eastern Tasmanian granites. Quite possibly, there was no such component, or it was too mafic to contribute significant zircon. Further data from the Mathinna Supergroup are needed.

Western Tasmania

Simple visual assessment shows that the inherited zircon age spectra of western Tasmanian granites (Figure 7) also share many common peaks, especially when samples that yielded the most grains are considered. As was found for eastern Tasmania, S- and I-type granites of this region have similar zircon age distributions to each other (Figures 8e, f), once again reflecting a distinct genetic relationship between these two granite

types. Any slight differences between these two western Tasmanian distributions probably reflect the smaller number of individual ages being compared. It is noteworthy that the ~498 Ma South West Cape Granite, in the extreme southwest of the island, yields a similar inheritance pattern (Figure 7g) to that of the other, younger, western Tasmanian granites, including the 351 Ma Sandblow Granite from King Island (Figure 8f), 420 km away.

The kernel functional estimation of Sircombe & Hazelton (2004) supports these conclusions. The plethora of shaded fields (upper right of Figure 10), clearly demonstrates that it is not statistically possible to distinguish any of the eight western Tasmanian inheritance spectra (rows Sandy Cape to Southwest Cape) from each other. The *P* value of 0.12 recorded in

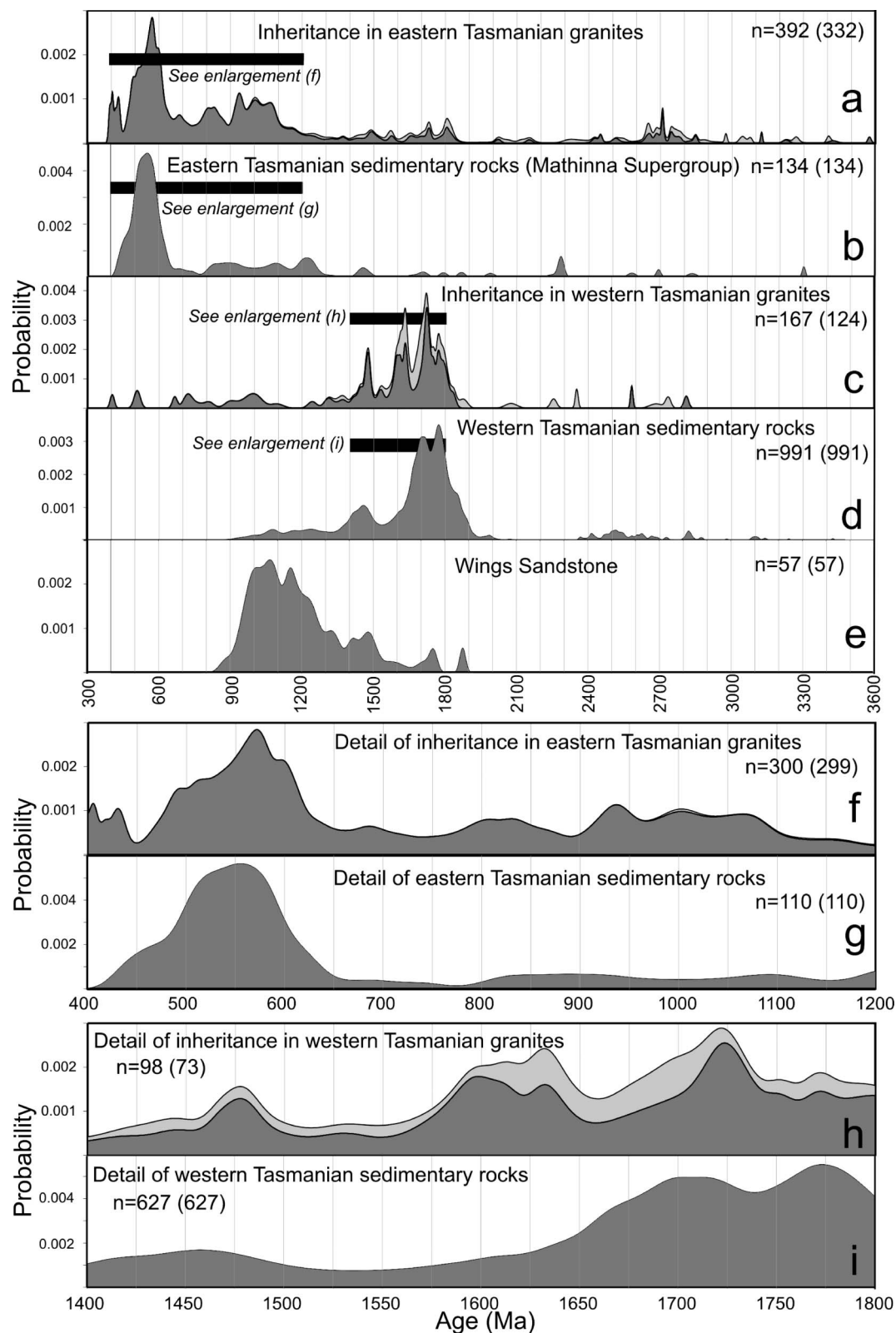


Figure 9 Comparison of age distributions for inherited zircon from granites with detrital zircons from sedimentary rocks. (a) Pooled zircon inheritance in granites from eastern Tasmania. (b) Detrital zircons from sedimentary rocks from eastern Tasmania (Black *et al.* 2004). (c) Zircon inheritance in granites from western Tasmania. (d) Detrital zircons from sedimentary rocks from western Tasmania studied by Black *et al.* (2004). (e) Detrital zircons from Wings Sandstone (western Tasmania: Black *et al.* 2004). (f) 400–1200 Ma inherited zircon from eastern Tasmanian granites. (g) 400–1200 Ma detrital zircon from sedimentary rocks in eastern Tasmania. (h) 1400–1800 Ma inherited zircon from western Tasmanian granites. (i) 1400–1800 Ma detrital zircon from sedimentary rocks in western Tasmania.

c=170288	SC	WF	HT	HW	HR	SA	CX	SW	GD	EG	LP	PO	LO	DD	HG	UB	AN	HB	BC	HI	DG	SZ	WY	TC
Sandy Cape		0.007	0.142	0.483	0.166	0.060	0.108	0.064	0.001	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0.02
Wombat Flat	13.84		0.71	0	0.47	0	0.056	0.126	0.145	0.037	0.024	0.011	0.011	0.010	0.003	0.052	0.001	0.010	0.016	0.004	0.071	0.006	0.009	0.734
Housetop	17.80	4.35		0.036	0.423	0.026	0.171	0.288	0.922	0.788	0.954	0.927	0.720	0.848	0.844	0.788	0.985	0.950	0.894	0.951	0.942	0.910	0.956	0.590
Heemskirk White	2.34	15.13	18.90		0.063	0.337	0.128	0.031	0	0.001	0	0	0	0	0	0	0	0	0	0	0	0	0	0.010
Heemskirk Red	12.70	6.20	7.85	13.07		0.021	0.276	0.266	0.355	0.199	0.267	0.125	0.044	0.075	0.045	0.099	0.032	0.290	0.034	0.048	0.052	0.059	0.139	0.705
Sandblow	1.80	13.20	16.97	2.56	11.82		0.269	0.123	0.001	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0.02
Cox Bight	5.84	8.56	12.20	6.85	7.45	4.87		0.830	0.017	0.001	0.001	0	0	0	0	0	0	0.05	0	0	0	0	0	0.2
SW Cape	7.15	7.38	11.04	8.50	7.33	6.46	2.14		0.04	0	0	0	0	0	0	0.01	0	0.07	0	0	0	0	0	0.34
Gardens	17.08	5.97	4.59	18.03	6.88	16.30	11.49	10.11		0.903	0.637	0.429	0.473	0.537	0.453	0.081	0.491	0.405	0.211	0.263	0.201	0.763	0.799	0.683
Gardens enclave	18.73	7.78	6.07	19.87	9.75	18.03	13.34	11.67	3.02		0.185	0.136	0.446	0.354	0.239	0.060	0.104	0.419	0.062	0.041	0.067	0.525	0.451	0.317
Long Point	18.52	6.03	3.30	19.41	7.41	17.68	12.85	11.74	3.44	5.41		0.680	0.240	0.351	0.307	0.032	0.801	0.337	0.326	0.774	0.385	0.473	0.823	0.361
Poimena	21.45	8.28	4.43	22.41	10.31	20.62	15.77	14.57	5.45	6.08	3.18		0.273	0.701	0.748	0.121	0.962	0.143	0.838	0.952	0.829	0.236	0.442	0.160
Lottah	22.33	9.56	6.43	23.61	12.89	21.63	16.87	15.23	6.42	4.71	6.30	4.63		0.866	0.732	0.077	0.240	0.393	0.249	0.222	0.374	0.460	0.538	0.119
Diddleum	22.16	9.12	5.68	23.36	12.12	21.44	16.61	15.09	5.92	4.86	5.27	3.27	1.42		0.970	0.097	0.344	0.118	0.479	0.508	0.576	0.470	0.634	0.118
Hogarth Road	22.22	9.10	5.51	23.37	11.91	21.47	16.63	15.17	5.89	5.15	4.91	2.66	2.08	0.70		0.108	0.711	0.260	0.696	0.641	0.557	0.381	0.585	0.064
Upper Blessington	25.21	11.40	8.09	26.48	15.23	24.56	19.88	18.59	10.88	10.69	8.73	6.25	6.83	6.18	5.97		0.003	0	0.320	0.157	0.683	0.014	0.018	0.089
Ansons Bay	20.31	7.18	3.40	21.29	9.36	19.47	14.62	13.42	4.59	5.52	2.22	1.19	4.78	3.56	3.08	6.84		0.041	0.809	0.954	0.732	0.111	0.594	0.093
Henbury	24.72	13.13	10.18	25.95	15.71	23.96	19.34	17.63	8.89	6.43	9.50	7.95	4.33	5.27	5.66	10.21	8.24		0.190	0.219	0.360	0.015	0.397	0.306
Bicheno	22.53	9.01	5.29	23.66	12.01	21.78	16.95	15.60	6.76	6.61	4.87	2.19	3.55	2.32	1.83	4.53	2.84	7.23		0.906	0.971	0.118	0.170	0.030
Hippolyte Rocks	20.90	7.72	3.96	21.86	9.76	20.06	15.22	14.08	5.30	6.26	2.75	0.96	5.13	3.80	3.22	6.40	1.00	8.55	2.61		0.763	0.227	0.436	0.108
Deep Glen Bay	22.47	8.75	4.97	23.59	11.84	21.72	16.92	15.67	7.26	7.50	5.08	2.57	4.72	3.53	3.10	3.81	3.12	8.55	1.49	2.71		0.258	0.279	0.059
Strzelecki	19.37	6.69	4.10	20.63	10.15	18.66	13.88	12.28	3.97	3.04	4.40	4.32	3.02	2.94	3.24	7.96	3.72	6.56	4.27	4.43	5.00		0.708	0.243
Wybalenna	18.68	6.12	3.37	19.82	8.90	17.89	13.05	11.59	2.83	3.06	3.15	3.81	4.11	3.63	3.65	8.52	2.95	7.30	4.51	3.74	5.14	1.59		0.356
Tombstone Creek	12.78	3.38	5.66	13.79	4.21	12.00	7.15	5.91	4.54	6.78	5.93	8.76	10.04	9.62	9.60	13.32	7.65	12.87	9.98	8.29	10.06	7.07	6.11	

concordant data only
shaded: $P > 0.05$
5

Figure 10 Matrix of comparison between the age spectra from different Tasmanian granites, based on concordant data only, following the kernel functional estimation method of Sircombe & Hazelton (2004). The value of c for this comparison is 170288. The values in the lower left half represent the calculated distances between the age-distribution estimations multiplied by 1000. The values in the upper right half are the probability P for testing equality of age distributions between each pair of samples. Shaded fields represent P values of greater than 0.05, which indicate that the null hypothesis of equal age distributions cannot be rejected at the 5% significance level.

Figure 12 for the comparison between western S- and I-type granite inheritance spectra fails to identify any significant difference at the 95% confidence level. Again, the combined detrital zircon spectrum of the sedimentary country rocks (from Black *et al.* 2004) is similar to, but distinct from, the granite spectra (Figure 11 inset, Figure 12).

However, inheritance age patterns of the western Tasmanian granites are dramatically different from those of eastern Tasmania. Whereas in the east, ages of about 500–600 Ma and about 900–1100 Ma are dominant (Figure 9a), western Tasmanian granites are characterised by 1450–1850 Ma inherited zircon (Figure 9c). A distinction of that magnitude represents a fundamental difference in the origin of the granites from the two regions, and the participation of different source components in their genesis. The inter-regional differences are demonstrated statistically in Figure 10, in which most unshaded fields are between individual eastern and western Tasmanian granites. This is illustrated graphically by the derived dendrogram (Figure 11), in which the granites are ordered statistically, and all but one of the western granites are grouped together at the top. The apparently anomalous position of the Housetop Granite, and to a lesser extent, the Heemskirk Red Granite and Wombat Flat Granite, probably represents a sampling problem, as these samples yielded the lowest number of inherited grains (5, 14, and 17, respectively). A statistical analysis of the combined spectra (Figure 12) confirms that, irrespective of whether I- or S-type granites are being compared, the chance of the western and eastern inheritance patterns being equivalent to each other is less than 0.1%.

Detrital zircon age spectra for sedimentary units from western Tasmania (Black *et al.* 2004) are mostly dominated by 1650–1900 Ma zircon (Figure 9d), which is very largely absent from the Mathinna Supergroup of eastern Tasmania (Figure 9b). However, a prominent concentration of similar ages was obtained from inherited zircon within western Tasmanian granites (compare Figures 9c and 9d). As was the case for eastern Tasmania, this strongly suggests that associated sedimentary successions (or possibly their precursors) played an important role in granite genesis.

The western Tasmanian granites contain a range of 400–750 Ma zircon (Figure 9c) that is almost entirely too young to be derived from the Proterozoic sandstones studied by Black *et al.* (2004), the youngest of which is probably the Middle Cryogenian (?) Port Sorell Formation. Possible local sources for this zircon are felsic magmatism at 750–780 and 580 Ma, both on King Island and mainland Tasmania, and the more voluminous Mt Read Volcanics and associated granitoids at 493–505 Ma (see above).

One difference, between the western Tasmanian granite (Figure 9c) and combined sedimentary (Figure 9d) age spectra, is that there appear to be relatively more ~1000 Ma ages in the former. In the latter, these are mainly from the Wings Sandstone in southwest Tasmania which, unlike other western Tasmanian Proterozoic successions, has a dominant ~1200–1000 Ma component (Figure 9e). This suggests provenance from a Grenville-aged terrane, such as the Albany-Fraser and Musgrave provinces in southern and central Australia, respectively. Granite of similar age has been documented recently from the South Tasman Rise (Fioretti *et al.* 2005). Similar aged zircons

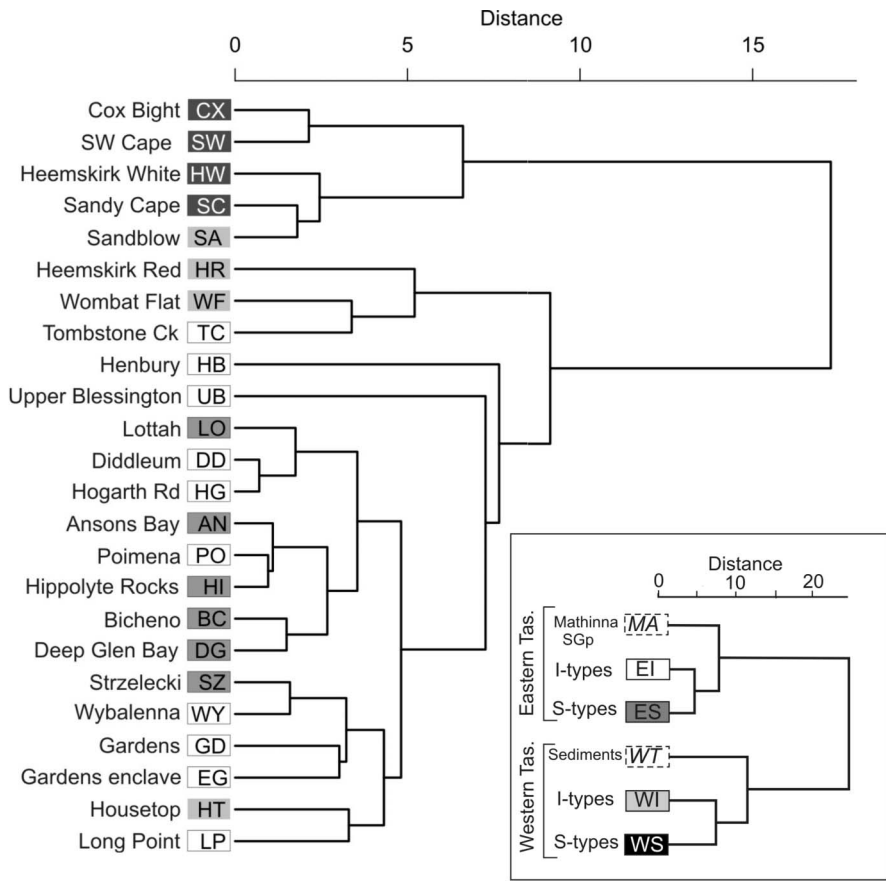


Figure 11 Dendrogram showing the degree of similarity between the inherited zircon age spectra of Figure 10. Clustering was achieved using an averaging algorithm, based on the distance values in the lower left half of the matrix (Figure 10). Inset: dendrogram showing comparison of eastern I-types, eastern S-types, western I-types and western S-types, based on matrix of Figure 12.

c=69218	EIG	ESG	MA	WIG	WSG	WSS
Eastern I-type granites		0.091	0	0	0	0
Eastern S-type granites	3.46		0	0	0	0
Mathinna Supergroup	6.49	6.88		0	0	0
Western I-type granites	16.94	15.32	20.13		0.136	0
Western S-type granites	22.75	21.24	26.15	6.24		0
Western sedimentary successions	29.53	27.60	32.66	12.97	7.70	

Figure 12 Matrix of comparison between the age spectra of I- and S-type granites and their sedimentary country rocks, in both eastern and western Tasmania, based on concordant data only.

have played an even more important role in eastern Tasmania (see above).

The most important difference, however, between the western Tasmanian granite and sedimentary age spectra (Figures 9c, d) occurs in the main data concentrations between ~1600 Ma and ~1900 Ma, which are enlarged in Figure 9 (h, i). It is important to note that the positions of the main age concentrations in the granites are unaffected by the inclusion of discordant data, presumably because significant radiogenic Pb loss has only occurred relatively recently. The granites contain a large possibly bimodal proportion of ~1600 Ma zircon, which may define separate events at ~1600 Ma and ~1630 Ma. Similar aged zircon is far less common in the sedimentary rocks, although there are slight to moderate concentrations in the Surprise Bay Formation, the Badger Head Group, the upper Oonah Formation and the Port Sorell Formation (Black *et al.* 2004).

Events at ~1600 Ma have not been previously noted from the geological record in Tasmania. This age

corresponds, however, with the voluminous Gawler Range Volcanics and coeval Hiltaba Suite granites of the Gawler Craton in South Australia (Creaser & White 1991; Creaser 1995). This felsic activity extended eastward as the Ninnerie Supersuite in the Curnamona Province (Page *et al.* 2005), and possibly southward to the Terre Adelie Craton, Antarctica (Peucat *et al.* 2002). It is therefore plausible that similar igneous rocks could underlie western Tasmania, and may have contributed ~1600 Ma zircon to the Devonian–Carboniferous granites.

However, a sedimentary origin for the ~1600 Ma zircons in the granites is also possible, since this would be expected to contribute a more complex spectral component, and our data do not entirely rule out the possibility of enrichment in zircon of other ages. Additional Paleoproterozoic components could relatively easily be obscured by the pronounced ~1700–1900 Ma components of the sedimentary age spectrum. A deep crustal but sedimentary origin for the ~1600 Ma

zircons in the granites is not inconsistent with the I-type nature of some of the Devonian–Carboniferous granites, because those characteristics could have been established by a higher proportion of a mantle-derived mafic source component.

Black *et al.* (2004) noted the general similarity of the zircon detrital age spectra from western Tasmanian Proterozoic sedimentary rocks with ages from the Wilgena, Moonta and Christie subdomains of the Gawler Craton, but discounted these as potential sources due to a dearth of ~1600 Ma zircon in the Tasmanian sequences. However, it seems plausible that the surficial Gawler Range Volcanics were eroded first, incorporating the ~1600 Ma zircons into the oldest derived sedimentary successions. Subsequent erosion of the Paleoproterozoic rocks underlying the Gawler Range Volcanics would have produced younger successions dominated by older (mainly ~1700 – ~1800 Ma) zircon; i.e. an ‘inverted zircon stratigraphy.’ In this context, it is worth noting that the Mesoproterozoic Surprise Bay Formation on King Island has peaks at about 1600 Ma and 1630 Ma, whereas zircons of this age are very rare in the other, younger, successions. It seems feasible, therefore, that many of the Mesoproterozoic to early Neoproterozoic sedimentary successions in western Tasmania could have been derived from the Gawler Craton, or an equivalent source terrane. The deepest felsic crust in the region might be composed of such rocks, or be dominated by ~1600 Ma igneous rocks.

ZIRCONIUM FRACTIONATION IN TASMANIAN GRANITES

Tasmanian granites, especially when considered in terms of suites, generally show a decline in Zr toward more felsic compositions (Figure 13). This is particularly evident in S-types (e.g. the Interview Suite from western Tasmania; Figure 13b), which probably reflects the lower solubility of zircon in peraluminous melts (Watson & Harrison 1983). Similar trends are seen in many I-type suites (e.g. Figure 13a), although in some (such as the Wybalenna Suite, Figure 13b) the decreases are slight. No Tasmanian granite shows the increase in Zr toward more evolved compositions characteristic of ‘high temperature I-types,’ typified by the Boggy Plain Suite of the Kosciusko Batholith (Wyborn *et al.* 1987; Chappell *et al.* 1998, 2000).

Some of the petrogenetic arguments made by Kemp *et al.* (2005) for I-type granites from the Lachlan Orogen, notably the Why Worrie Suite of the Bega Batholith, also apply to the Tasmanian granites. In particular, the low abundance of inherited zircon is difficult to reconcile with a restite-unmixing model.

The I-type Russells Road Suite (Figure 13a) from the Scottsdale Batholith, from which we have examined inherited zircon from three contrasting plutons, is a useful example. This suite shows linear trends on many simple two-element variation diagrams, including SiO_2 –Zr (Figure 13a), similar to those interpreted by Chappell

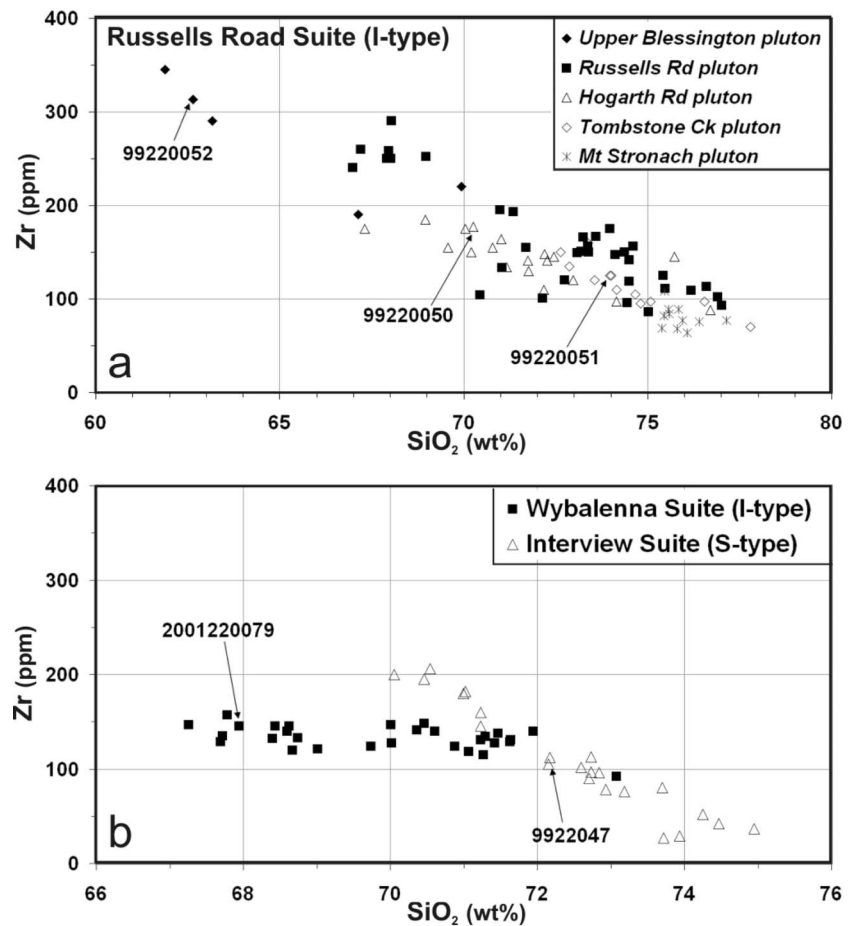


Figure 13 Plots of zirconium (ppm) against SiO_2 (wt%) for representative Tasmanian granite suites. (a) Russells Road Suite, with constituent plutons distinguished. (b) Wybalenna Suite (six plutons) and Interview Suite (four plutons). Each data point represents an analysed sample; those also used for SHRIMP dating are labelled.

and co-workers as indicative of restite unmixing. In this model, the most mafic rocks (such as our sample 99220052 of the Upper Blessington Granodiorite) would be interpreted as containing the most entrained restite. However, inherited zircon is rare in this rock; despite being actively sought; only nine grains were observed (Figure 5g). The more felsic Hogarth Road Granite and Tombstone Creek Granite contain a higher proportion of inherited zircon (Figures 5h, i), despite their overall lower Zr contents, consistent with a larger contribution from a sedimentary source component. A more plausible explanation for the negative correlation of Zr with SiO₂ in the Russells Road Suite is fractionation of magmatic zircon.

ISOTOPIC DATA

Results

Analytical details are given in Appendix 1 and data, with estimated uncertainties, are presented in Table 2. The isotopic ratios of the granites have been corrected to initial values at U–Pb ages on magmatic zircon, as determined by SHRIMP (Black *et al.* 2005). Implicit in this procedure is the assumption that the whole rock Rb–Sr, Nd–Sm and U–Th–Pb systems closed at effectively the same time as the U–Pb system in magmatic zircon. The validity of this assumption, particularly with regard to the Rb–Sr system in fractionated granites, is discussed below.

Several of the more felsic granites have Sm/Nd values greater than average crustal values, probably due to fractionation of REE-bearing accessory minerals (e.g. Sawka *et al.* 1990; Chappell 1999). Therefore, the conventional depleted mantle Nd model ages, which assume minimal REE fractionation since separation from mantle, are unreliable. For this reason, we have also calculated two-stage Nd model ages (Table 2), in which a crustal ¹⁴⁷Sm/¹⁴⁴Nd of 0.11 (Sm/Nd ~0.182) is assumed prior to granite generation (Liew & Hofmann 1988). We consider that these provide more realistic estimates of mean crustal residence times, especially for fractionated samples. For the sedimentary rocks, which are of interest as potential source components, we have corrected the isotopic ratios to 380 Ma, about the mean age of the granites. Initial ratios are plotted, together with selected previously published isotopic data (Table 3), in Figures 14–17.

Discussion

GENERAL OBSERVATIONS

In the Lachlan Orogen, McCulloch & Chappell (1982) showed that the Sr and Nd isotopic composition of granites from the Berridale and Kosciusko Batholiths define a hyperbolic array, from isotopically primitive I-types with low ⁸⁷Sr/⁸⁶Sr and variable ε_{Nd}, through to S-types with higher ⁸⁷Sr/⁸⁶Sr and low ε_{Nd}. The most isotopically evolved Lachlan Orogen granites are S-types associated with regional metamorphic complexes (McCulloch & Chappell 1982; Gray 1990). At the

more primitive end of the array, the I-type field has been expanded by further data, mainly from the Bega and Moruya Batholiths (Chappell *et al.* 1990; McCulloch & Woodhead 1993; Keay *et al.* 1997). McCulloch & Chappell (1982) and Gray (1984) noted that the isotopic array can be interpreted as a mixing curve between mafic mantle-derived magmas and a crustal component derived from the lower Paleozoic turbidites.

Our new isotopic data for the Mathinna Supergroup, together with that of Sun & Higgins (1996), support its broad correlation with the lower Paleozoic turbidites of the Lachlan Orogen (Gray 1990; McCulloch & Woodhead 1993). At ~380 Ma, both terranes have ⁸⁷Sr/⁸⁶Sr ratios of ~0.718 to 0.726 and ε_{Nd} of –12 to –9.5 (Figure 14).

The unfractionated or weakly fractionated I-type granites of eastern Tasmania lie close to and partly overlap the Lachlan I-type field on a plot of initial ε_{Nd} against ⁸⁷Sr/⁸⁶Sr (Figures 14, 15), although the Blue Tier and Scottsdale Batholiths each form separate arrays with a flatter slope. In contrast, there are no Tasmanian granites in the Lachlan S-type field. With one exception, the eastern Tasmanian S-types and all the western Tasmanian granites lie above and to the right of the Lachlan array, at higher (less negative) ε_{Nd} and/or higher ⁸⁷Sr/⁸⁶Sr.

The two-stage Nd model ages of eastern Tasmanian granites (1054–1390 Ma) are considerably younger than those of the Mathinna Supergroup (single stage model ages of 1871–1919 Ma) but overlap with those of the western Tasmanian granites (1138–1600 Ma), which also tend to be younger than western Tasmanian sedimentary successions (single stage model ages of 1495–1834 Ma). This could indicate a juvenile component in the granites, both in eastern and western Tasmania.

McCulloch & Woodhead (1993) showed that the initial Pb isotope ratios of about twenty mostly I-type granites from the Moruya, Bega and Berridale Batholiths, in the Lachlan Orogen, display very limited variation (Figure 16). This was attributed to homogenisation of Pb isotope compositions in the granite source rocks immediately prior to magmatism, probably by means of deep, crustal-scale advection by aqueous fluids. Hogan (1995) argued, however, that this process would be unlikely to homogenise Pb isotope compositions unless there had been ‘repeated fluid circulation, by pervasive grain-scale infiltration, in crustal-scale convection cells,’ and that Sr isotopes would also be affected.

Overall, Tasmanian granites show more variation in their Pb isotope ratios than those of the Lachlan Orogen (Figure 16). On a plot of initial ²⁰⁷Pb/²⁰⁴Pb against ²⁰⁶Pb/²⁰⁴Pb (Figure 16a), they define a broad array with positive slope. The eastern Tasmanian I-type granites from the Blue Tier and Scottsdale Batholiths again plot in separate well-defined narrow fields, with the former overlapping with that of the I-type Bega Batholith. The lead isotopic composition of eastern Tasmanian S-type granites is slightly more varied, but overlaps with that of the I-types. The greatest dispersal is displayed by western Tasmanian granites. Again, this could reflect the more variable and generally more radiogenic Pb isotope composition of their host sedimentary successions. Within the western group (of only seven samples) there is also no obvious distinction between the lead

Table 2 Isotopic data for granite and host rock samples.

Pluton/unit	Abb ¹	Bath ²	Type ³	GA No ⁴	Age ⁵ (Ma)	$\pm 2\sigma$	$^{87}\text{Sr}/^{86}\text{Sr}_{\text{m}}$	$\pm 2\sigma$ ($\times 10^6$)	Sr (ppm)	Rb (ppm)	$^{87}\text{Sr}/^{86}\text{Sr}_{\text{t}}$	$\pm 2\sigma$ ($\times 10^5$)	Nd ⁶ (ppm)	Sm ⁶ (ppm)
Eastern Tasmanian granites														
The Gardens	GD	BTB	IMu	99220055	400	± 2	0.716860	± 12.3	251.3	155.3	0.70667	± 29	12.83	2.95
Gardens enclave	EG	BTB	IMu	99220056	399	± 2	0.715603	± 12.4	238.1	128.6	0.70672	± 25	27.90	6.34
Long Point	LP	BTB	IMu	99220058	396	± 2	0.717754	± 12.7	235.7	151.5	0.70726	± 30	26.22	5.27
Poimena	PO	BTB	IFu	99220057	387	± 3	0.740215	± 12.7	131.0	259.9	0.70850	± 92	26.45	5.48
Lottah	LO	BTB	SFh	99220061	378	± 2	1.215587	± 16.2	14.9	477.4	0.69216	± 3681	16.61	5.28
(<i>isotope dilution</i>)							1.213840		15.3	486.3	0.69367	± 758		
Diddleum	DD	SB	IMu	99220048	390	± 2	0.711262	± 14.0	328.8	90.7	0.70682	± 13	26.75	4.73
Upper Blessington	UB	SB	IMu	99220052	389	± 2	0.711791	± 13.3	286.9	72.6	0.70773	± 12	29.67	5.56
Hogarth Road	HG	SB	IFu	99220050	388	± 2	0.721044	± 14.0	192.4	146.2	0.70889	± 35	29.44	5.49
(<i>repeat</i>)							0.720966	± 14.2			0.70881	± 35		
Tombstone Creek	TC	SB	IFu	99220051	386	± 3	0.752245	± 12.1	95.7	253.8	0.70990	± 124	22.56	4.94
Ansons Bay	AN	EB	SFh	99220054	384	± 2	0.751781	± 11.5	94.8	238.9	0.71171	± 116	44.27	10.26
Henbury	HB	BL	IFh	2001220070	377	± 3	1.518571	± 20.0	13.8	720.0	0.64518	± 6595	57.95	16.21
(<i>isotope dilution</i>)							1.524660		14.4	728.5	0.67855	± 1321		
Bichen	BC	SFf	SFf	2001220071	381	± 3	0.828671	± 12.6	51.0	379.8	0.71028	± 345	19.53	4.91
Hippolyte Rocks	HI	SFf	SFf	2002220081	374	± 3	0.820981	± 15.4	51.5	383.2	0.70514	± 337	29.01	7.41
(<i>isotope dilution</i>)							0.821460		51.7	399.1	0.70132	± 96		
Deep Glen Bay	DG	SFh	SFh	2001220075	374	± 4	1.197175	± 13.4	14.7	482.0	0.66811	± 3788	17.03	5.46
(<i>isotope dilution</i>)							1.205530		14.8	489.2	0.67317	± 890		
Strzelecki	SZ	SFh	SFh	2001220080	385	± 4	0.718007	± 12.5	227.7	185.1	0.70510	± 39	25.78	4.88
Wybalenna	WY	IMu	IMu	2001220079	401	± 4	0.719209	± 11.9	255.5	174.0	0.70796	± 34	28.84	5.90
Western Tasmanian granites														
Housetop	HT	IFu	IFu	99220042	374	± 2	0.749962	± 14.1	87.9	229.8	0.70956	± 116	45.59	8.51
Wombat Flat	WF	IFu	IFu	99220043	373	± 2	0.737589	± 12.3	150.1	228.7	0.71410	± 68	45.59	8.51
Sandy Cape	SC	SFf	SFf	99220047	362	± 2	0.877355	± 13.4	51.9	416.2	0.75566	± 350	27.06	6.04
Heemskirk White	HW	SFf	SFf	99220045	361	± 2	0.904229	± 14.8	33.4	362.6	0.73974	± 599	28.22	6.56
Heemskirk Red	HR	IFf	IFf	99220046	360	± 2	0.887226	± 11.7	27.5	320.5	0.71142	± 735	53.19	10.67
Sandblow (Grassy)	SA	IMu	IMu	99220059	351	± 2	0.718269	± 10.2	472.5	211.0	0.71181	± 19	54.09	8.37
Cox Bight	CX	SFf	SFf	2002220085	370	± 3	0.829330	± 14.1	58.7	442.7	0.71305	± 338	10.67	4.03
Eastern Tasmanian sedimentary sequences (Mathinna Supergroup) at 380Ma														
Stony Head Sst	sh			98220033	[380]		0.751672	± 14.2	37.4	70.2	0.72216	± 99	24.79	4.94
Queen of the Earth	qe			2001220078	[380]		0.756074	± 15.0	33.8	82.0	0.71792	± 136	24.79	4.94
Western Tasmanian sedimentary sequences at 380Ma														
Upper Oonah Fm	uo			99220060	[380]		0.799445	± 13.6	33.9	81.4	0.76152	± 135	35.20	6.57
Oonah Fm	oh			95220027	[380]		0.981626	± 15.1	10.8	89.2	0.84887	± 258	35.20	6.57
Badger Head Gp	bh			98220034	[380]		0.850434	± 13.6	14.3	62.6	0.78095	± 505	29.34	5.45
Port Sorell Fm	ps			2001220077	[380]		0.859083	± 19.3	8.8	42.8	0.78182	± 891	27.31	4.97
Fraser Fm	ff			98220036	[380]		0.816391	± 13.8	53.1	130.6	0.77748	± 110	27.70	5.12
Surprise Bay Fm	su			2001220076	[380]		0.762315	± 12.7	141.5	107.0	0.75041	± 34	30.18	6.42
Bowry Fm ¹⁰	bf			93220006	[380]		0.717353	± 15.7	3.8	0.6	0.71488	± 256	14.59	2.89
Western Tasmanian mafic rocks at 380Ma														
Forth MC amphibolite	fa			95220020	[380]		0.706589	± 18.7	77	6	0.70537	± 13	18.48	5.13
Arthur MC amphibolite	aa			95220018	[380]		0.708413	± 12.3	284	3	0.70825	± 04	18.48	5.13

¹Abbreviations used in Figures 4, 11–15. ²BTB: Blue Tier Batholith; SB: Scottsdale Batholith; EB: Eddystone Batholith; BL: Ben Lomond Batholith. ³I: I-type; S: S-type; M: mafic (<70% SiO₂); F: felsic (>70% SiO₂); u: unfractionated or weakly fractionated; f: fractionated (>250 ppm Rb); h: highly fractionated (>400 ppm Rb, <20 ppm Sr). ⁴Geoscience Australia Sample No. ⁵U–Pb SHRIMP age from Black *et al.* (2005), isotopic ratios of sedimentary successions calculated at 380 Ma. ⁶Nd and Sm values by isotope dilution differ slightly from ICPMS values in Table 1.

Table 2 (Continued).

$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}(\text{m})$	$\pm 2\sigma$ ($\times 10^6$)	T_{DM}^7 (Ma)	T_{DM}^8 (Ma)	ϵ_{Nd}^9	$\pm 2\sigma$	Pb (ppm)	U (ppm)	Th (ppm)	$^{206}/^{238}\text{Pb}_\text{m}$	$^{207}/^{235}\text{Pb}_\text{m}$	$^{206}/^{238}\text{Pb}_\text{i}$	$\pm 2\sigma$	$^{207}/^{235}\text{Pb}_\text{i}$	$\pm 2\sigma$	$^{208}/^{232}\text{Pb}_\text{i}$	$\pm 2\sigma$
0.1389	0.512204	± 38	1667	1340	-5.53	± 0.39	22.1	6.8	10.8	19.458	15.670	38.986	18.183	15.600	± 0.064	38.331	± 0.032
0.1373	0.512215	± 33	1616	1319	-5.24	± 0.34	15.1	6.0	9.0	19.739	15.684	39.108	18.090	15.594	± 0.116	38.307	± 0.056
0.1215	0.512144	± 62	1469	1359	-5.84	± 0.61	16.3	4.3	16.2	19.233	15.660	39.448	18.154	15.601	± 0.075	38.126	± 0.086
0.1253	0.512152	± 38	1515	1361	-5.97	± 0.38	30.3	10.0	23.0	19.478	15.679	39.144	18.149	15.607	± 0.052	38.158	± 0.039
0.1921	0.512479	± 40	3218	1140	-2.88	± 0.41	24.0	15.1	15.5	20.449	15.712	38.920	17.952	15.576	± 0.116	38.092	± 0.039
0.1069	0.512272	± 39	1115	1135	-2.67	± 0.39	13.9	2.7	14.7	19.298	15.693	39.802	18.501	15.650	± 0.071	38.406	± 0.105
0.1133	0.512258	± 57	1199	1175	-3.27	± 0.56	14.6	2.1	11.1	19.249	15.664	39.452	18.655	15.632	± 0.056	38.457	± 0.071
0.1128	0.512231	± 63	1232	1211	-3.79	± 0.62	21.5	3.8	20.7	19.202	15.641	39.383	18.487	15.602	± 0.043	38.130	± 0.064
0.1324	0.512234	± 26	1224	1205	-3.71	± 0.26				19.217	15.682	39.488	18.501	15.643	± 0.043	38.233	± 0.064
0.1401	0.512230	± 25	1503	1279	-4.80	± 0.26	43.9	15.6	30.0	19.812	15.684	39.023	18.383	15.607	± 0.045	38.134	± 0.028
0.1401	0.512222	± 37	1657	1315	-5.34	± 0.38	31.5	2.9	22.2	18.900	15.638	39.040	18.535	15.618	± 0.021	38.139	± 0.034
0.1692	0.512408	± 42	2032	1158	-3.16	± 0.43	40.5	23.4	53.1	21.717	15.780	39.852	19.365	15.652	± 0.077	38.127	± 0.056
0.1519	0.512274	± 27	1822	1284	-4.92	± 0.28	26.4	9.7	20.5	19.707	15.679	39.079	18.254	15.600	± 0.063	38.082	± 0.043
0.1546	0.512386	± 31	1643	1139	-2.92	± 0.32	25.6	6.1	22.3	19.368	15.640	39.281	18.444	15.590	± 0.042	38.187	± 0.049
0.1939	0.512457	± 33	3540	1173	-3.41	± 0.35	24.0	6.5	22.1	19.536	15.651	39.159	18.480	15.594	± 0.051	38.002	± 0.055
0.1144	0.512350	± 32	1081	1054	-1.56	± 0.33	11.0	5.8	15.2	21.099	15.775	39.858	18.935	15.657	± 0.205	38.014	± 0.173
0.1237	0.512127	± 31	1527	1390	-6.23	± 0.32	18.8	7.8	22.2	19.735	15.694	39.594	17.999	15.599	± 0.100	37.989	± 0.083
0.0968	0.512245	± 29	1058	1138	-2.91	± 0.29	9.5	8.0	59.2	23.984	15.904	47.086	20.163	15.698	± 0.410	37.952	± 0.980
0.1129	0.512178	± 32	1307	1283	-4.99	± 0.32	16.0	8.2	25.0	20.826	15.743	40.435	18.767	15.631	± 0.136	38.407	± 0.134
0.1351	0.511995	± 31	1956	1600	-9.70	± 0.31	30.0	6.2	15.9	19.990	15.698	38.991	19.208	15.656	± 0.032	38.343	± 0.025
0.1405	0.512106	± 43	1880	1467	-7.79	± 0.43	23.0	18.2	21.6	20.624	15.726	39.665	17.591	15.563	± 0.146	38.502	± 0.056
0.1213	0.512214	± 31	1360	1260	-4.81	± 0.32	22.5	11.9	52.3	20.757	15.722	41.484	18.685	15.611	± 0.102	38.539	± 0.144
0.0936	0.512146	± 32	1148	1264	-4.98	± 0.32	37.5	15.1	38.8	19.744	15.684	39.574	18.265	15.605	± 0.050	38.344	± 0.041
0.2287	0.512438	± 61		1310	-5.42	± 0.62	37.9	7.2	14.2	19.744	15.699	38.824	19.014	15.659	± 0.025	38.359	± 0.016
0.1263	0.511936	± 47	1871		-10.29	± 0.94	15.5	2.5	15.9	18.946	15.679	39.160	18.313	15.644	± 0.054	37.860	± 0.422
0.1205	0.511832	± 23	1919		-12.03	± 0.48	28.6	1.8	12.5	18.598	15.643	38.849	18.359	15.630	± 0.021	38.300	± 0.139
0.1123	0.511790	± 41	1834		-12.46	± 0.81	4.5	2.9	11.8	23.181	15.997	42.975	20.369	15.845	± 0.639	39.280	± 1.198
0.1129	0.511808	± 35	1818		-12.13	± 0.70	6.0	3.2	13.7	22.792	15.950	42.953	20.478	15.824	± 0.399	39.753	± 1.026
0.1124	0.511811	± 29	1806		-12.04	± 0.59	2.5	2.2	7.1	28.745	16.462	46.584	24.461	16.230	± 1.735	42.119	± 1.895
0.1101	0.511798	± 29	1789		-12.20	± 0.58	3.0	2.0	8.3	23.690	16.050	42.698	20.757	15.891	± 0.999	38.796	± 1.457
0.1118	0.511812	± 42	1796		-12.00	± 0.83	19.5	4.2	13.2	19.741	15.711	39.223	18.886	15.665	± 0.052	38.355	± 0.233
0.1286	0.512084	± 26	1678		-7.51	± 0.53	22.3	1.9	8.3	18.134	15.539	38.047	17.802	15.521	± 0.028	37.588	± 0.079
0.1199	0.512111	± 43	1495		-6.56	± 0.86	2.0	0.1	4.0	25.707	16.170	50.388	25.462	16.157	± 0.364	47.227	± 1.601
0.1716	0.512816	± 56	903		4.69	± 1.12	2.0	0.5	1.2	19.876	15.697	39.117	18.883	15.643	± 0.568	38.347	± 0.385
0.1680	0.512673	± 62	1244		2.08	± 1.23	8.0	0.5	2.2	18.522	15.593	38.405	18.281	15.580	± 0.074	38.063	± 0.045

⁷Calculated using depleted mantle parameters from Michard *et al.* (1985); values for samples with $^{147}\text{Sm}/^{144}\text{Nd} > 0.14$ are shown in italics. No value is quoted for the Cox Bight Granite, which has a higher $^{147}\text{Sm}/^{144}\text{Nd}$ than depleted mantle. ⁸Calculated after Liew & Hofmann (1988), assuming a crustal $^{147}\text{Sm}/^{144}\text{Nd} = 0.1100$. ⁹Modern CHUR (0.512638) from Goldstein *et al.* (1984). ¹⁰Metasedimentary member.

Table 3 Previously published combined Sr–Nd isotope data (eastern Tasmania).

Pluton/unit	Abb ¹	Sample(s)	Bath.	Type	⁸⁷ Sr/ ⁸⁶ Sr _(i)	2σ	ε _{Ndi}	2σ
St Marys Porphyrite	SM	mean	BTB	IMu	~0.7075	±0.0002	~−5.6	±0.2
Pyengana Granodiorite	PY	C40517	BTB	IMu	0.7061	±0.0010	−4.8	±0.4
Mt Pearson Granite	PE	C40524	BTB	I	0.7082	±0.0019	−6.2	±0.6
Lottah Granite	LO	various	BTB	SFh	~0.715	±0.002	~−3.1	±0.4
Mt Paris Granite	PA	C40548	BTB	SFh	0.7163	±0.0064	−2.0	±0.4
Hogans Road Diorite		84161005	BTB	IMu			5.8	±0.4
Coles Bay Granite	CO	C40533		IFh	0.7070	±0.0005	−2.5	±0.2
Boobyalla Granite	BY	R62583	EB	SFf	0.7119	±0.0008	−5.1	±0.2
Mathinna SGp (at 380Ma)		C43304A			0.7259		−11.2	±0.6
Mathinna SGp (at 380Ma)		C40572			0.7263		−10.4	±0.4

Compiled from Cocker (1977, 1982), Turner *et al.* (1986) and Sun & Higgins (1996).

¹Refers to Figures 14 and 15. Other abbreviations as in Table 2.

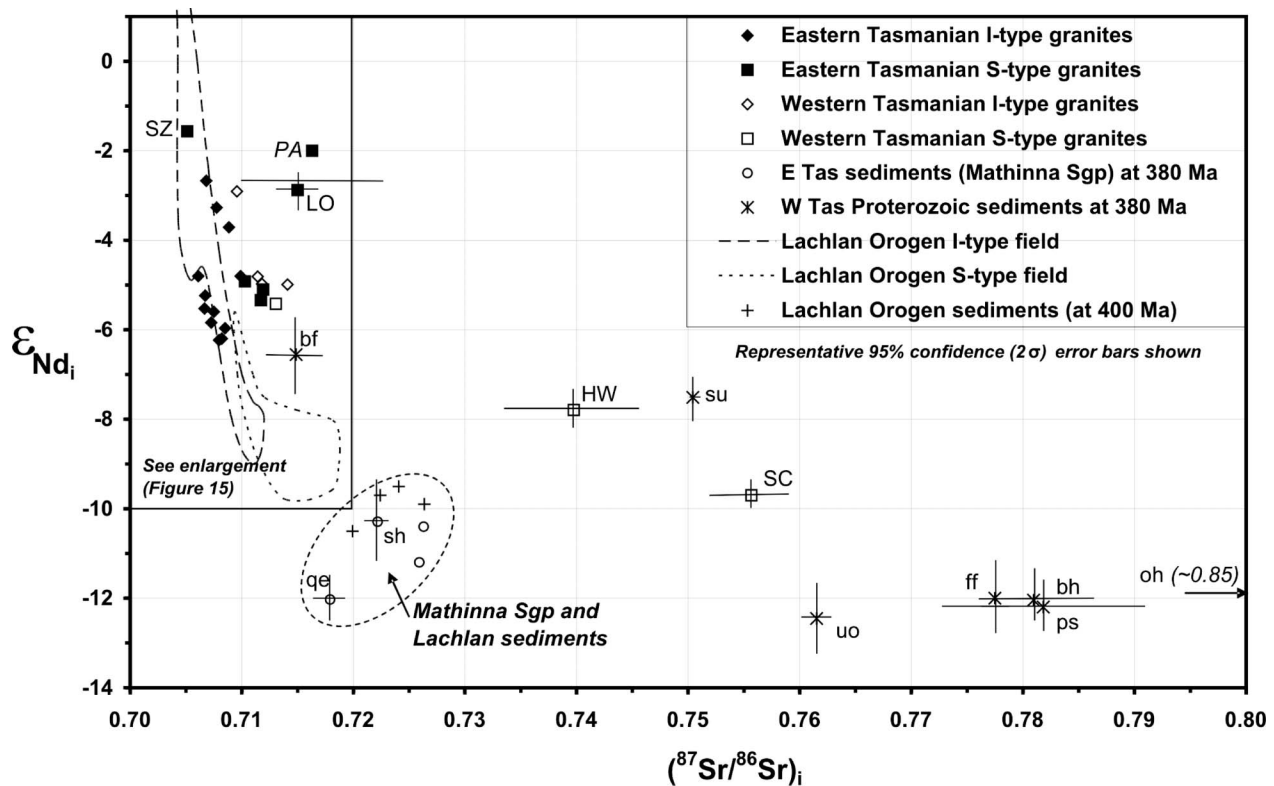


Figure 14 Plot of ϵ_{Nd} – $^{87}Sr/^{86}Sr$ for Tasmanian Devonian–Carboniferous granites, (corrected to their emplacement ages as discussed in text) and host sedimentary successions (corrected to 380 Ma). Labels for individual plutons and sedimentary successions from Table 2 (new data) and Table 3 (previously published data—italics). Field for Lachlan Orogen S-type granites from McCulloch & Chappell (1982); their I-type field is redrawn using additional data from Chappell *et al.* (1990), McCulloch & Woodhead (1993) and Keay *et al.* (1997).

isotopic compositions of I- and S-types. Also noteworthy is that strongly fractionated granites (which include most western granites, and the Henbury, Deep Glen Bay, Hippolyte Rocks and Lottah Granites from the east) also show a relatively large dispersal in both $^{207}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$, regardless of whether they are designated as I- or S-type.

Initial values of the thorogenic ratio $^{208}Pb/^{204}Pb$ are subject to relatively greater uncertainties (Figure 16b), but appear to be higher in the western Tasmanian granites (especially the S-types) than in those from the east.

EASTERN TASMANIAN I-TYPES

Scottsdale Batholith

The Scottsdale Batholith consists of eight I-type plutons, emplaced during a relatively short period (~5 million years) in a progressive, crudely nested sequence from relatively mafic granodiorite to alkali feldspar granite (Black *et al.* 2005; Figure 2). We have studied the Upper Blessington Granodiorite, Hogarth Road Granite and Tombstone Creek Granite which belong to the Russells Road Suite, together with the Diddleum Granodiorite from the slightly different Diddleum Suite. Our new

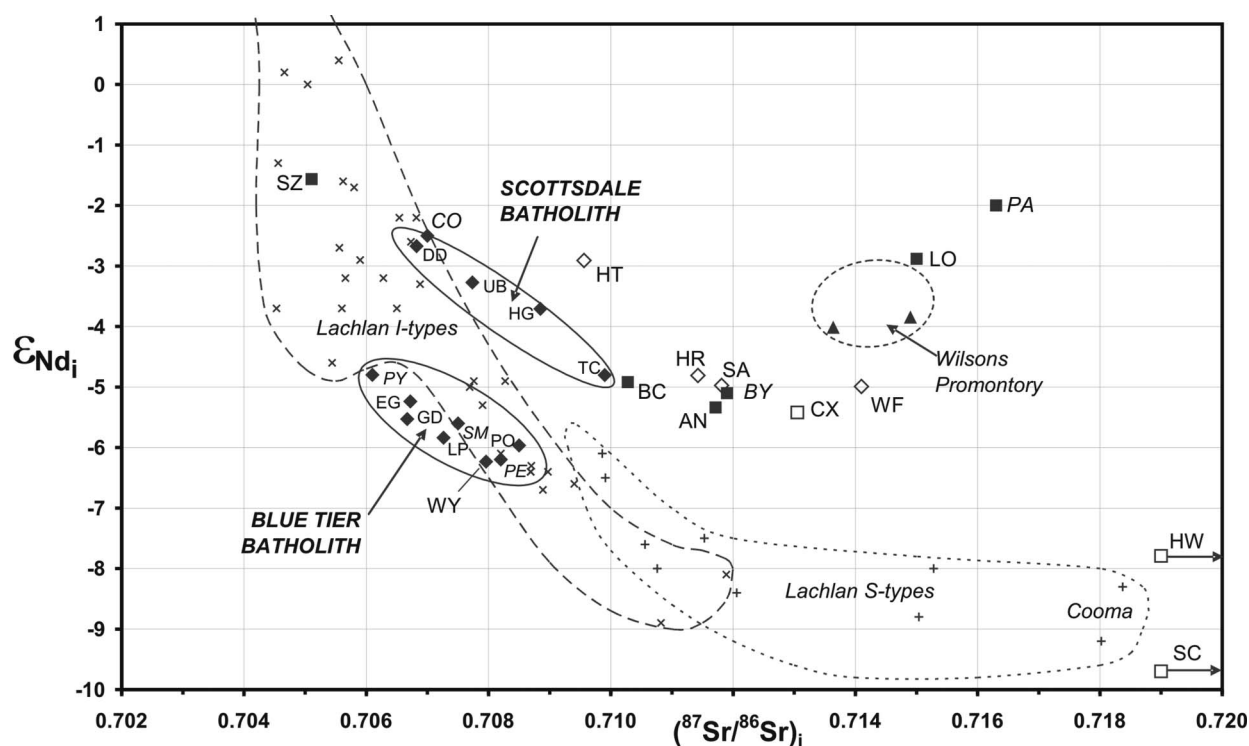


Figure 15 Enlargement of part of Figure 14 (note greatly expanded horizontal scale). Symbols as for Figure 14; data points for individual Lachlan I-types (diagonal crosses) and S-types (upright crosses) also shown. The field for Wilsons Promontory S-types (filled triangles: data from Elburg & Nicholls 1995; Elburg 1996) is drawn to encompass a leucogranite ($\epsilon_{\text{Nd}} - 3.03$) for which $^{87}\text{Sr}/^{86}\text{Sr}$ was not determined.

isotopic data produce a short but well-defined steeply negative alignment on the ϵ_{Nd} -initial $^{87}\text{Sr}/^{86}\text{Sr}$ diagram (Figure 15). Furthermore, ϵ_{Nd} progressively decreases, and initial $^{87}\text{Sr}/^{86}\text{Sr}$ increases, in successively younger plutons (Figures 18a, b). In the Russells Road Suite, these trends also correlate with the temporal progression to more felsic compositions (Figure 18c). All of these features strongly suggest that the plutons of the Scottsdale Batholith were derived from the same, or closely related, magmas. The trend towards less radiogenic Nd (lower ϵ_{Nd}) with time, however, obviously could not have occurred in a closed magma chamber. Also, the rate of increase in initial $^{87}\text{Sr}/^{86}\text{Sr}$ is too fast to have occurred in a closed system, implying an unrealistic $^{87}\text{Rb}/^{86}\text{Sr}$ of ~ 48 , compared with 0.9 to 7.9 in the actual samples. Instead, isotopic differences strongly suggest a small but progressively increasing input from a sedimentary component and/or a decreasing input from an isotopically primitive component or components.

Initial Pb isotope ratios from the Scottsdale Batholith tend to become less radiogenic, concurrent with the decrease in ϵ_{Nd} and increase in $^{87}\text{Sr}/^{86}\text{Sr}$, in successively younger and more felsic plutons (Figures 16, 17). The Pb isotope ratios for the felsic Tombstone Creek Granite are not far removed from those of the Blue Tier Batholith. The decrease in $^{206}\text{Pb}/^{204}\text{Pb}$ is consistent with an increasing input from the Mathinna Supergroup, but this cannot account for decrease in $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$, unless our two samples from the Mathinna Supergroup are unrepresentative of the succession.

Although the chemical trends in the Scottsdale Batholith could be interpreted in terms of restite unmixing, perhaps together with fractional crystallisation in the more felsic bodies, the isotopic data show that a simple closed-system model is inadequate. Rather, progressive dilution of an initially isotopically more primitive magma by contamination or assimilation/fractional crystallisation (AFC) seems probable.

Blue Tier Batholith

The larger Blue Tier Batholith was divided, using mainly petrographic criteria, into about twenty plutons by Groves *et al.* (1977) and Cocker (1977), and these have been largely substantiated by subsequent detailed mapping (e.g. McClenaghan *et al.* 1987). Early workers (Groves *et al.* 1977; McCarthy & Groves 1979) advocated an *in situ* progressive fractional crystallisation model, but this was shown to be untenable by subsequent field, geochronological and Sr isotope evidence (Cocker 1982; Higgins *et al.* 1985). These later authors favoured a model whereby the various granite types were generated by the partial melting of a range of protoliths. Although some of the felsic bodies have been difficult to classify, most are now regarded as I-types (e.g. McClenaghan 2006). In the Blue Tier Batholith, S-types appear largely restricted to a few strongly fractionated granites, such as the Lottah and Mt Paris Granites which are considered in a later section.

Our new data are from the Gardens and Long Point Granodiorites and the large, more felsic Poimena

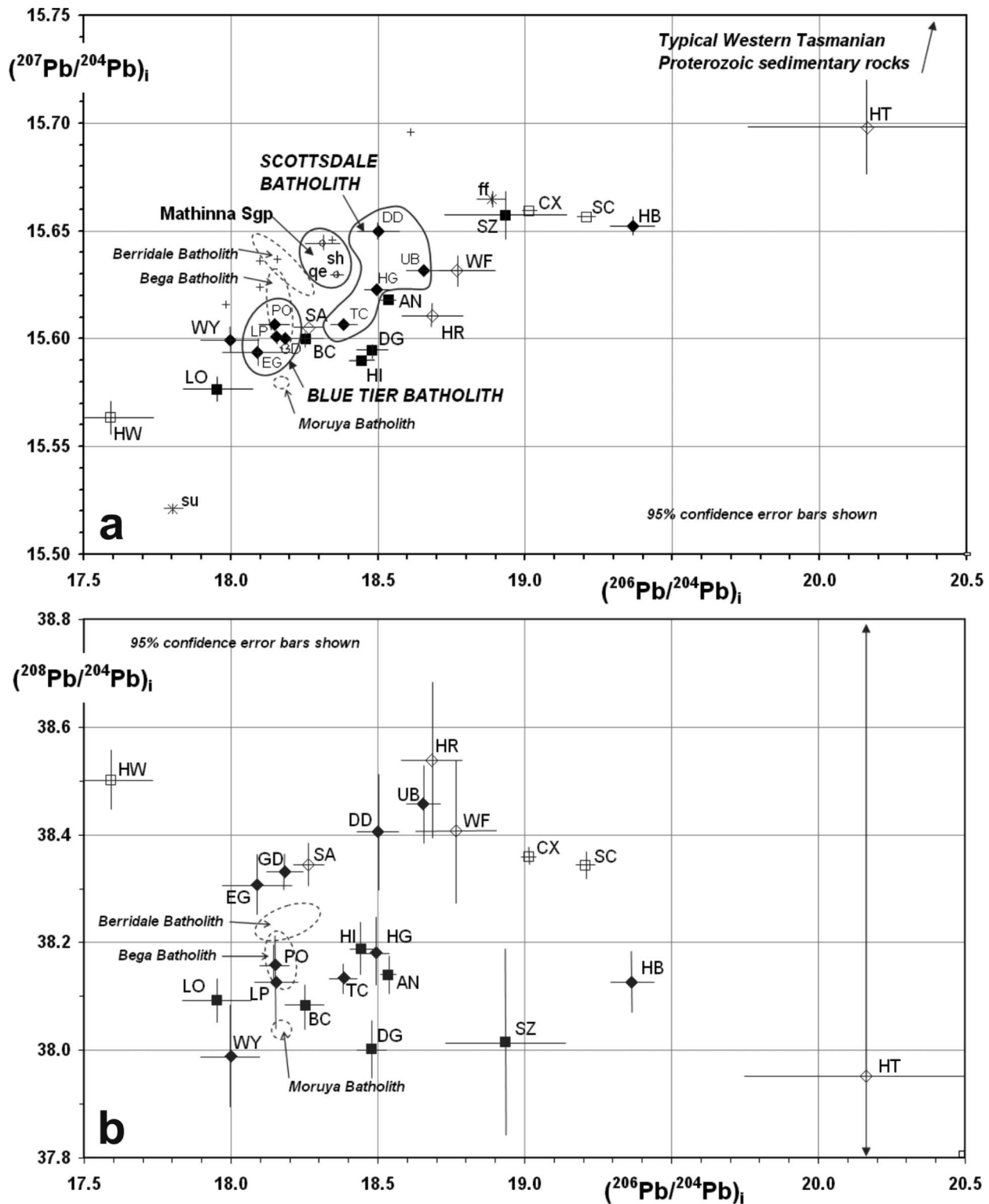


Figure 16 (a) Initial $^{207}\text{Pb}/^{204}\text{Pb}$ vs $^{206}\text{Pb}/^{204}\text{Pb}$ for Tasmanian granites and sedimentary successions. Fields for Moruya, Bega and Berridale Batholiths drawn from data of McCulloch & Woodhead (1993). (b) Initial $^{208}\text{Pb}/^{204}\text{Pb}$ vs $^{206}\text{Pb}/^{204}\text{Pb}$. Symbols as for Figure 14.

Granite. Also plotted on Figure 15 are previously published Sr–Nd data from the Pyengana Granodiorite, the St Marys Porphyry (a largely extrusive body chemically equivalent to granodiorite) and the felsic Mt Pearson Granite (Cocker 1982; Turner *et al.* 1986; Sun & Higgins 1996).

Taken together, these data define a tight field with similar $^{87}\text{Sr}/^{86}\text{Sr}$ but lower and less varied ϵ_{Nd} than the Scottsdale Batholith samples (Figures 15, 18), even though the bodies were emplaced over a longer time interval (~13 million years) and are considered to belong to different suites. Once again, there is a

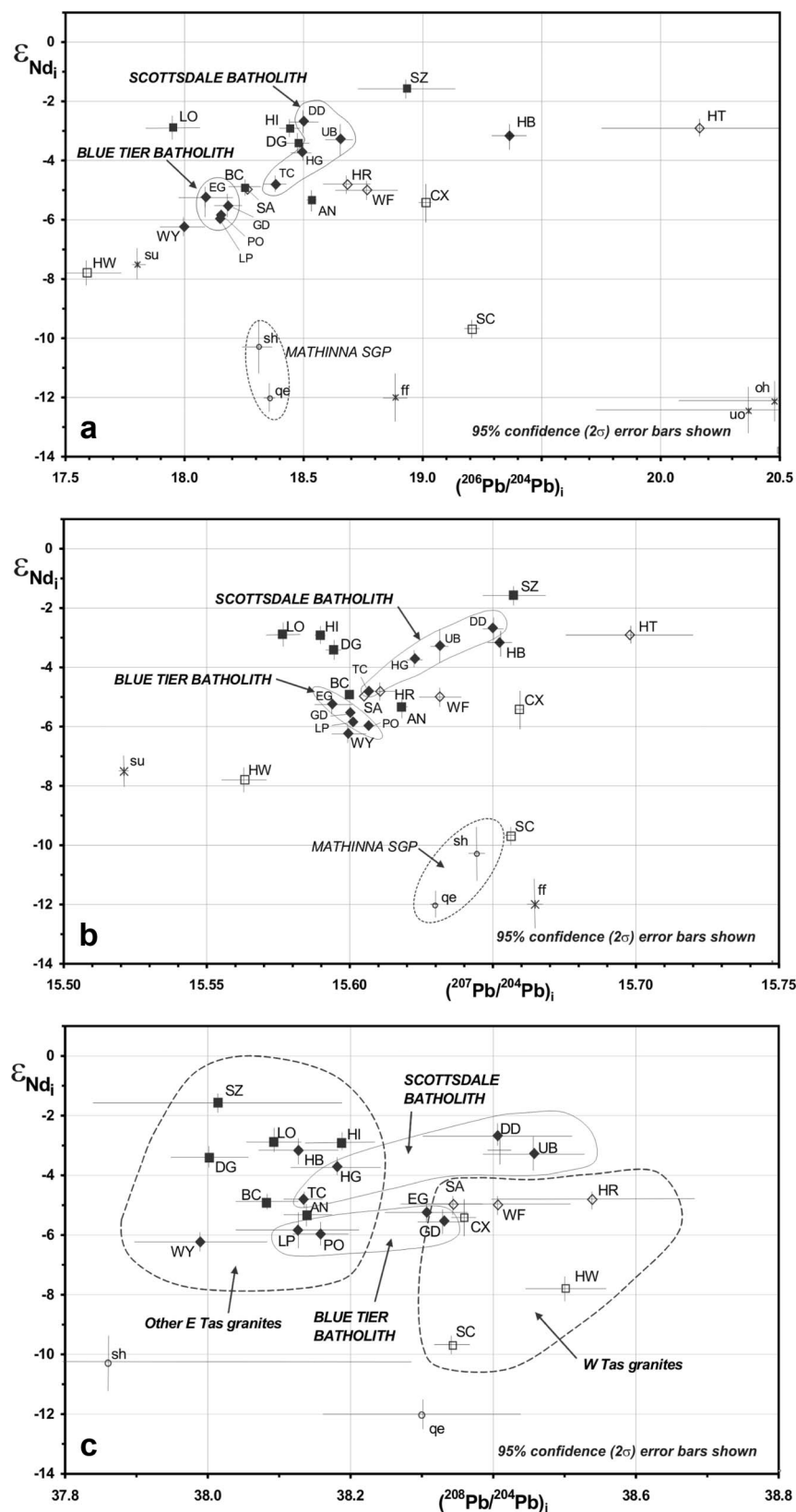


Figure 17 (a) Initial ϵ_{Nd} vs $^{206}\text{Pb}/^{204}\text{Pb}$ for Tasmanian granites and sedimentary successions. (b) Initial ϵ_{Nd} vs $^{207}\text{Pb}/^{204}\text{Pb}$ (c) Initial ϵ_{Nd} vs $^{208}\text{Pb}/^{204}\text{Pb}$. Symbols as for Figure 14.

negative correlation between $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd} . Although the three dated plutons may be derived from different source rocks, there is a temporal change to more felsic compositions, also correlating with small increases in $^{87}\text{Sr}/^{86}\text{Sr}$ and decreases in ϵ_{Nd} (Figure 18).

Initial Pb isotope ratios from the Blue Tier Batholith define a more restricted field at less radiogenic compositions, especially for $^{206}\text{Pb}/^{204}\text{Pb}$, than those from the Scottsdale Batholith (Figures 16, 17). As both batholiths intrude the same sedimentary succession (Mathinna Supergroup), these differences are more likely to have

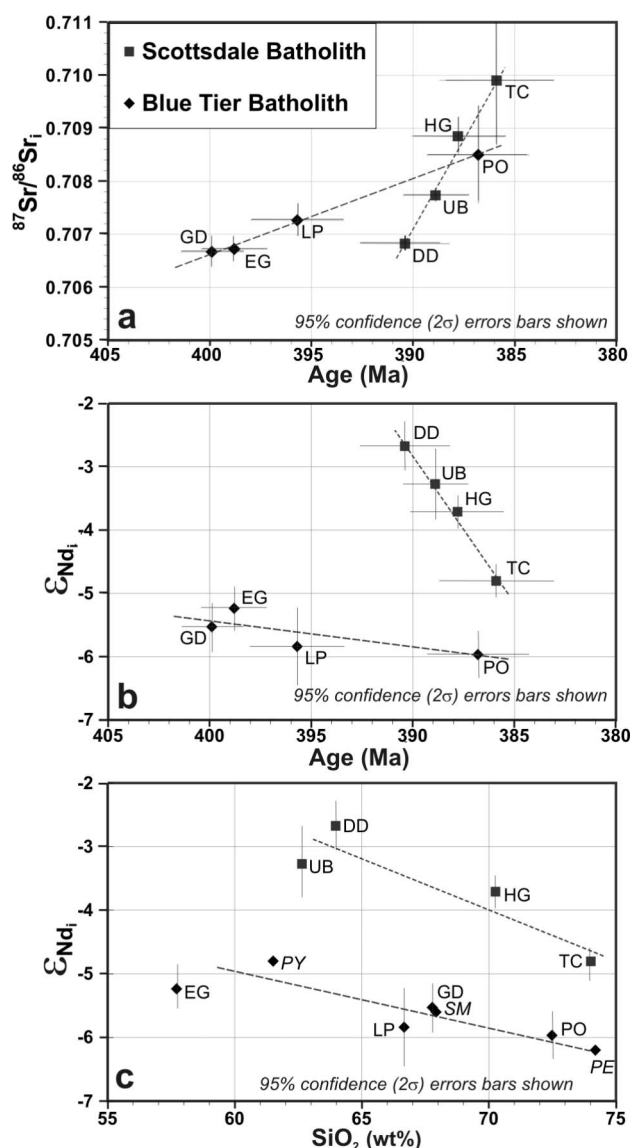


Figure 18 Progressive change in isotopic ratios with emplacement age (SHRIMP) and composition in the Scottsdale and Blue Tier Batholiths. (a) $^{87}\text{Sr}/^{86}\text{Sr}$ –age. (b) ϵ_{Nd} –age. (c) ϵ_{Nd} – SiO_2 .

originated in mantle or lower crustal igneous source components.

Other eastern Tasmanian I-types

The Wybalenna Granodiorite has isotopic ratios (Sr, Nd and Pb) similar to those of I-types from the Blue Tier Batholith, including the Gardens and Long Point Granodiorites, which were emplaced at about the same time (~400 Ma). The Henbury Granite is discussed below, together with other strongly fractionated bodies.

EASTERN TASMANIAN S-TYPES

The Ansons Bay and Bicheno Granites, together with the Boobyalla Granite (Cocker 1982; Sun & Higgins 1996),

have slightly higher initial Sr ratios (0.7103–0.7117) and similar ϵ_{Nd} (~–5) to the more felsic of the eastern I-types described above (Figure 15). In particular, they lie on a direct extension of the Scottsdale Batholith trend, and the first two, at least, also have comparable Pb isotope ratios (Figure 16a; Table 2). A direct relationship, whereby the S-types might represent further assimilation of a Tombstone Creek-type magma by sedimentary rocks of the Mathinna Supergroup, is precluded by major and trace elements (e.g. Na_2O , P_2O_5 , Sr; Table 1 and further data in GA and MRT databases) and the spatial separation of the bodies (Figure 2). The ϵ_{Nd} values (~–5) are, however, well above those of Lachlan Orogen S-type granites (McCulloch & Chappell 1982).

It is important to note, however, that the Lachlan S-type field is derived from a limited number of unfractionated, mostly relatively mafic (i.e. <70% SiO_2) bodies from the Kosciusko and Berridale Batholiths (initial Sr ratios ~0.710–0.715), together with the anatectic Cooma Granodiorite (initial ratio ~0.718). These are generally dissimilar to the felsic and mostly fractionated Tasmanian S-types. Elburg & Nicholls (1995) and Elburg (1996) have reported ϵ_{Nd} values of –4.0 and –3.0 from felsic S-types from the Wilsons Promontory Batholith, which have been considered to be a northern extension of the eastern Tasmanian granites (e.g., Chappell *et al.* 1988).

The similarity of the Pb isotopic characteristics of these eastern Tasmanian S-type granites with the I-types does not preclude a larger sedimentary source component, as the Mathinna Supergroup also has comparable ratios and elemental concentrations of Pb (~15–30 ppm vs 14–44 ppm in the granites).

Our sample from the Strzelecki Granite on Flinders Island has high ϵ_{Nd} (–1.6) and low initial $^{87}\text{Sr}/^{86}\text{Sr}$ (~0.7051), particularly for an S-type granite. It is, however, from a darker-coloured, finer-grained, less felsic (~71.1% SiO_2) phase which, at outcrop scale, appears to have mingled with a more felsic (~76% SiO_2), lighter-coloured and coarser-grained phase, more typical of the pluton as a whole. Both the major element chemistry and isotopic ratios of the sample are broadly consistent with an 80:20 mixture of the more felsic host granite and a mantle-derived basaltic melt. Because the former has low absolute concentrations of both Sr and Nd (Table 1, analysis SZ(f)), the isotopic ratios are likely to be dominated by the mantle-derived component, whereas the bulk composition is granitic. This interpretation is supported by the lower initial ϵ_{Nd} of –4.8 of the main, more felsic granite (Appendix 1; provided by D. C. Champion, pers. comm. 2009). Thus, we consider that this sample, while interesting as a local example of magma mingling, is unrepresentative at a regional scale.

EASTERN TASMANIAN STRONGLY FRACTIONATED GRANITES

The S-type Hippolyte Rocks, Deep Glen Bay and Lottah Granites, and the I-type Henbury Granite, are all very felsic, strongly fractionated, alkali-feldspar granites, enriched in Rb and depleted in Sr (Figure 4a). Since age corrections are approximately proportional to the product of parent/daughter ratio and the age, these

features result in large uncertainties in initial $^{87}\text{Sr}/^{86}\text{Sr}$. If XRF data for Rb and Sr are used, three of these bodies yield unrealistic initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, below depleted mantle values (~ 0.7026 ; e.g. McCulloch & Black 1984; Taylor & McLennan 1985), with very large uncertainties (Table 2). In an effort to improve precision and accuracy, we have also analysed Rb and Sr in these granites by isotope dilution at The University of Melbourne (R. Maas, analyst). However, these data yield, for all four granites, initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (at $> 95\%$ confidence levels) wholly below depleted mantle values (Table 2). This suggests a systematic error.

Black *et al.* (2005) noted that previously published K–Ar and Rb–Sr ages from Tasmanian granites are typically ~ 10 million years less than U–Pb SHRIMP zircon ages from the same intrusion. Since the latter are considered to more accurately indicate the time of emplacement and crystallisation, they attributed the discrepancy to residual heat associated with cooling and, by implication, the lower closure temperatures of the K–Ar and Rb–Sr systems. As even the Rb–Sr whole rock ages tend to be younger, a protracted period of open system behaviour at pluton scale for ~ 10 million years (not just internal redistribution of Rb and Sr between minerals) is implied.

These granites display petrographic evidence for hydrothermal alteration, such as ubiquitous partial sericitisation of feldspar and common alteration of biotite to muscovite and/or chlorite. For some time after their crystallisation, fluids may have circulated within and around the plutons, redistributing Rb and Sr and, in particular, leading to loss of radiogenic Sr. Based on the isotopic data, and on the assumption that there has been no significant weathering of these samples, some of the alteration must have continued until at least 367 Ma (Lottah Granite), 350 Ma (Henbury Granite), and 348 Ma (Deep Glen Bay Granite). Thus, correcting single-sample Sr isotope measurements to the U–Pb SHRIMP ages may lead to a systematic underestimation of initial $^{87}\text{Sr}/^{86}\text{Sr}$, particularly for strongly fractionated granites with high Rb/Sr.

One sample (84162108) of Lottah Granite previously studied by Mackenzie *et al.* (1988) came from the same locality as our sample 99220061. It yielded an excellent five-point mineral isochron with an age of 366 ± 3 Ma and an initial ratio of 0.715 ± 0.002 . The adoption of that age, rather than the zircon age of 378 Ma, changes the initial $^{87}\text{Sr}/^{86}\text{Sr}$ of our sample from 0.694 to 0.710 ± 0.015 , in much closer agreement with the value of Mackenzie *et al.* (1988). A more precise estimate of the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of sample 84162108 can be calculated from the plagioclase analysis, because of its relatively low $^{87}\text{Rb}/^{86}\text{Sr}$ (21.03). This yields 0.7143 at 366 Ma (or 0.7107 at the SHRIMP age of 378 Ma), which is probably the most robust estimate available for the initial $^{87}\text{Sr}/^{86}\text{Sr}$ of the Lottah Granite. However, we are unable to quote reliable initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for the other three bodies.

It is worth noting that Cocker (1977, 1982) obtained similar initial Sr ratios of 0.708–0.716 from the fractionated Mt Paris Granite, which closely resembles the nearby Lottah Granite. Probably the most reliable determination is a four-point mineral isochron which yielded an initial ratio of 0.7163 ± 0.0064 at 362 ± 7 Ma.

A striking feature of strongly fractionated eastern Tasmanian granites is their high and relatively uniform ϵ_{Nd} . Our value from the Lottah Granite ($\epsilon_{\text{Nd}} -2.9$) compares well with previous data from this body (-3.4 to -2.1 excluding a sample of doubtful assignment) and the Mt Paris Granite (-4.0 to -2.0) (Sun & Higgins 1996). The Deep Glen Bay Granite ($\epsilon_{\text{Nd}} -3.4$) and the Hippolyte Rocks Granite (-2.9) are petrographically and chemically very similar S-types of the same age (374 Ma), only 18 km apart, and may be part of the same pluton. The other fractionated S-types analysed by Sun & Higgins (1996), both from the Babel Island Suite, were the Mt William Granite ($\epsilon_{\text{Nd}} -5.8$ to -3.3) and the Babel Island Granite (-5.7 to -3.0). The only strongly fractionated eastern Tasmanian I-type that we have studied is the Henbury Granite ($\epsilon_{\text{Nd}} -3.2$), but Sun & Higgins (1996) report -2.5 for the Coles Bay Granite.

These ϵ_{Nd} values are comparable to those from the Scottsdale Batholith ($\epsilon_{\text{Nd}} -4.8$ to -2.7), especially the more mafic Diddleum and Upper Blessington Granodiorites. The very felsic Coles Bay Granite also has a very similar initial $^{87}\text{Sr}/^{86}\text{Sr}$ to the otherwise very different Diddleum Granodiorite (Tables 2, 3; Figure 15). Derivation of fractionated I-types from a parent magma similar to the Diddleum Granodiorite might be postulated but, as described above, the Scottsdale Batholith displays a trend to lower ϵ_{Nd} with successively younger, more felsic plutons.

The only less fractionated S-type with high ϵ_{Nd} that could represent a parent of the strongly fractionated S-types is the Strzelecki Granite, but this rock shows clear evidence of contamination by mafic magma. Mackenzie *et al.* (1988) and Sun & Higgins (1996) both noted the high ϵ_{Nd} of the Lottah Granite, and tentatively suggested a contribution from an underplated mantle-derived magma, also providing a heat source. However, these plutons, unlike the Strzelecki Granite, show little field, major or trace element evidence for magma mingling.

The Pb isotope data from these fractionated granites lie close to the fields defined by other eastern Tasmanian granites, but are slightly more dispersed (Figures 16, 17). The Henbury pluton has the highest $^{206}\text{Pb}/^{204}\text{Pb}$, and the Lottah pluton has the lowest $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$. This might be attributed to slight mobility of Pb in hydrothermal fluids, but the consistency of values (particularly $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$) for the similar Deep Glen Bay and Hippolyte Rocks suggests that the calculated initial ratios are reliable.

WESTERN TASMANIAN I-TYPES

Our initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for the Housetop, Wombat Flat and Sandblow Granites have low estimated uncertainties and are in good agreement with previous data from the former two bodies (Sawka *et al.* 1990). On the Sr–Nd plot (Figure 15), they plot above and/or to the right of the both the Lachlan Orogen I-type array and the Scottsdale Batholith. The Housetop Granite is isotopically the most primitive, whereas the unfractionated and less felsic Sandblow and Wombat Flat Granites have higher initial $^{87}\text{Sr}/^{86}\text{Sr}$ and slightly lower ϵ_{Nd} (~ -5.0).

Although our initial $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd} values for the strongly fractionated Heemskirk Red Granite are similar to the other western I-types, the former (0.7114 ± 0.0074), has a much larger uncertainty, due to higher Rb/Sr. This value is apparently lower than, but actually within error of, that of 0.7190 ± 0.0051 estimated by Brooks & Compston (1965) in a detailed Rb–Sr study of the Heemskirk Batholith. Brooks (1966), reworking their data, concluded that both the Heemskirk Red and White Granites were emplaced with a range of initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, and subsequently the isotopic systems were disturbed by migration of radiogenic Sr from potash feldspar to plagioclase.

The Pb isotope signatures of the Sandblow, Wombat Flat and Heemskirk Red granites are similar to those of the mostly more mafic I-types of eastern Tasmania, particularly the Scottsdale Batholith (Figure 16a, b). The Housetop Granite, however, has the most uranogenic Pb ratios (highest $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$) of all the granites, even allowing for the relatively large estimated uncertainties. Its initial $^{208}\text{Pb}/^{204}\text{Pb}$ ratio is far too imprecise (due to high Th/Pb) to be useful.

WESTERN TASMANIAN S-TYPES

The three western S-type granites are all moderately to strongly fractionated bodies with considerable uncertainty in their initial $^{87}\text{Sr}/^{86}\text{Sr}$. Our values for the Heemskirk White Granite (0.7397 ± 0.0060) and the Sandy Cape Granite (0.7557 ± 0.0035), however, agree very well with the isochron determinations of Brooks (1966) and Sawka *et al.* (1990) respectively, although the latter yield slightly younger ages (by 4–17 million years).

On a Sr–Nd plot (Figures 14, 15), all three lie outside the field of Lachlan Orogen S-types. The Cox Bight Granite lies close to some of the western Tasmanian I-types, whereas both the more strongly peraluminous Sandy Cape and Heemskirk White Granites have extremely high initial $^{87}\text{Sr}/^{86}\text{Sr}$ (~ 0.756), far higher than the most radiogenic Lachlan Orogen granites (~ 0.718). An obvious interpretation is that this reflects source components isotopically similar to their host sedimentary successions, which are much older and have more radiogenic Sr than the Mathinna Supergroup (Figure 14; Table 2). The ϵ_{Nd} of the western S-types is however higher than that of most of these western sedimentary successions, which are mostly similar to the Mathinna Supergroup (~ -12).

In contrast, the Pb isotopic compositions of the western S-types are more scattered than, but overall not consistently distinguishable from, the I-types (Figure 16, 17). The Cox Bight and Sandy Cape Granites have relatively uranogenic Pb, whereas the Heemskirk White Granite has the least of all the granites. This does not necessarily exclude the possibility of a sedimentary source component. For example, Table 2 shows that the two most fertile of the western Tasmanian sedimentary samples (from the Fraser and Surprise Bay Formations, both from King Island) have similar Pb isotopic compositions to the granites and, if they are representative of the assimilate, would not have much affected those ratios. Although the other, less fertile, samples from the Tasmanian mainland have

more uranogenic Pb compositions at the time of granite generation, they have markedly lower Pb contents (≤ 4.5 ppm).

If Pb has been lost from potash feldspars during alteration or weathering of the Heemskirk White Granite, it would result in higher U/Pb and over-correction of $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$, an effect that would be exacerbated if the lost Pb were relatively radiogenic. These processes possibly account for the anomalously low initial ratios.

PETROGENETIC IMPLICATIONS

We consider that restite unmixing cannot, at least in itself, account for the compositional variation observed within Tasmanian granites, notably the isotopic trends within the Russells Road Suite of the Scottsdale Batholith, and is inconsistent with the generally low proportion of inherited zircon. Also, the similarity of the inherited zircon spectra of I- and S-type granites and their country rocks, in eastern and western Tasmania respectively, are difficult to reconcile with the concept that ‘granites image their source rocks.’ Furthermore, the isotopic characteristics of strongly fractionated granites, particularly the high ϵ_{Nd} of fractionated S-types, differ from unfractionated granites, and they cannot have been derived from them (or similar magmas) merely by closed system crystal fractionation.

Rather, we favour a model similar to that suggested by Collins (1996, 1998) and Keay *et al.* (1997), in which much of the compositional and isotopic variation is generated by mixing, in varied proportions, between one or more mafic components, directly or indirectly derived from the mantle, and a felsic component derived from anatexis of sedimentary rocks.

MANTLE COMPONENT

At least one component was mafic, mantle-derived magma, which also provided a heat source for further melting and assimilation. The composition and high temperature origin of this component suggest that it contributed no inherited zircon to the granites. There is some direct field evidence for the presence of such a component, particularly in eastern Tasmania, in dolerite dykes associated with the granites, rare diorites (McClenaghan 1984) and mesoscopic features within the granites, such as mafic schlieren, mafic enclaves and compositional variation.

Only limited isotopic data are available from these rocks. No rocks as isotopically primitive as the microdiorite at Bingie Bingie Point ($^{87}\text{Sr}/^{86}\text{Sr} \geq 0.70322$; $\epsilon_{\text{Nd}} \leq +8.0$), which was used as a hypothetical component by Keay *et al.* (1997), are known from eastern Tasmania. Sun & Higgins (1996) report initial ϵ_{Nd} of $+4.7$ for the Hogans Road Diorite, a series of small bodies intruding the Haleys New Country pluton. These are comparable to values from the tonalites of the Moruya Batholith ($+2.7$ to $+4.3$; Chappell *et al.* 1990), considered by Keay *et al.* (1997) to be themselves the mixing products of mantle-derived magmas and dacitic partial

melts of lower crust. Two of the Devonian dolerites from northeastern Tasmania gave values of -2.3 and -1.8 (Sun & Higgins 1996), suggesting some upper crustal influence. No Sr isotope data are available for these mafic rocks, but the general trend of both the Lachlan array and the eastern Tasmanian I-types granites, and data for the Strzelecki Granite (Figures 14, 15), indicate that an initial $^{87}\text{Sr}/^{86}\text{Sr}$ of ~ 0.704 is likely for this component.

The western Tasmanian granites are consistently more felsic, but their ϵ_{Nd} (up to -2.9) is generally more positive than that of their host rocks and also requires at least one isotopically primitive source component. Local field evidence within the granites themselves for a mantle contribution is sparse, but mafic enclaves occur in the I-type Sandblow Granite (Wesolowski *et al.* 1985) and in the S-type Three Hummock Island Granite (Everard *et al.* 1997).

LOWER CRUSTAL COMPONENT

A mafic to intermediate, but isotopically relatively primitive component derived from partial melting of Neoproterozoic to Cambrian lower crust may be required by major element constraints, as suggested by Collins (1996, 1998), but neither the inherited zircon nor the isotopic data provide any firm constraints on it. Subtle apparent differences between the zircon age spectra of the Eastern Tasmanian granites and that of the Mathinna Supergroup could be interpreted as indicating other older (1100 Ma and 800 Ma) sources, but further data from the Mathinna Supergroup are needed to test this hypothesis. Any mafic source would be unlikely to contribute significant zircon.

The isotopic characteristics of this component are probably only slightly more evolved than the mantle-derived component; for instance, the Cambrian Heathcote tholeiite (Nelson *et al.* 1984) used in the model of Keay *et al.* (1997) has $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7054 and ϵ_{Nd} of $+4.6$. Rocks of appropriate age are not exposed in eastern Tasmania, but crop out in western Tasmania; for example, amphibolites in the Forth and Arthur Metamorphic Complexes have values (at 380 Ma) of 0.7066 to 0.7084 and $+4.7$ and $+2.1$ (Table 2). These rocks are more mafic than envisaged for this component, but partial melts of them, which would have the same isotopic compositions, may be appropriate.

The chemical and isotopic differences (Figure 15) between the Scottsdale and Blue Tier Batholiths and, by extension, between granite suites elsewhere within the Lachlan Orogen, may reflect regional-scale differences in the isotopically primitive components. These could be discrete mantle melts of differing composition, lower crustal components of different composition, or different mixing ratios between the two. For example, it might be that the mantle component for the Scottsdale Batholith had a slightly higher $^{87}\text{Sr}/^{86}\text{Sr}$, or that it was derived from a higher proportion of a lower crustal component (with higher $^{87}\text{Sr}/^{86}\text{Sr}$) relative to the mantle component, than the Blue Tier Batholith. Differences in source component composition, or relative proportion, might also explain the higher

$\text{Na}_2\text{O}/\text{K}_2\text{O}$ of the Scottsdale Batholith. As discussed below, however, different mixing processes (simple assimilation *vs* assimilation-fractional crystallisation) will also affect the isotopic arrays.

SEDIMENTARY COMPONENT- EASTERN TASMANIA

The exposed sedimentary country rocks of eastern and western Tasmania provide, or at least may be representative of, another source or sources for the granites in the respective regions. In eastern Tasmania, the inherited zircon age spectra of both the I- and S-type granites are consistent with input from the Mathinna Supergroup, or its protolith(s). Isotopic trends, particularly in the Scottsdale Batholith and the Blue Tier Batholith (apart from the Lottah Granite) are also consistent with progressive contamination of more mafic magmas by the Mathinna Supergroup (Figures 14, 15, 18).

Mineralogically, the Mathinna Supergroup is dominated by quartz and muscovite. Available geochemical data (~ 85 analyses from MRT and Geoscience Australia databases) show the succession is not particularly fertile: Na_2O is usually less than 2%, and often less than 1%; CaO less than 1.5% and mostly less than 0.5%. Nevertheless, simple calculations show that it is capable of producing large volumes of felsic magma. For instance, at 400 MPa (~ 15 km depth), its mean composition (Table 4) is capable of yielding up to 14.1% of haplogranite partial melt ($\text{Q}_{30}\text{Ab}_{47}\text{Or}_{23}$) under water-saturated conditions at 655°C , or 20.8% melt ($\text{Q}_{34}\text{Ab}_{32}\text{Or}_{34}$) under anhydrous conditions at 1000°C (Steiner *et al.* 1975). In both cases, the limiting parameter is Na_2O (that is, the exhaustion of albite in the residue). The actual melt will, of course, be peraluminous and contain small amounts of CaO and other components.

At 380 Ma, the $^{87}\text{Sr}/^{86}\text{Sr}$ of the Mathinna Supergroup lay between 0.718 and 0.726, and ϵ_{Nd} was between -12.0 and -10.3 (Tables 2, 3). The mean elemental concentrations of Sr and Nd, which are important parameters in mixing and assimilation models, are 59.6 ppm and 29.7 ppm respectively (available data, $n=44$).

Due to the relative isotopic homogeneity of the Mathinna Supergroup, and its likely chemical homogeneity at scales (1–10 km) relevant to granite genesis, we do not consider that isotopic or chemical differences between specific plutons, batholiths or suites in eastern Tasmania are inherited from the sedimentary component.

SEDIMENTARY COMPONENTS-WESTERN TASMANIA

The exposed sedimentary successions in western Tasmania are much more diverse in age, sedimentology, and chemical and isotopic composition than the Mathinna Supergroup, and are not necessarily representative of those at depth at the time of granite generation. Therefore, it is not possible to specify a particular isotopic composition for the sedimentary source component in western Tasmanian granites; very likely, several were involved. Broadly, the higher $^{87}\text{Sr}/^{86}\text{Sr}$ of these granites, particularly the S-types, correlates with

the higher $^{87}\text{Sr}/^{86}\text{Sr}$ (at ~ 380 Ma) of the western sedimentary rocks.

Most of our sedimentary samples from western Tasmania are chemically mature, with negligible Na_2O and CaO (Table 1), and are incapable of partial melting at realistic temperatures to produce a granite melt. The samples, however, were chosen primarily for detrital zircon studies, and are biased towards relatively infertile lithologies such as orthoquartzites. Sedimentary rocks more likely to be representative of potential source rocks for granites are also exposed in western Tasmania, even in the Early Neoproterozoic siliciclastic successions. For instance, the average composition of pelitic units within the Rocky Cape Group (Table 4) is comparable to that of the Mathinna Supergroup, and at 400 MPa (for instance), could also have generated about 15% and 22% of haplogranite melt under water-saturated and anhydrous conditions, respectively.

In the west, however, an additional component of mid- to deep-crustal origin, not represented in the exposed sedimentary successions, is clearly identified by a high proportion of ~ 1600 Ma zircon. It might be the equivalent, albeit a considerably older one, to the deep crustal igneous source component inferred by Keay *et al.* (1997) and Collins (1998) for the Lachlan Orogen granites. In western Tasmania, however, this source component is likely to be significantly more felsic than mafic rocks because, in contrast to the Lachlan Orogen, it has yielded a significant proportion of the inherited zircon budget. It is also feasible that the ~ 1600 Ma zircon-bearing, deep crustal source component is of sedimentary derivation.

Table 4 Mean major element composition of representative sedimentary successions.

	Mathinna SGp	Rocky Cape Gp
	<i>n</i> = 84	<i>n</i> = 35
SiO_2	76.52	72.12
TiO_2	0.60	0.66
Al_2O_3	12.77	16.32
Fe_2O_3	1.75	1.34
FeO	2.28	2.94
MnO	0.05	0.04
MgO	1.50	1.42
CaO	0.31	0.10
Na_2O	0.78	0.83
K_2O	3.32	4.18
P_2O_5	0.12	0.06
Igneous (CIPW) Norm		
<i>Q</i>	55.87	47.59
<i>C</i>	7.62	10.37
<i>Or</i>	19.60	24.69
<i>Ab</i>	6.64	7.06
<i>An</i>	0.74	0.14
<i>Hy</i>	5.55	6.83
<i>Mt</i>	2.54	1.94
<i>Il</i>	1.15	1.25
<i>Ap</i>	0.29	0.13

Mathinna Supergroup data from MRT and Geoscience Australia databases; Rocky Cape Group pelite data from Everard (2005) and Everard *et al.* (2007).

Numerical modelling of isotopic trends

EASTERN TASMANIA

The three component mixing model of Collins (1996, 1998) and Keay *et al.* (1997) can be adapted and extended to encompass the unusual isotopic and other features of eastern Tasmanian granites (Figure 19).

Keay *et al.* (1997) used the equations of De Paolo & Wasserburg (1979) to model Sr–Nd isotopic trends in the Lachlan Orogen by the simple assimilation of Ordovician turbidites by mafic melts. First, simple mixing between the mantle-derived mafic component and the lower crustal component (or a partial melt thereof) may result in products comparable to the isotopically most primitive granitoids in the Lachlan Orogen, or eastern Tasmania. For instance, a 20:80 ratio of mantle and lower crustal components (with elemental and isotopic Sr and Nd compositions as in Figure 19) produces a mix with $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7048 and ϵ_{Nd} of +4.3, similar to the Hogans Road Diorite (ϵ_{Nd} +4.7, Sun & Higgins 1996) or the Tuross Head Tonalite (0.7041, +3.8; Keay *et al.* 1997). Then, this primitive end-member is itself mixed with the average Mathinna Supergroup composition, either by direct assimilation or by mixing with a partial melt. Unfractionated I-type granites, both in Tasmania and in the Lachlan Orogen, and the relatively mafic Lachlan Orogen S-types, generally lie within the compositional triangle, in $^{87}\text{Sr}/^{86}\text{Sr}$ – ϵ_{Nd} space, between realistic estimates of the three components (Figure 19).

The effect of fractional crystallisation during assimilation (that is, AFC processes) on isotopic ratios can be calculated using the equations of De Paolo (1981). In addition to the composition of the components, a key parameter is the mass ratio (*r*) of assimilation to fractional crystallisation rates. This mechanism was considered by Keay *et al.* (1997), but rejected as being of no more than minor importance, because the calculated trends diverge to high ϵ_{Nd} and high $^{87}\text{Sr}/^{86}\text{Sr}$ values that do not fit Lachlan Orogen data (Figure 19). Tasmanian S-type and/or fractionated granites, however, lie in this region, which is not accessible by simple mixing. Furthermore, even though isotopic data from the Blue Tier and Scottsdale Batholith I-types plot within the composition triangle, each batholith defines a flatter array than the calculated trends for simple assimilation (Figure 19), and more closely fits trends for AFC with appropriate *r* values.

STRONGLY FRACTIONATED GRANITES

Although their major element compositions lie close to the haplogranite minimum, strongly fractionated I- and S-type granites are most easily distinguished by the contrasting behaviour of P_2O_5 , REE, Y and Th. These elements are controlled by accessory phases such as apatite, monazite and xenotime, which in turn are affected by the alumina saturation state of the melt (e.g. Sawka *et al.* 1990; Chappell & White 1992, 1998; Chappell 1999). The implication, according to proponents of the restite model, is that fractionated granites also retain an imprint, albeit a more subtle one, derived from their source rocks.

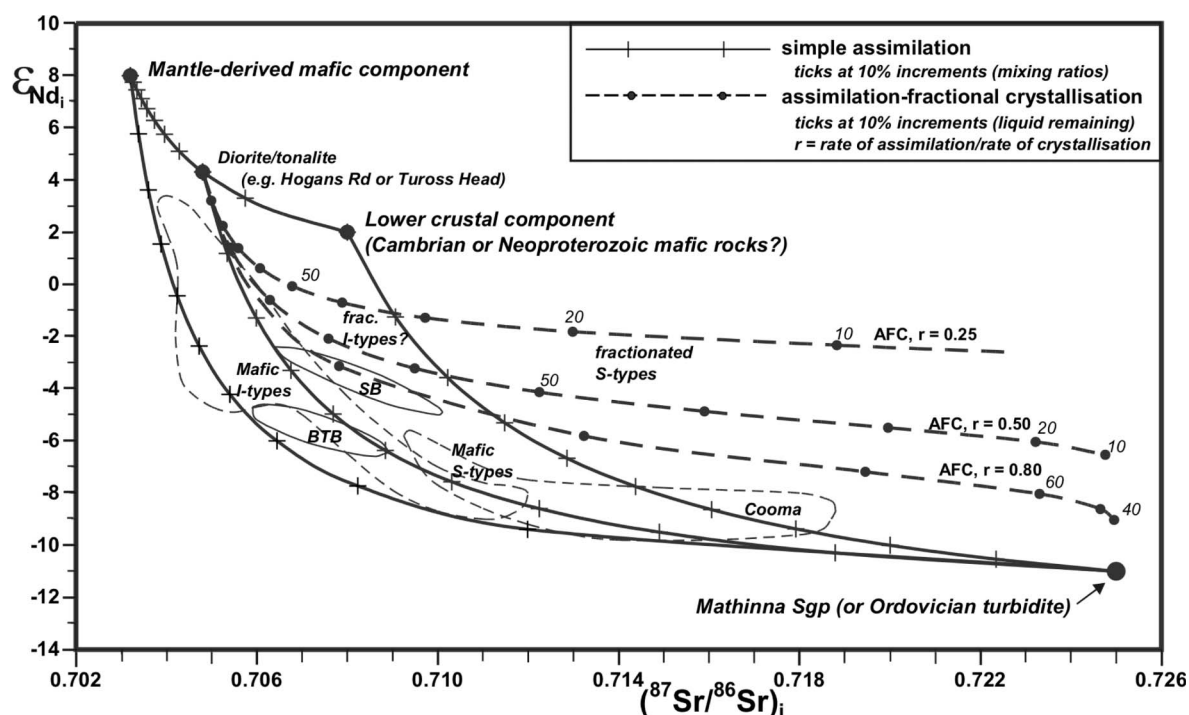


Figure 19 Diagram showing $\epsilon_{\text{Nd}}-^{87}\text{Sr}/^{86}\text{Sr}$ fields for eastern Tasmanian and Lachlan Orogen granites, and a multi-component mixing model for their origin (partly after Keay *et al.* 1997). The components are mantle-derived mafic magma (800 ppm Sr, 25 ppm Nd, $^{87}\text{Sr}/^{86}\text{Sr}$ 0.7032, ϵ_{Nd} +8), a Late Neoproterozoic mafic crustal source (100 ppm Sr, 10 ppm Nd, $^{87}\text{Sr}/^{86}\text{Sr}$ 0.708, ϵ_{Nd} +2) and mean Mathinna Supergroup turbidite (60 ppm Sr, 30 ppm Nd, $^{87}\text{Sr}/^{86}\text{Sr}$ 0.725, ϵ_{Nd} -11). Calculated trends are shown for the mixing of the mafic and the sedimentary components by both simple assimilation (equations of De Paolo & Wasserburg 1979), and assimilation-fractional crystallisation (equations of De Paolo 1981). Bulk distribution coefficients (D) of 2 and 0.1 are used for Sr and Nd respectively. BTB: Blue Tier Batholith; SB: Scottsdale Batholith.

Combined assimilation-fractional crystallisation (AFC), however, can account for the isotopic characteristics (such as high ϵ_{Nd}) of strongly fractionated granites. In this model, they are generated directly by extended assimilation of lower or middle crust by fractionating, initially mafic magmas originating in the mantle, and not derived from hypothetical unfractionated granitic parents. Numerical AFC modelling shows that, in this case, the Sr–Nd isotopic composition does not directly approach that of the assimilant (e.g. Figure 19 and Keay *et al.* 1997, their figure 3). Because Sr is strongly compatible in feldspars, it is rapidly depleted in the melt, in which its isotopic composition rapidly approaches that of the assimilant. In contrast, Nd is at least initially incompatible, and the melt retains a mantle-like high ϵ_{Nd} signature.

Whether, and at what stage, the melt becomes peraluminous will be controlled by the compositions of the initial mafic component and the assimilant, the ratio of assimilation to fractional crystallisation (that is, the r parameter) and the extent of the process (that is, the f parameter, the amount of liquid remaining). Obviously, strongly fractionated melts that have assimilated little sedimentary material (that is, low r) will be more likely to remain metaluminous and classify as I-types. Some A-type granites, which are thought to have been derived by the direct fractionation of basaltic magma (that is, $r=0$; e.g. Turner *et al.* 1992), may represent an extreme end-member of this spectrum of processes. Melts that

assimilate chemically immature sedimentary rocks (with appreciable Na_2O and CaO) as they fractionate, will also be more likely to generate fractionated metaluminous I-types, whereas those that assimilate chemically mature, peraluminous sedimentary rocks (such as pelites) will produce fractionated S-types.

An important implication is that Tasmanian S-type granites, which are almost all moderately to strongly fractionated, are fundamentally different to the more mafic S-types of the Lachlan Orogen. They have more in common with strongly fractionated I-types, with which they are closely associated, both spatially and temporally in the Heemskirk and Ben Lomond Batholiths (Figures 2, 3).

WESTERN TASMANIA

We have not attempted to formulate a similar quantitative model for the less diverse granites of western Tasmania, from which we have fewer data, because of the likely greater heterogeneity of the sedimentary components. Their relatively high ϵ_{Nd} at moderate to high $^{87}\text{Sr}/^{86}\text{Sr}$ (Figures 14, 15), however, suggests that AFC processes have also been important in their evolution, consistent with their mostly fractionated character. Their $^{87}\text{Sr}/^{86}\text{Sr}$, particularly of the Sandy Cape and Heemskirk White Granites, probably also reflects the isotopically evolved composition of the western Tasmanian sedimentary successions.

GEODYNAMIC IMPLICATIONS

One important difference between (eastern) Tasmania and the mainland Lachlan Orogen is the character and timing of S-type magmatism. The Tasmanian S-type granites are felsic, usually fractionated, and not associated with high temperature-low pressure regional metamorphic complexes. As shown above, they are generated from the interplay of mantle-derived mafic magmas ($\pm$ relatively mafic lower crust) and the sedimentary succession of the Mathinna Supergroup by assimilation-fractional crystallisation (AFC) processes, rather than simple assimilation or mingling with anatectic melts. They are post-tectonic and, in contrast to much of the Lachlan Orogen, post-date early, more-mafic I-types (Black *et al.* 2005).

Collins & Hobbs (2001) suggested that, in the eastern Lachlan Orogen, the change from mafic calcalkaline or shoshonitic volcanism to felsic S-type magmatism in the Early Silurian (~ 430 Ma) was in response to the crustal thickening and burial of Ordovician turbidites in an early Benambran deformation at ~ 440 – 450 Ma. Collins & Richards (2008) proposed a more general plate tectonic model, in which the deformation and crustal thickening, coupled with temporary cessation of arc magmatism, occurred in response to slab advance and flattening of subduction, perhaps caused by a buoyancy anomaly such as an oceanic plateau or continental ribbon on the subducting plate. The ensuing S-type plutonism is thought to indicate the slab retreat phase as the arc re-establishes, with increased heat flux into the newly thickened metasedimentary lower crust. Collins & Richards (2008) proposed that with further retreat (rollback) and steepening of the slab, there is a switch to an extensional regime with crustal thinning, resulting in a reversion to I-type magmatism. They proposed that four cycles of slab rollback, extension and I-type magmatism, followed by flat subduction, crustal thickening and S-type magmatism, could be recognised in the Tasmanides, from the Middle Cambrian to the Late Carboniferous.

In Tasmania, however, there is minimal magmatism and no undisputed deformation between the Early Ordovician and Early Devonian, when continuous, quiet shelf and turbidite deposition prevailed in western and eastern Tasmania, respectively. Mid-Paleozoic tectonism and associated felsic magmatism appears to be restricted to a single cycle, from ~ 400 – 350 Ma, and was probably terminated after complete blocking of the subduction zone by collision with the thick, western Tasmanian lithosphere. In the Lachlan Orogen, in contrast, the accretion of thinner and narrower oceanic plateaux or continental ribbons caused flat subduction, contraction and only temporary cessation of arc magmatism.

Tasmanian S-type (and felsic I-type) granites appear to have been generated in response to crustal thickening during the Tabberabberan Orogeny, but there is little evidence for flat subduction or increased heat flow at the time. The granites themselves have narrow contact aureoles, and are not associated with regional metamorphic complexes. Mantle-derived mafic magmas, which were an important source component of the

granites, may have been either insufficiently voluminous, or too cool, to directly melt and assimilate large volumes of buried sedimentary rocks of the Mathinna Supergroup. Instead, concurrent assimilation-fractional crystallisation, in which some of the heat for crustal melting was supplied by latent heat of crystallisation, was necessary to generate granites. In Tasmania, except perhaps on King Island, there was no reversion to typical I-type granodiorites, and the final plutons were strongly fractionated granites produced by low *r*-value assimilation-fractional crystallisation.

The Tasmanian S-type granites may have more in common with non-extensional or post-kinematic Lachlan Orogen S-types ‘formed after final closure of the western Lachlan backarc as the entire Lachlan Orogen amalgamated’ (Collins & Richards 2008). This apparently refers to S-type intrusions such as the Strathbogie and Pyalong Granites in the Melbourne and Bendigo Zones of Victoria. Rossiter & Gray (2008) also tentatively suggested a correlation between their (Victorian) Central Granite Superprovince, defined mainly on the basis of high barium contents (> 750 ppm Ba at 70% SiO_2) and low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, and eastern Tasmanian granites. This idea, however, needs careful re-evaluation, particularly as few Tasmanian granites (including unfractionated ones) are as high in Ba as the central Victorian examples.

We favour a model in which the early, pre-tectonic, I-type granodiorites of eastern Tasmania were produced over an east-dipping subduction zone, with the trench located to the west. Some westward slab rollback may have occurred, accounting for the ~ 10 million year age difference between the earliest granodiorites in the Blue Tier Batholith and Scottsdale Batholith, but probably neither migration of the magmatic front nor extension was as marked as in the eastern Lachlan Orogen. At this time (~ 400 – 390 Ma), the western Tasmanian terrane was a separate microcontinental block or part of a passive margin, lying well to the west (Figure 20a). The Tabberabberan Orogeny in Tasmania (~ 389 Ma, Black *et al.* 2005) probably represents closure of the intervening ocean crust, and collision of the two terranes, and resulted in crustal thickening sufficient to also allow the generation of S-type granites, at first in the already magmatically active eastern terrane (Figure 20b). Following amalgamation, slab breakoff may have resulted in a short period of continued subduction-related magmatism, followed by upwelling of hot asthenospheric mantle induced by foundering of the slab, resulting in underplating of the lithosphere by mafic magmas. After about 377 Ma, the eastern terrane had cooled considerably and the tectonic regime reverted to extension, allowing the mafic magmas to ascend to the upper crust (preserved as dolerite dyke swarms intruding northeastern Tasmanian granites). By now, only limited further crustal melting, fuelled by latent heat of crystallisation of mafic magmas, could occur, generating strongly fractionated granites by low *r*-value AFC. In western Tasmania, mafic underplating of the lithosphere also occurred, but because of the initially cooler crust, the thermal response was delayed relative to the east. Peak crustal temperatures were probably lower

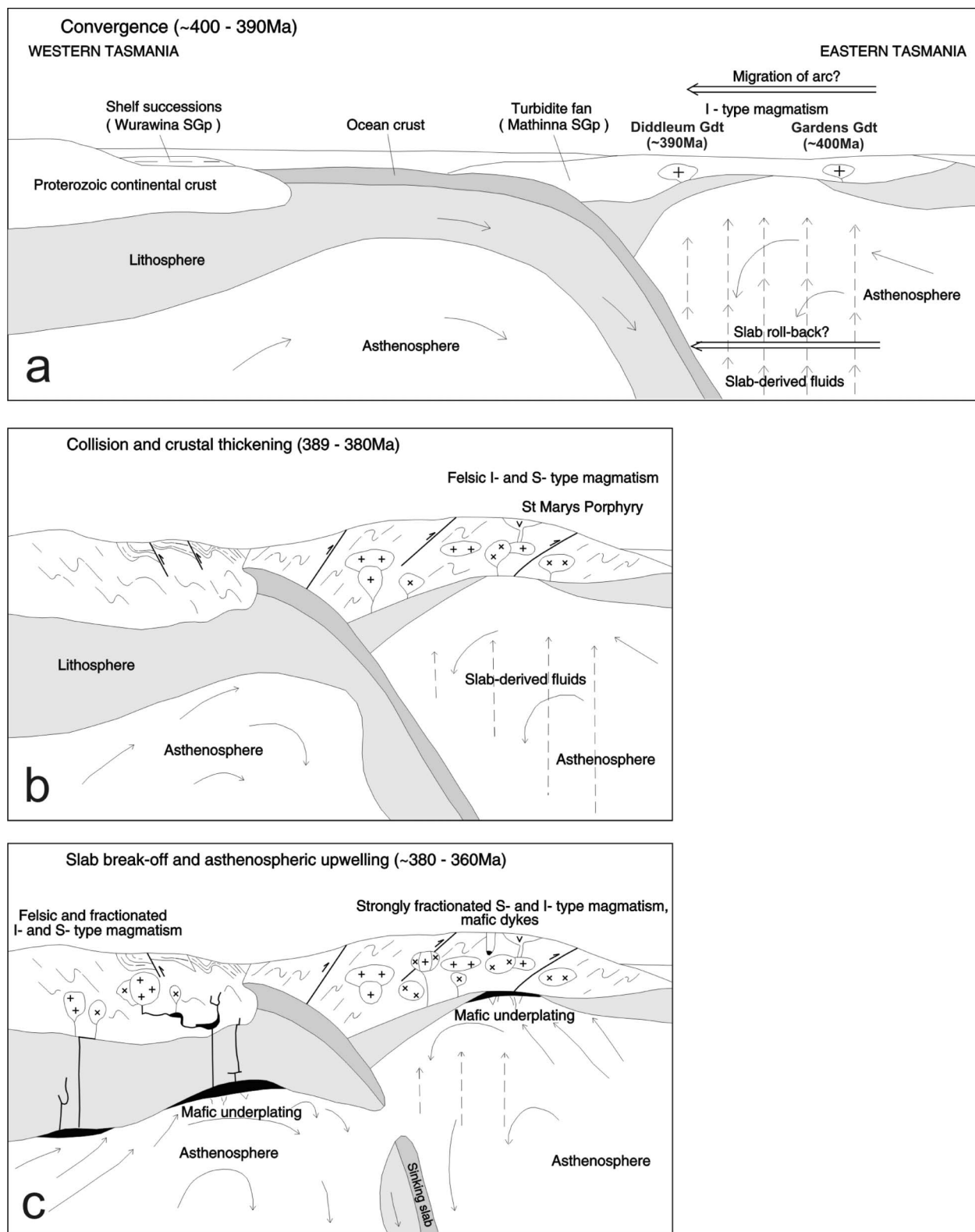


Figure 20 Cartoons depicting possible tectonic setting for Devonian–Carboniferous granite magmatism in Tasmania. (a) ~400–390 Ma: plate convergence with emplacement of arc-related, I-type granodiorites and rare coeval volcanics in eastern Tasmania; possible moderate slab rollback and extension. (b) ~389–380 Ma: crustal thickening and shortening following terrane collision (Tabberabberan Orogeny); generation of felsic S- and I-type granites in eastern Tasmania. (c) ~380–360 Ma: slab breakoff induces asthenospheric upwelling and underplating of lithosphere by mafic magmas; emplacement of late, strongly fractionated granites and dolerite dykes in eastern Tasmania; ascending mafic magmas induce rising crustal temperatures and generation of fractionated granites in western Tasmania.

and attained later, and only felsic, mostly fractionated granites were produced (Figure 20c).

CONCLUSIONS

- (1) The inherited zircon age spectra of the Eastern and Western Tasmanian granites suggest that their respective, and different, enclosing sedimentary terranes were an important source component. Alternatively, in the case of western Tasmania, the granites and the sedimentary rocks might have shared a common source.
- (2) There is no systematic difference in the detrital zircon age spectra between the I-types and S-types in either terrane. This tends to discount the idea that 'granites form an image of their source rocks' (Chappell 1979), that is, that the two types are derived respectively from distinct meta-igneous (infracrustal) and metasedimentary (supracrustal) sources. Rather, it suggests a genetic link: that 'the S-type magmas are heavily contaminated I-magmas' (Collins 1998).
- (3) There is no evidence from the granites for a Proterozoic felsic basement, perhaps similar to the rocks of western Tasmania, beneath the Mathinna Supergroup. Otherwise, it would be expected to be reflected in the inherited zircon. A more mafic basement (which would not contribute significant zircon) remains possible.
- (4) The low proportion of inherited zircon (particularly in I-types) also discounts the idea that the linear trends displayed within suites by Zr (and, by implication, those of other trace elements) when plotted against SiO₂ or total FeO is due to restite unmixing.
- (5) The inherited zircon age spectra of the western Tasmanian granites contain a significant peak at ~1600 Ma, not prominent in the host sedimentary successions, which may be derived, either directly or indirectly, from the Gawler Range Volcanics of South Australia or a correlative of it.
- (6) The correlation of isotopic ratios with major element chemistry and age in the Scottsdale and Blue Tier Batholiths is suggestive of an increasing component from anatexis or assimilation of sedimentary rock (Mathinna Supergroup), and is inconsistent with simple restite unmixing in a closed magma chamber (which would maintain isotopic compositions).
- (7) Unfractionated I-type granodiorites in eastern Tasmania are isotopically similar to similar I-types in the Lachlan Orogen.
- (8) Tasmanian S-types, which are mostly fractionated, are isotopically unlike the typically more-mafic S-types of the Lachlan Orogen, in particular having higher ϵ_{Nd} .
- (9) Higher $^{87}\text{Sr}/^{86}\text{Sr}$ in western Tasmania granites, especially S-types, is consistent with a contribution from their older, isotopically more evolved enclosing sedimentary terrane, although many of their surface country rocks are too mature and infertile to represent this component.
- (10) The Pb isotope data from Tasmanian granites, particularly in the west, have a much larger spread than obtained from the Bega and Berridale Batholiths. This probably reflects the greater diversity of their sedimentary source rocks.
- (11) Mantle-derived mafic magmas were an important source component for all Tasmanian granites.
- (12) Direct generation of strongly fractionated granites by AFC processes may explain their high ϵ_{Nd} and the close spatial and temporal association of fractionated I- and S-types. Thus, fractionated granites can be generated directly in the lower crust, and need not be derived from unfractionated parents.
- (13) Further isotopic studies, particularly of granite-associated mafic rocks and strongly fractionated granites, are warranted in Tasmania. Mineral separation techniques will probably be required to obtain meaningful Sr isotope data from the latter.

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APPENDIX 1: SAMPLE AND ANALYTICAL DETAILS

Samples and stratigraphic nomenclature

Sample locations have been published previously (Black *et al.* 2004, 2005; Turner *et al.* 1998). Note that the Grassy Granite of Black *et al.* (2005) is now known formally as the Sandblow Granite. The Proterozoic successions of western and eastern King Island referred to by Black *et al.* (2004), are now known as the Surprise Bay

Formation (Seymour *et al.* 2007) and the Fraser Formation (Direen & Jago 2008), respectively.

Major and trace element geochemistry

New whole rock major- and trace-element analyses were obtained at the geochemistry laboratory of Geoscience Australia (GA) (Table 1) and are also accessible from the GA database OZCHEM (<http://www.ga.gov.au/gda/>). Rock samples were prepared for chemical analysis by

jaw crushing (2–5 kg) and then grinding of a subsample (50–70 g) in a tungsten carbide ring mill. Abundances of major and trace elements were determined by XRF and ICP-MS. Major and minor elements (Si, Ti, Al, Fe, Mn, Mg, Ca, Na, K, P, S) were determined by wavelength-dispersive XRF on fused disks using methods similar to those of Norrish & Hutton (1969). Precision for these elements is better than $\pm 1\%$ of the reported values. As, Ba, Cr, Cu, Ni, Sc, V, Zn and Zr were determined by wavelength-dispersive XRF on pressed pellets, using methods similar to those described by Norrish & Chappell (1977). Selected trace elements (Cs, Ga, Nb, Pb, Rb, Sb, Sn, Sr, Ta, Th, U, Y, and the rare earth elements) were analysed at GA by ICP-MS (Perkin Elmer ELAN 6000) using methods similar to those of Eggins *et al.* (1997), but on solutions obtained by dissolution of fused glass disks (Pyke 2000). ICP-MS results were calibrated using prepared run standard solutions of varying concentrations, which themselves were calibrated against a series of International Standards. Precisions are 5%, increasing to 10% at low levels. Agreement between XRF and ICP-MS are within 10%. FeO abundances were determined at GA by digestion and electrochemical titration using a modified methodology based on Shapiro & Brannock (1962).

SHRIMP dating of inherited zircons

Most of the technique employed for U–Pb zircon dating is given in Black *et al.* (2005). One difference is that the majority of the analyses were obtained on the SHRIMP I ion-microprobe at the RSES, Australian National University, rather than the SHRIMP II instrument. The data are accessible from OZCHRON, the geochronology database of Geoscience Australia (<http://www.ga.gov.au/oracle/ozchron/index.jsp>).

Sr, Nd and Pb isotopes

Sr, Nd and Pb isotopic compositions of whole rock samples of the granites and selected country rocks were determined by J. D. Foden at the School of Earth and Environmental Sciences, University of Adelaide, on a Finnigan MAT 262 Thermal Ionisation Mass Spectrometer operated in static mode. Details of the procedures used to determine the radiogenic isotopes are provided in Elburg & Foden (1998) and Elburg *et al.* (2003).

For making age corrections, the elemental concentrations of Sm and Nd were determined by isotope dilution, and therefore have high precision, even at low concentrations, which we take nominally as 1 relative % (2σ). The elemental concentrations of Rb and Sr were determined by XRF, and U, Th and Pb by ICP-MS. We estimate the uncertainty in these elements (at 2σ) as 1 relative %, or the detection limit (Rb 0.3 ppm, Sr 0.5, U 0.07, Th 0.01, Pb 0.5), whichever is the greater.

APPENDIX 2: ADDITIONAL ND ISOTOPE DATA FROM FRACTIONATED GRANITES IN TASMANIA

Appendix 2 Table

Pluton	Province	Batholith	Type	Sample no.	mE (AGD66)	mN (AGD66)	Age (est.)	SiO ₂ wt%	Rb ppm	Sr ppm	Nd ppm	Sm ppm	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd _(m)	T _{DM2} (Ma)	ϵ_{Nd}
Royal George	Eastern	Ben Lomond	SFf	R000547	573470	5366620	376.5	75.68	420	31	25.83	6.82	0.1596	0.512384	1291	–3.2
Gipps Creek	Eastern	Ben Lomond	SFf	R004493	556420	5385000	376.5	72.67	480	68	22.12	5.39	0.1474	0.512282	1396	–4.6
Mt Stronach	Eastern	Scottsdale	IFf	SB6	554100	5440500	395	75.95	370	5	12.78	4.23	0.2000	0.512611	1113	–0.7
Strzelecki	Eastern		SFf	R006956	588560	5544970	384.9	76.12	320	51	23.18	5.95	0.1553	0.512289	1418	–4.8
Conical Rocks	Western		SFf	99220044	329007	5379562	357	75.38	340	32	19.97	4.86	0.1473	0.512196	1515	–6.4
Meredith	Western	Meredith	IFf	G401385	353500	5387300	373	77.76	340	6	18.70	5.51	0.1784	0.512383	1358	–4.1

Courtesy of D. C. Champion. R. Maas, analyst. Abbreviations as in Table 2.