Australian Journal of Earth Sciences (2001) **48**, 25–41

What caused the Early Silurian change from mafic to silicic (S-type) magmatism in the eastern Lachlan Fold Belt?

W. J. COLLINS[1]* AND B. E. HOBBS[2]

[1]*Department of Geology, University of Newcastle, Newcastle, NSW 2308, Australia.*
[2]*Australian Geodynamics CRC, CSIRO Exploration and Mining, Private Bag, PO Wembley, WA 6014, Australia.*

One of the most significant, but poorly understood, tectonic events in the east Lachlan Fold Belt is that which caused the shift from mafic, mantle-derived calc-alkaline/shoshonitic volcanism in the Late Ordovician to silicic (S-type) plutonism and volcanism in the late Early Silurian. We suggest that this chemical/isotopic shift required major changes in crustal architecture, but not tectonic setting, and simply involved ongoing subduction-related magmatism following burial of the pre-existing, active intraoceanic arc by overthrusting Ordovician sediments during Late Ordovician – Early Silurian (pre-Benambran) deformation, associated with regional northeast–southwest shortening. A review of ‘type’ Benambran deformation from the type area (central Lachlan Fold Belt) shows that it is constrained to a north-northwest-trending belt at *ca* 430 Ma (late Early Silurian), associated with high-grade metamorphism and S-type granite generation. Similar features were associated with *ca* 430 Ma deformation in east Lachlan Fold Belt, highlighted by the Cooma Complex, and formed within a separate north-trending belt that included the S-type Kosciuszko, Murrumbidgee, Young and Wyangala Batholiths. As Ordovician turbidites were partially melted at *ca* 430 Ma, they must have been buried already to ~20 km before the ‘type’ Benambran deformation. We suggest that this burial occurred during earlier northeast–southwest shortening associated with regional oblique folds and thrusts, loosely referred to previously as latitudinal or east–west structures. This event also caused the earliest Silurian uplift in the central Lachlan Fold Belt (Benambran highlands), which pre-dated the ‘type’ Benambran deformation and is constrained as latest Ordovician – earliest Silurian (*ca* 450–440 Ma) in age. The south- to southwest-verging, earliest Silurian folds and thrusts in the Tabberabbera Zone are considered to be associated with these early oblique structures, although similar deformation in that zone probably continued into the Devonian. We term these ‘pre’- and ‘type’-Benambran events as ‘early’ and ‘late’ for historical reasons, although we do not consider that they are necessarily related. Heat-flow modelling suggests that burial of ‘average’ Ordovician turbidites during early Benambran deformation at 450–440 Ma, to form a 30 km-thick crustal pile, cannot provide sufficient heat to induce mid-crustal melting at *ca* 430 Ma by internal heat generation alone. An external, mantle heat source is required, best illustrated by the mafic *ca* 430 Ma, Micalong Swamp Igneous Complex in the S-type Young Batholith. Modern heat-flow constraints also indicate that the lower crust cannot be felsic and, along with petrological evidence, appears to preclude older continental ‘basement terranes’ as sources for the S-type granites. Restriction of the S-type batholiths into two discrete, oblique, linear belts in the central and east Lachlan Fold Belt supports a model of separate magmatic arc/subduction zone complexes, consistent with the existence of adjacent, structurally imbricated turbidite zones with opposite tectonic vergence, inferred by other workers to be independent accretionary prisms. Arc magmas associated with this ‘double convergent’ subduction system in the east Lachlan Fold Belt were heavily contaminated by Ordovician sediment, recently buried during the early Benambran deformation, causing the shift from mafic to silicic (S-type) magmatism. In contrast, the central Lachlan Fold Belt magmatic arc, represented by the Wagga–Omeo Zone, only began in the Early Silurian in response to subduction associated with the early Benambran northeast–southwest shortening. The model requires that the S-type and subsequent I-type (Late Silurian – Devonian) granites of the Lachlan Fold Belt were associated with ongoing, subduction-related tectonic activity.

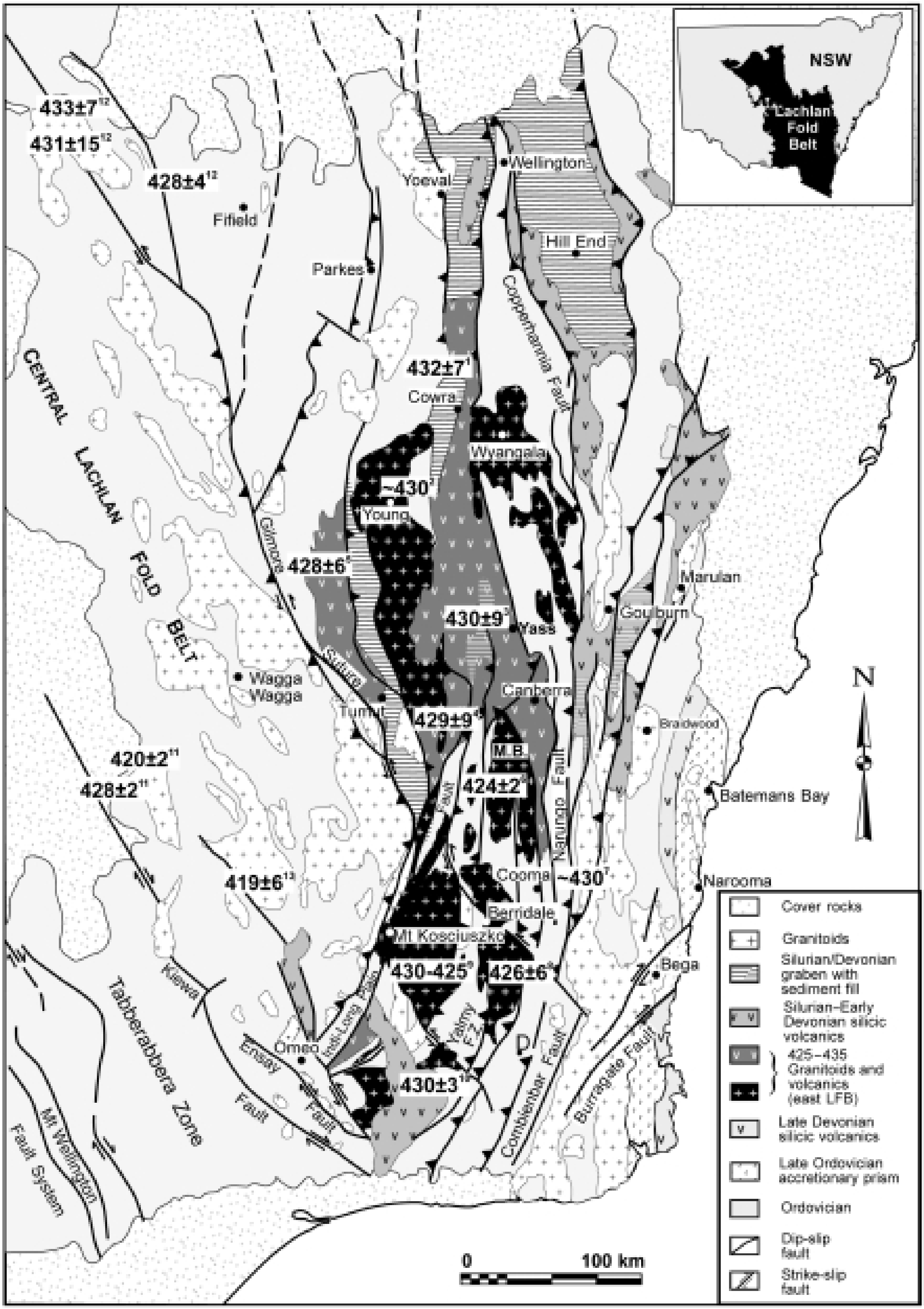
KEY WORDS: basement terranes, Benambran, heat flow, magmatic arcs, S-type granite, subduction.

INTRODUCTION

One of the characteristic features of Ordovician tectonism in the east Lachlan Fold Belt is the occurrence of calc-alkaline to shoshonitic basalts, which have been recognised as part of an intraoceanic arc environment (Glen *et al.* 1998), consistent with early tectonic interpretations (Oversby 1971; Scheibner 1973). Aside from typical trace-element abundances of intraoceanic arcs, critical evidence is the isotopic composition of the volcanics: ϵ_{Nd} values range from + 6.5 to + 8.0 and from + 4.2 to + 6.7 for the Early and Late Ordovician basalts, respectively, indicating minimal crustal contamination (Wyborn & Sun 1993; Watkins in Glen *et al.* 1998). Very low $^{207/204}$Pb and $^{208/204}$Pb isotopic ratios for the mafic rocks also indicate an exclusive mantle origin (Carr *et al.* 1995), which is consistent with an intraoceanic arc environment.

*Corresponding author: bcollins@geology.newcastle.edu.au

In contrast, Early Silurian magmatism is characterised by extensive silicic S-type granites and comagmatic volcanics, focused in a north-trending belt of batholiths in the east Lachlan Fold Belt and a north-northwest-trending belt in the central Lachlan Fold Belt (Figure 1). Compilation of recent U–Pb SHRIMP ages indicates that most of these batholiths were emplaced between 435 and 425 Ma. These S-type granitic rocks have evolved isotopic compositions,



with ϵ_{Nd} values ranging from –5.8 to –9.2 (McCulloch & Chappell 1982), with more evolved compositions similar to the –9.5 to –10.5 ϵ_{Nd} range for Ordovician Lachlan Fold Belt turbidites (McCulloch & Woodhead 1993). These isotopic compositions require that the S-type granites be derived from a large component of metasedimentary material, either Proterozoic continental crust (Chappell *et al.* 1988) or Ordovician sedimentary rocks (Gray 1984, 1990; Collins 1996, 1998). The major focus of this paper, therefore, is to understand what caused the marked chemical and isotopic shift from mantle to continental source rocks between the Late Ordovician and Early Silurian.

The shift from mafic to silicic volcanism followed a fundamental change from deep- to shallow-water facies sedimentation, which occurred in response to major regional uplift of the east and central Lachlan Fold Belt. The uplift is generally explained by widespread crustal thickening loosely associated with Benambran deformation (Crook *et al.* 1973). However, the Benambran deformation was probably contemporaneous with silicic magmatism in the type area, as discussed below, and cannot account for the burial and necessary instantaneous partial melting of the Ordovician turbidites during migmatite formation and S-type granite generation. This paper summarises the evidence for two stages of Benambran deformation (Table 1), an early stage equating to what have been loosely called ‘east–west’ or ‘latitudinal’ structures (Glen 1992), overprinted by a late stage of intense Benambran deformation, which more closely relates to that described in the type area, the Omeo Complex, where it was originally defined. We argue that the early Benambran structures were produced by regional northeast–southwest shortening in the latest Ordovician – earliest Silurian and we assess its contribution to heat flow and granite production in the middle–deep crust during the late Early Silurian, when the ‘type’ or late Benambran deformation occurred. Finally, we develop a tectonic setting to account for the geological evolution of the east and central Lachlan Fold Belt during the two Benambran events. The Tucker and McKerrow (1995) time-scale is applied throughout.

Figure 1 Geological map of Lachlan Fold Belt (LFB) showing the major features and locations discussed in this paper. North-trending magmatic belt of S-type igneous rocks in east Lachlan Fold Belt is highlighted. Age determinations in million years are from (numbers refer to superscripts after the ages): 1, Canowindra Volcanics (Pogson & Watkins 1998); 2, Young Batholith, based on comagmatic relation with the Goobarragandra Volcanics (Owen & Wyborn 1979); 3, Micalong Swamp Basic Igneous Complex (Owen & Wyborn 1979); 4, Goobarragandra Volcanics (Owen & Wyborn 1979); 5, Frampton Volcanics (Stuart-Smith *et al.* 1992); 6, Murrumbidgee Batholith (Roddick & Compston 1976); 7, Cooma granite (Williams 1995); 8, minimum age for S-type plutons, based on age of late I-type Maffra Adamellite, Berridale Batholith (Ireland & Gibson 1998); 9, Kosciuszko Batholith (Chappell *et al.* 1991b); 10, Deddick Granodiorite, Kosciuszko Batholith (Maas *et al.* 1998); 11, Yabba Adamellite 428 Ma, Bethanga Granite 420 Ma (Keay *et al.* 1999); 12, Tarran Volcanics 433 Ma, Erimeran Granite 431 Ma, Mineral Hill Volcanics 428 Ma, Wagga Batholith (Isaacs *et al.* 2000); 13, Koetong Adamellite, Koetong Batholith (Anderson *et al.* 1996). D, Delegate; F, fault; F.Z., fault zone; M.B., Murrumbidgee Batholith.

‘TYPE’ BENAMBRAN DEFORMATION

Definition

The Benambran deformation event, as described in the literature, is enigmatic and complex. The type area is a fault contact and inferred unconformity between metamorphosed Lower to Middle Ordovician metasediments (Pinnack Sandstone) and overlying Middle Silurian Mitta Mitta Rhyolite of the Wombat Creek Group in the Omeo Complex (Talent 1965; VandenBerg *et al.* 1998). The time interval encompassed by the ‘type’ unconformity is at least 30 million years, during which time a variety of tectonic events have been inferred. These events include the Late Ordovician – Early Silurian regional facies change in the

Table 1 Summary of Late Ordovician – Early Silurian deformation events in the east and central Lachlan Fold Belt (LFB).

Age	450–440 Ma (Early Benambran)	435–423 Ma (Late Benambran)
Wagga–Omeo (central LFB)	Uplift of ‘Benambran highlands’. Formation of carbonate platform.	Type area: high-grade metamorphism, migmatisation, S-type granite emplacement, NE–SW shortening.
Tabberabbera Zone (central LFB)	N–NE-directed thrusting and folding.	N–NE-directed thrusting and folding.
Yalmy Fault Zone (east LFB)	Inferred ‘E–W’-trending folds.	Influx from ‘Benambran highlands’. Meridional folding, E-directed thrusting (‘second’ Benambran event). S-type granite emplacement.
Cooma Complex – Murrumbidgee Batholith – Tantangara – Brindabella (east LFB)	‘First’ Benambran event. Uplift of Benambran highlands.	‘Second’ Benambran event. Tight meridional folds, W-directed thrusts. High-grade metamorphism, migmatisation. S-type granite emplacement.
Wyangala Batholith (east LFB)	Oblique-trending thrusts.	Tight–isoclinal meridional folds. S-type granite emplacement.
Narooma Complex (east LFB)	Recumbent ‘E–W’-trending folds. NE-directed thrusting.	?

east Lachlan Fold Belt, resulting from uplift of 'Benambran highlands' (Crook *et al.* 1973; VandenBerg 1998). This is consistent with the 'first' Benambran event regional uplift defined by Owen and Wyborn (1979), which 'uplifted and folded' the Wagga–Omeo Zone in the Late Ordovician (pre-early Llandovery). However, the north-northeast to north-east-trending Yalmy Fault Zone (Glen & VandenBerg 1987), which deforms Upper Llandovery sedimentary rocks derived from those highlands, has also been assigned to the Benambran deformation (Glen 1992; VandenBerg 1998). This event appears to correlate with the 'second' Benambran event of Owen and Wyborn (1979), which also folded Upper Llandovery sediments into tight meridional folds. Nonetheless, east–west-trending folds are also inferred to be part of regional Benambran dextral transpression (Powell 1983; Fergusson & Coney 1992; Glen 1992). Thus, it appears that several, possibly unrelated, tectonic events have been grouped as Benambran deformation.

Initially, we separated the two events into 'pre-' and 'type' Benambran, because we felt that they were not necessarily related. It has been suggested that we retain the term 'Benambran' for both events, which we do largely for historical reasons, particularly as Owen and Wyborn (1979) described the first and second Benambran fold episodes (Table 1). However, we emphasise that it is more important to consider the structures in terms of convergent-margin tectonic complexity, rather than assigning them to discrete deformation events.

A more precise age for the 'type' or late Benambran event comes from structural and geochronological analysis of the high-grade metamorphic rocks in the Omeo Complex. At Omeo, major deformation produced multiple foliations, which are associated with high-grade metamorphism and plutonism in the Omeo Metamorphic Complex (Morand 1988; Willman *et al.* 1999). The first foliation is a transposed surface (S_1) that contains rootless intrafolial folds; the second (S_2) is the dominant foliation in the region, being the axial surface of tight F_2 folds. These folds are upright and northwest-trending in the low-grade rocks and northeast-inclined to recumbent in the high-grade rocks, verging to the south or southwest, suggesting southwest-directed tectonic transport. Later mesoscale, east–west-trending, upright F_3 folds were localised and not of apparent regional significance.

High-grade metamorphism occurred early in the deformation event because sillimanite needles and migmatitic leucosomes define the S_1 foliation (Morand 1990). The high-T, low-P metamorphic zones culminate with migmatitic rims around large granitic plutons (Fagan 1979; Morand 1990) indicating an intimate relationship between plutonism, metamorphism and deformation during the 'type' or late Benambran event.

Recently, the syntectonic, synmetamorphic, S-type Yabba granite and an early I-type tonalite, from approximately 40 km north-northwest of Omeo, have been precisely dated by the U–Pb SHRIMP method on zircon at 428.4 ± 2.1 Ma and 428.5 ± 1.4 Ma, respectively (Keay *et al.* 1999). Provided that the structural correlations of Fagan (1979) apply throughout the Omeo Complex, the ages tightly constrain 'type' or late Benambran deformation at *ca* 430–426 Ma, which is late Llandovery.

Late 'type' Benambran deformation elsewhere

TABBERABBERA ZONE

Fergusson (1987, 1998a) documented south- to southwest-verging folding and thrusting in the Tabberabbera Zone, which deforms the Cabbannah Group, a thick, quartz-rich Lower Silurian turbidite succession correlated with the Yalmy Group in the east Lachlan Fold Belt. Foster *et al.* (1999) dated micas in highly deformed slates formed during the southwest-verging thrusting that yielded an Ar–Ar age spectra range from 440 to 410 Ma. The oldest ages are in the northernmost region, near the present Kiewa Fault Zone and micas become younger progressively southwest, in the direction of thrusting. These structures are consistent in age, geometry and vergence with those of the high-grade rocks in the Omeo Metamorphic Complex, described above. Ar–Ar age spectra from similar rocks in the west Lachlan Fold Belt have been interpreted as indicating reactivation during discrete tectonic events (VandenBerg 1999), rather than migrating deformation fronts (Gray *et al.* 1997; Foster *et al.* 1998, 1999), but this has not been shown for the Tabberabbera Zone, although more analyses are needed.

Deformation in the Tabberabbera Zone, therefore, appears to have occurred semi-continuously from Late Ordovician to Early Devonian by south- to southwest-verging thrusting and folding. This time span encompasses the entire range of previously defined Benambran deformation events, but has similar orientation and tectonic vergence to the late Benambran deformation in the central Lachlan Fold Belt.

YALMY FAULT ZONE

The Yalmy Fault Zone (Glen & VandenBerg 1987) thrusts Upper Ordovician Bendoc Group over middle to upper? Llandoverian Yalmy Group (Lewis *et al.* 1994). Recent U–Pb zircon age determinations of Kosciuszko Batholith plutons that intrude the Yalmy Fault Zone indicate that deformation is of a similar age to late Benambran deformation, and not mid-Silurian as originally inferred. The zircon age for the post-tectonic Deddick Granodiorite is 430 ± 3 Ma (Maas *et al.* 1998), and for the Nunniong and Amboyne granites, it is 430 ± 13 Ma and 427 ± 3 Ma, respectively (VandenBerg 1998). Given the middle to late? Llandovery age of the Yalmy Group, the Yalmy Fault Zone is tightly constrained at *ca* 435–425 Ma, coeval with the late Benambran event (Table 1).

The Yalmy Fault Zone can be equated with the 'second' Benambran event of Owen and Wyborn (1979), which is tightly constrained between deposition of the lower Llandoverian Tantangara Formation and upper Llandoverian Peppercorn Formation. This is consistent with a *ca* 430 Ma age, using the Tucker and McKerrow (1995) time-scale. The deformation produced tight meridional folds and a penetrative slaty cleavage that can be traced southward across the Berridale Fault into structures of the Yalmy Fault Zone. Owen and Wyborn (1979) also suggested that the 'second' Benambran event was associated with regional high-temperature metamorphism and large-scale anatexis, typical of events associated with the late 'type' Benambran deformation in the Wagga–Omeo Zone.

Fergusson and VandenBerg (1990) correlated deformation of the Yalmy Fault Zone with folding and thrusting to the northeast in the Bungonia region, east of Goulburn (Figure 1), and termed the inferred structure the Bungonia–Delegate fold and thrust zone. However, the correlation is problematic because Fergusson and VandenBerg (1990) indicated that deformation at Bungonia must have occurred during the Late Silurian to Early Devonian. Fergusson (1998b) reiterated this time range for the earliest deformation (D_1) at Bungonia. The non-correlation is significant because it suggests that the Bungonia–Delegate thrust belt is composed of different-aged deformation structures.

COOMA COMPLEX – MURRUMBIDGEE BATHOLITH

The earliest recognised regional structures in the low-grade regions of the Cooma Complex are upright to inclined, west-verging, tight meridional folds (F_1: Hopwood 1976). These are overprinted by a variably dipping crenulated surface (S_2), which is overprinted by late retrogressive (D_3) shear zones (Hopwood 1976). A similar structural sequence exists in the high-grade rocks (Hopwood 1976), but there the earliest foliation is a well-developed, layer-parallel gneissosity (S_1), rather than bedding. Johnson and Vernon (1995) agreed with the structural sequence defined by Hopwood (1976), but added two earlier schistosities in the low-grade rocks, based on crenulated internal foliations in porphyroblasts. Accordingly, the D_1, D_2 and D_3 structures of Hopwood (1976) are D_3, D_4 and D_5, according to Johnson and Vernon (1995) and the latter terminology is used here. The second foliation (S_2) of Johnson and Vernon (1995) appears to correlate with the regional S_1, as defined by Granath (1980), but their S_1 foliation does not appear to be regionally significant. Both Hopwood (1976) and Johnson and Vernon (1995) considered that the regional inclined to upright, north-trending (F_3) fold structures, between the high- and low-grade regions, were the same fold generation.

In contrast, recent mapping at Cooma (Figure 2) has revealed that the regional upright north-trending folds of the high-grade (migmatite) zone are a later generation (F_5) than those (F_3) in the low-grade region. As the high-grade zone is approached from the west, the axial-plane slaty (S_3) cleavage becomes the 'stripy-' or 'corduroy-gneiss' in the cordierite–andalusite–K-feldspar metamorphic zone at Spring Creek (Johnson & Vernon 1995). At Spring Creek, lithological layering (S_0) outlines the western limb of the F_3 Spring Creek Anticline (Figure 2, top left), but ~1 km farther eastward in Snake Creek, lithological layering is strongly transposed parallel to S_3, and the dominant foliation is a composite S_0/S_3 fabric. The change coincides with the pervasive development of stromatic migmatites, which define the high-grade zone of the Cooma Complex, and the boundary is broadly coincident with the enveloping S_3 surface (Figure 2). In this high-grade zone west of Cooma township, S_0/S_3 is folded about box-like, east–west-trending, conjugate F_4 folds, which are refolded by a north–south-trending set of regional upright F_5 folds (Figure 2). Farther north, east of Snake Creek, S_0/S_3 outlines tight, upright, meridional F_5 folds throughout the migmatites, similar to F_3 in geometry and orientation, but erroneously correlated with the F_3 folds of the low-grade zones (Hopwood 1976).

The Cooma granite intruded syn-D_5 as subvertical sheets aligned subparallel to F_5 axial planes, but it also has the S_5 foliation, which is the gneissic fabric in the pluton. At several localities (Figure 3), abundant migmatitic enclaves in the granite preserve S_0/S_3 orientations that show mesoscale, upright F_5 fold structures. Within the F_5 hinge zones, a well-developed S_5 axial-plane foliation contains smaller gneissose enclaves and remnants of D_3 leucosomes that are aligned parallel to S_5 and oblique to the S_3 orientations of migmatites in the limbs. Also shown in F_5 limb regions are migmatites that have dissagregated along the S_3 foliation to form Cooma granite. This indicates transport of the small enclaves along S_3 into F_5 hinge zones during magma flow, and subsequent migration along the S_5 axial-planar surface during F_5 folding. Therefore, the Cooma granite developed and intruded during D_3–D_5, associated with ongoing regional east–west-compressive deformation.

The most recent estimate for the age of the Cooma granite is 435–430 Ma (Williams 1995), which is middle–late Llandovery. As inferred Yalmy Group sandstones were folded during D_3 deformation at Cooma (Glen in Lewis *et al.* 1994), as well as the S-type Bennetts Creek Volcanics (White *et al.* 1977), an extremely short burial time is required for these sediments to reach the estimated 10 km depth of the Cooma Complex (Ellis & Obata 1992) before metamorphism. The *ca* 435–430 Ma age of these folds, and their general north–south orientation, is similar to the Yalmy Fault Zone structures, which implies that they are all age equivalents of the late Benambran event (Table 2).

The regional north-trending F_3 and subconcordant F_5 folds continue northward for almost 100 km through the S-type Murrumbidgee Batholith, although their effects progressively weaken in that direction. The high-grade metamorphics of the Cooma Complex also extend northward through the batholith, for almost 50 km. The Murrumbidgee granite plutons have S_3 foliation and therefore pre-date the Cooma granite. Nonetheless, the plutons cut S_3 locally, suggesting syn-D_3 emplacement. This is consistent with the character of the gneissosity in the granites, which is defined by large grains of strongly aligned biotite, plagioclase and quartz, with minimal evidence for recrystallisation. We interpret this as a high-temperature submagmatic flow foliation, distinguishable from the lower grade mylonitic fabric associated with the Murrumbidgee Fault, which has a distinctive mylonitic S/C fabric characterised by a strong retrograde stretching lineation and intense recrystallisation of quartz. Given the similar structural timing of emplacement as the Cooma granite, the age of the Murrumbidgee Batholith is also probably *ca* 435–430 Ma, rather than the Rb–Sr isochron age of 424 ± 2 Ma determined by Roddick and Compston (1976).

The regional north-trending S_3 cleavage in the low-grade rocks extends westward into the Tantangara region, where it is part of the 'second' Benambran event described by Owen and Wyborn (1979). This foliation is of a similar age to the *ca* 430 Ma Cooma granite at Cooma and to the 430 Ma plutons of the southern Kosciuszko Batholith (described above), which confirms the Yalmy Fault Zone as late Benambran and part of late Llandovery regional deformation. Thus, it appears that the Cooma granite and the major S-type batholiths of Kosciuszko and

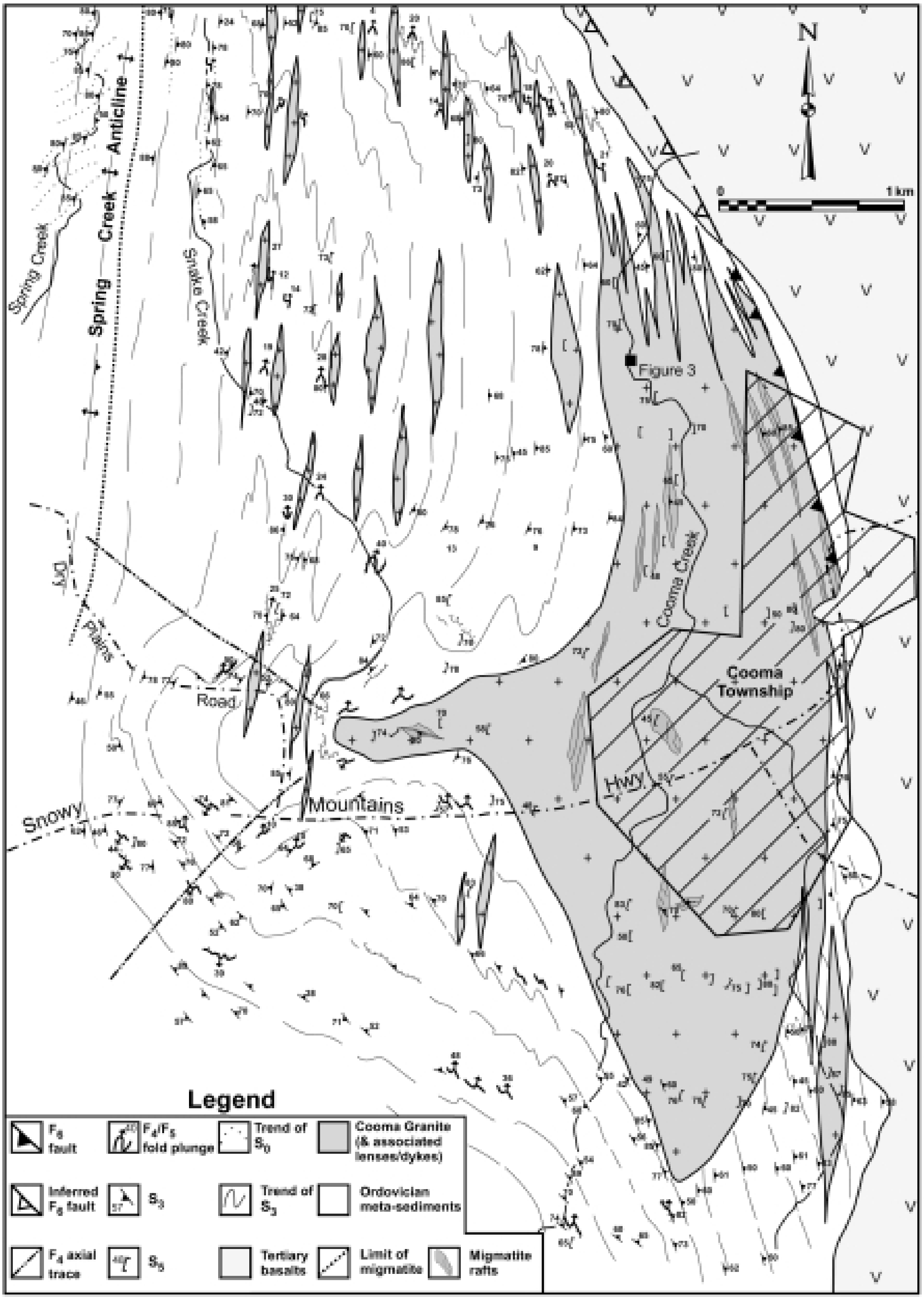


Figure 2 Structural map of the Cooma Complex showing outline of migmatitic foliation (S_3). The pluton developed its shape during D_4–D_5, as it partly outlines the F_4 structure, but granite lenses and dykes also cut this structure and are aligned concordant with S_5. See Figure 1 for location of Cooma township.

Murrumbidgee, and probably Berridale, were all emplaced during the same regional tectonic event at *ca* 430 Ma.

WYANGALA BATHOLITH

Hobbs (1965) and Morand (1987) inferred that a regional deformation event produced tight to isoclinal meridional folds prior to intrusion of S-type plutons in the Wyangala Batholith. However, weak magmatic foliations parallel to the major structures were recorded in some granitoids, which suggests that the batholith was 'late tectonic' with respect to that deformation (Paterson *et al.* 1990). Furthermore, individual plutons are elongate parallel to the major structures, locally sheet-like and indicate

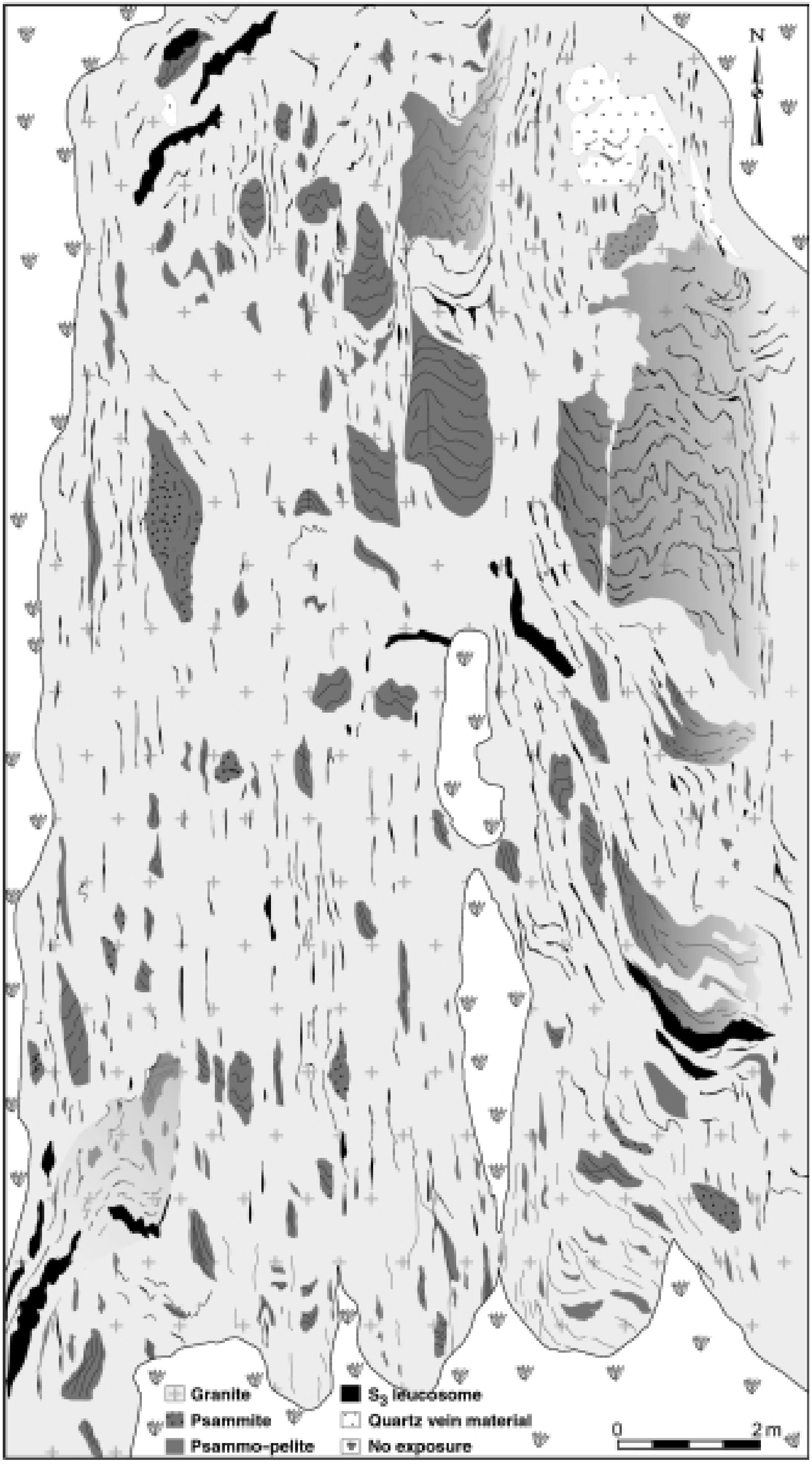


Figure 3 Detail of internal structure outlined by migmatitic enclaves in the Cooma granite. S_3 in the migmatites defines a north-trending F_5 fold that has been intruded by the Cooma granite. In places, the granite appears to be segregated from the migmatite (bottom right), in others it has 'exploded' the F_5 folds (top left), or has concentrated into the axial plane (S_5), where small enclaves are highly aligned (top right). The well-developed subvertical S_5 foliation in these zones reflects ongoing D_5 deformation (see Figure 2 for location). Lighter shades of grey reflect the progressive transition from metasediment to granite.

regional structural control, similar to that seen in the Murrumbidgee Batholith. Fold-axis orientations and asymmetries in the wall rocks usually indicate that deformation occurred during east- to northeast-directed thrusting, consistent with dextral transpression (Paterson *et al.* 1990).

The age of the early meridional structures at Wyangala can only be inferred. Previously, they had been assigned as Silurian–Devonian (Morand 1987), Middle Silurian (Paterson *et al.* 1990) and Early Silurian (Glen & Wyborn 1997). A more precise age for the meridional pre- to syn-batholith deformation can perhaps be gleaned from the 432 ± 7 Ma U–Pb SHRIMP age of the S-type Canowindra Volcanics (Pogson & Watkins 1998). These volcanics are comagmatic with, and intruded by, the S-type Cowra Granite, which marks the northern termination of the Wyangala Batholith (Chappell *et al.* 1991a). The early to middle Wenlockian age of the Canowindra Volcanics, based on fossil assemblages (Pogson & Watkins 1998), is consistent with the lower U–Pb age limit. A minimum age for the early S-type plutons comes from the 425 ± 4.7 Ma I-type Carcoar granite (Lawrie *et al.* 1998), which is a late phase of the batholith. This evidence suggests that the 'pre- to syn-batholith' deformation occurred in the Early Silurian, between 435 and 425 Ma, coeval with the late Benambran deformation.

Summary

The 430 Ma late Benambran deformation is part of a major north–northwest-trending structural, magmatic and high-grade metamorphic belt that constitutes most of the central Lachlan Fold Belt. Similar-trending structures also occur in the adjacent, lower grade Tabberabbera Zone and are part of a south- to southwest-verging, southward-migrating deformation belt that existed throughout the Silurian (Foster *et al.* 1999). The Yalmy Fault Zone is also *ca* 430 Ma, and can be traced through a north-trending belt of syntectonic S-type batholiths in the east Lachlan Fold Belt, which include Kosciuszko, Berridale, Murrumbidgee, Young and Wyangala. These structures are coeval with the late Benambran deformation, but are not part of the central Lachlan Fold Belt. Rather, they form a separate structural, magmatic and (locally) high-grade metamorphic belt confined to the east Lachlan Fold Belt.

EARLY BENAMBRAN DEFORMATION

Many structures have been recognised in the east Lachlan Fold Belt that pre-date the late Benambran features described above. They are generally described as latitudinal or 'east–west' (Glen 1992) because they formed at a high angle to the meridional Benambran structures, but we argue that they were originally northwest-trending structures that formed parallel to, and contemporaneously with, the older (~ 440 Ma) structures in the Omeo and Tabberabberra Zones of the central Lachlan Fold Belt.

Some of the best known early recumbent fold structures in the east Lachlan Fold Belt are those at Bermagui (Williams 1971), 25 km south of Narooma (Figure 1). Miller and Gray (1996) demonstrated that these north-northwest-trending folds developed above a décollement of tectonic melange at Narooma and Batemans Bay, and suggested that they are part of an accretionary complex that developed at the eastern extremity of the east Lachlan Fold Belt. Stretching lineations and discrete shear bands in the melange at Narooma suggest an early southwest–northeast compression associated with the recumbent folding and imbrication. The structures are consistent with oblique (northeast-directed) thrusting above a west-dipping subduction zone (Miller & Gray 1996). Ar–Ar analysis of white micas that formed in the melange, near the base of the complex, constrain the age of thrusting during accretion to 450–445 Ma (Offler *et al.* 1998), which is latest Ordovician.

Elsewhere in the east Lachlan Fold Belt, evidence for early oblique folding and thrusting is sporadically developed. Glen (1992 p. 363, figure 7) summarised the available evidence, noting 'latitudinal' folds and cleavage in the Girilambone Group west of Parkes, adjacent to the Tumut trough, around the Wyangala Batholith, and in a meridional belt extending from east of Canberra to Delegate, near the Victorian border (Figure 1). These early oblique structures deform Upper Ordovician sedimentary rocks and have been refolded into meridional orientations, commonly by the late Benambran deformation. Thus, the time interval of early Benambran deformation is Late Ordovician to late Early Silurian.

The early oblique structures in the Girilambone Group are correlated with structures in the Wagga–Omeo Zone (Fergusson & Coney 1992; Glen 1992). A maximum age is given by the Middle to Upper Ordovician Girilambone Group (Iwata *et al.* 1995) and these structures are overprinted by meridional folds that probably extend eastward into the Parkes volcanic belt, but which underlie the Middle Silurian Forbes Group (Sherwin 1996). Therefore, the overprinting event is likely to be late Benambran and the earlier east–west folding is probably early Benambran (> 430 Ma).

In the Tumut region, Stuart-Smith *et al.* (1992) suggested that major east–west recumbent folding and thrust-faulting occurred at *ca* 425 Ma, although their analysis has been disputed by Scheibner and Basden (1996) among others. In particular, a major difficulty exists with the timing of deformation and duration of the Tumut Trough. The model of Stuart-Smith *et al.* (1992) requires that the Benambran deformation post-dated eruption of the Frampton Volcanics at 428 ± 6 Ma, but if so, an impossibly short-time interval is required for that deformation, followed by extension, deposition and emplacement of the North Mooney Complex late in Tumut Trough history, at 426 ± 6 Ma. Most other workers place the east–west structures, along with all Benambran deformation, before initiation of the Tumut Trough.

Farther east, around the Wyangala Batholith, Glen and Wyborn (1997) suggested that Upper Ordovician black shale units represented early 'east–west-trending' thrusts, which were overprinted by tight to isoclinal, meridional, pre- to syn-batholith 'F_1' folds. This interpretation would also imply that early thrusting was not associated with penetrative cleavage development, as the 'F_1' folds should otherwise have an axial-plane crenulation cleavage in the shale units. The lack of early east–west cleavage suggests that east–west thrusting occurred at high crustal levels, where competency contrasts are most common and strain may be partitioned entirely within very narrow fault surfaces. The age of the earlier east–west folds is post-Upper Ordovician Warbisco Shale and pre-Wyangala Batholith emplacement at *ca* 430 Ma. Therefore, the highly oblique structures must be earliest Silurian, or possibly latest Ordovician, and pre-date late Benambran deformation.

Southward from the Wyangala Batholith, pre-Benambran folds exist in the Molong–Wyangala Zone of Scheibner and Basden (1996), which becomes the Kuark Zone in Victoria (Gray 1988), the eastern boundary apparently being the Combienbar Fault. These structures include recumbent folds at Bradleys Creek, near Canberra, which pre-date the Upper Silurian volcanics of the Ngunawal Basin (Carson & Rickard 1998). An unconformity between the Llandoverian Ryrie Formation and the underlying Upper Ordovician Bendoc Group east of Cooma (Lewis *et al.* 1994) could also be of pre-late Benambran age. The most precise age constraint on the early east–west structures is at Delegate, where east–west folds pre-date the middle to upper Llandovery Tombong beds (Glen & VandenBerg 1987). This indicates that east–west structures in the Molong–Wyangala–Kuark Zone are pre-middle Llandovery (pre-435 Ma), are overprinted by the Yalmy Fault Zone, and therefore pre-date late Benambran deformation (Table 1).

INFLUX FROM THE 'BENAMBRAN HIGHLANDS'

In the southern part of the east Lachlan Fold Belt, 'Benambran deformation' is supposedly marked by an Early Silurian westward influx of continental detritus, which is presumably derived from the Wagga–Omeo Zone (Crook *et al.* 1973; Owen & Wyborn 1979). The influx is represented by the lower to middle Llandoverian Yalmy Group and equivalents (Lewis *et al.* 1994), which include the lower Llandoverian Tantangara Formation in the Tantangara–Brindabella region (Owen & Wyborn 1979), the Black Mountain Sandstone at Canberra and the Mundoonen Sandstone in the Yass region (Crook *et al.* 1973). Crook *et al.* (1973) described the facies change as an influx of terrigenous detritus derived as proximal submarine fans off the 'Benambran highlands'. Glen (in White & Chappell 1989 p. 26) described some very localised 'grits' from the Yalmy Group that contained abundant vein quartz, basalt and 'felted amphibolite fragments', which suggest a source other than the underlying low-grade turbidites.

Uplift associated with the 'Benambran highlands' is recorded as widespread, but discontinuous, Lower to Middle Silurian carbonate lenses in the central Lachlan Fold Belt, which suggests persistent shallow shelf conditions (VandenBerg 1998; Willman *et al.* 1999). These conditions include the Claire Creek and Pyles limestone lenses in the Limestone Creek graben (VandenBerg *et al.* 1998). In the east Lachlan Fold Belt, shedding of material from the uplifted block extends as far north as Yass (Mundoonen sandstone: Crook *et al.* 1973), but does not extend eastward across the Combienbar Fault (Lewis *et al.* 1994). The more distal eastward (Rylie Formation) and northward (Mundoonen Sandstone) turbidites appear to be the youngest, extending into the Wenlockian, which may indicate a diachronous shedding of material away from the 'Benambran highlands' to the southwest. Also the detritus does not extend northward to the Macquarie Volcanic arc, where the Upper Ordovician black shale sequence is faulted against mafic volcanic and volcaniclastic units of the Upper Ordovician Cabonne Group (Glen & Wyborn 1997).

This regional facies change at the Silurian–Ordovician boundary appears to relate to an event that pre-dates the late Llandoverian (*ca* 430 Ma) late Benambran deformation. To account for the burial of Ordovician rocks to mid-crustal levels prior to high-grade metamorphism in the Cooma Complex, Glen (1992) suggested that earliest Benambran thrusting probably began in the Late Ordovician, and suggested that local lenses of quartzite in the Gisbornian–Bolindian Bendoc Group may represent the initial detrital influx in response to that deformation. However, it has been demonstrated that sediments associated with this influx were deformed by late Benambran deformation in the Yalmy fold and thrust zone. This suggests that the 'Benambran' uplift was associated with an earlier (early Benambran) tectonic event.

HEAT FLOW ASSOCIATED WITH EARLY CRUSTAL THICKENING

An early (450–440 Ma) set of sporadically developed, but widespread, oblique folds and thrusts, which are overprinted by the late Benambran deformation, have been identified throughout the east Lachlan Fold Belt. These early structures exist on the New South Wales south coast, in the Wyangala–Kuark Zone, around the Wyangala Batholith, in the Tumut Zone, and probably farther west in the Girilambone Group. The 440 Ma northwest-trending folds in the Tabberabbera Zone (Foster *et al.* 1999) are coeval and may be part of this deformation.

Regional early Benambran convergence resulted in widespread crustal thickening in the central and southern parts of the east Lachlan Fold Belt. Given that measured stratigraphic thicknesses of the Ordovician turbidite units are generally much less than 5 km (VandenBerg & Stewart 1992), direct evidence for considerable crustal thickening is simply the existence of Middle Ordovician (or younger) sediments in the middle crust (10–15 km depth) of the Omeo Complex in the central Lachlan Fold Belt at *ca* 430 Ma (Keay *et al.* 1999). A similar argument can be made for the east Lachlan Fold Belt using the 430 Ma Cooma Complex, which contains Lower–Middle Ordovician sediments buried to 300–400 MPa.

Given that burial of Ordovician sediments must have occurred prior to the development of the migmatite complexes, and that these sediments are possible candidates for a source component of the S-type granites of the east and central Lachlan Fold Belt (Collins 1996, 1998), a crucial question is: can thickening of the crust at 450–440 Ma increase the temperature of the crust so as to induce melting of the Ordovician sediments after 10–20 million years? In particular, can such melting occur without the addition of an extra heat source (from the mantle) or is internal heat generation by radioactive decay of K, Th and U within the crust, plus a normal mantle thermal flux, sufficient to cause melting in the time limits imposed by the geological data? In what follows, thermal transport is assumed to be solely by conduction following Fourier's Law of heat conduction with internal heat generation. We use the mantle thermal conductivity estimate of 4.5 $Wm^{-1} K^{-1}$ from Stacey (1992).

A constraint on the thermal structure of the east Lachlan Fold Belt lithosphere at 430 Ma is given by the present-day thermal structure. In the following discussion, we develop a model of the present-day lithosphere beneath southeast Australia and then proceed to develop thermal models for the period 430 Ma. The refraction seismic data for the east Lachlan Fold Belt shows a marked jump in velocity at 22 km depth from 6.0 to 6.5 km s^{-1}. The lowest crust has velocities approximately 6.8 to 7.0 km s^{-1}. The Moho is marked by a jump in velocity from 7.0 to 8.0 km s^{-1} at a depth of 43–44 km (Finlayson *et al*. 1998).

An additional constraint on the present-day thermal structure of the east Lachlan Fold Belt lithosphere comes from seismic results (Muirhead & Drummond 1991) and SKIPPY results (van der Hilst *et al*. 1998), which define the lithosphere/asthenosphere boundary at 125 km. If we take the definition of this boundary to be 0.85 T_m, where T_m is the solidus temperature of the upper mantle (Pollack & Chapman 1977), the temperature for the base of the lithosphere at 125 km is 1500°C and at 100 km is 1430°C. The remaining constraint on the present-day thermal structure is given by the modern-day surface heat flow, which varies from 60 to 120 mWm^{-2} in eastern Australia (Cull 1982). However, much of the east Lachlan Fold Belt surface heat flux is between 60 and 80 mWm^{-2} with higher fluxes occurring further to the west and southwest.

The average K, U and Th contents of Ordovician K-bearing (micaceous) sediments, the Cooma granite, migmatites surrounding the Cooma granite, and of S- and I-granites of the east Lachlan Fold Belt are given in Table 2. In the Ordovician/Silurian, these heat-production rates would have been higher by approximately 8% and, where relevant, these higher internal heat-production rates are used in the calculations below. Table 2 also indicates that melting or partial melting of the Ordovician sediments, forming migmatites and Cooma granite (Munksgaard 1988), tended to partition heat-producing elements into the melt products, which implies that K, Th and U were concentrated in the upper crust during granite generation and emplacement.

Given the above constraints, a number of scenarios have been explored using a 1-D finite-difference code to calculate the steady-state lithospheric geotherm, the resultant surface heat flux and the time taken to reach steady state following a thermal perturbation within the lithosphere. These scenarios are: (i) top 22 km of the present crust composed of K-rich Ordovician sediment and associated metamorphic rocks of similar composition; lower 22 km of present crust composed of mafic granulite (probably of Neoproterozoic–Cambrian age); (ii) top 22 km of the present crust 50% K-rich Ordovician sediment, 50% quartz-rich Ordovician sediment and associated metamorphic rocks of similar composition; lower 22 km of present crust mafic granulite; (iii) top 10 km of present crust comprised mainly of granites with next 12 km composed of granulites depleted in radiogenic heat-production elements through melting; lower 22 km of present crust mafic granulite; (iv) top 5 km of present crust comprising 80% granite and 20% K-rich sediments, giving an internal heat-production rate of 3.0 μWm^{-3}; remainder of crust down to 22 km consisting of metamorphic rocks depleted exponentially in heat producing elements, and a lower 22 km of mafic granulite; D = 12 km (where D is the length for crustal heat production to decrease to 1/e of its maximum value and e is the base of the natural logarithms; and (v) same as in (iv) except D = 6 km.

The results of these scenarios are given in Tables 3 and 4 together with the parameters used in the calculations. It is clear that simple models of the heat-production rates for the east Lachlan Fold Belt give modern surface heat-flow rates (Q in Table 3) that are higher than the typical 60–80 μWm^{-2} for the east Lachlan Fold Belt, but are comparable with those recorded further to the west (Cull 1982). However, a model that assumes an exponential decrease in heat-production rate down to 22 km with the top 5 km

Table 2 Heat-production elements and heat-production rates for Ordovician sedimentary rocks and Silurian granites/metamorphic rocks.

Rock type	K_2O (%)	U (ppm)	Th (ppm)	Heat-production rate (μWm^{-3})
Ordovician sedimentary rocks	3.54	3.5	16.9	2.43
Cooma Granite	3.90	3.5	18.8	2.59
Cooma Migmatite	5.18	4.1	19.3	2.88
Average I-type granite	3.38	5.0	20.0	3.03
Average S-type granite	4.00	5.0	19.0	3.01

Table 3 Results of steady-state thermal calculations for five scenarios of crustal structure with lithosphere 125 km thick.

Quantity	Scenario 1	2	3	4	5
A0–10 km	2.43	1.22	3.0	Top 5 km—3 μWm^{-3}	
A11–22 km	2.43	1.22	1.5	Exponential drop-off to 22 km	
				D = 12 km	D = 6 km
A23–44 km	0.1	0.1	0.1	0.1	0.1
A45–100 km	0.0	0.0	0.0	0.0	0.0
T_{10}	316	254	276	274	242
T_{22}	510	435	475	449	408
T_{44}	794	741	769	751	722
T^a	8.7	9.4	9.0	9.3	9.6
Q_M	39.2	42.2	40.6	41.6	43.2
Q_C	57.7	29.0	48.0	28.4	17.8
Q_S	96.9	71.2	88.6	70.0	61.0

A is assumed internal heat-production rate in μWm^{-3}.
Q_S is the calculated surface heat flow in mWm^{-2}.
Q_C is the calculated contribution to surface heat flow from crustal sources in mWm^{-2}.
Q_M is the calculated thermal flux through the Moho in mWm^{-2}.
T_z is the calculated steady-state temperature in °C at the depth z in km.
T^a is the calculated steady-state temperature gradient from the Moho down to the base of the lithosphere in °C km^{-1}.
D is the characteristic length for crustal radioactive heat generation to decrease to 1/e of its maximum value where e is the base of natural logarithms.
Crustal thermal conductivity is 3.0 $Wm^{-1}K^{-1}$; thermal diffusivity is 10^{-6} m^2s^{-1}; mantle thermal conductivity is 4.5 $Wm^{-1}K^{-1}$; specific heat is 10^3 $Jkg^{-1}K^{-1}$.

composed of 80% granite plus 20% K-rich Ordovician sediments, and the lower 22 km of the crust composed of mafic granulites (option 4), gives a surface heat flow of 70 mWm^{-2} and a Moho temperature of 750°C. Such a heat flow is consistent with present-day observations (Cull 1982). In view of the differentiation indicated between S-type granite and Ordovician sediment in Table 2, we opt for a model in which the upper crust was differentiated by anatexis.

These models point to the following two conclusions: (i) modern heat-production elements in the east Lachlan Fold Belt are concentrated in the uppermost few kilometres; and (ii) the presence of K-rich granulites in the lower crust is very unlikely. Certainly, the lower crust is composed of material of very low heat-production rates: for instance, a felsic gneiss with 4% K_2O and zero concentrations of U and Th would have a heat production rate of 0.3 μWm^{-3}; a lower crust (22 km) composed of such material would add an extra 4.4 mWm^{-2} to the above values making the exponential model barely tenable.

We now proceed to develop a model for the pre-thickened Ordovician lithosphere, consisting initially of 5 km of typical K-rich Ordovician sediment resting on 15 km of mafic (Cambrian?) crust. With heat-production rates of 2.65 μWm^{-3} for the Ordovician rocks, 0.1 μWm^{-3} for the mafic granulites and the base of the lithosphere at 100 km (with a temperature of 1400°C), one arrives at an Ordovician Moho temperature of 395°C and a temperature at the base of the Ordovician pile of 109°C.

The crust is then thickened so that the Ordovician pile is 30 km thick (to conform with seismic refraction data plus ~8 km of post-Silurian erosion) and homogeneous in heat-production rate (2.65 μWm^{-3}), the mafic granulites are thickened to 22 km and the base of the lithosphere is at 132 km (temperature of 1520°C). Given this crustal architecture, the maximum temperature during peak metamorphism is 951°C at the Moho, which is reached 160 million years after thickening begins (assuming no erosion). The Moho reaches 820°C after approximately 60 million years. The peak temperature at the base of the Ordovician is 715°C after 160 million years, while the temperature is 650°C after 60 million years.

Table 4 Results of steady-state thermal calculations for five scenarios of crustal structure with lithosphere 105 km thick.

Quantity	Scenario 1	2	3	4	5
A0–10 km	2.43	1.22	3.0	Top 5 km—3 μWm^{-3}	
A11–22 km	2.43	1.22	1.5	Exponential drop-off to 22 km	
				D = 12 km	D = 6 km
A23–44 km	0.1	0.1	0.1	0.1	0.1
A45–100 km	0.0	0.0	0.0	0.0	0.0
T_{10}	315	259	294	277	249
T_{22}	557	486	520	497	461
T_{44}	856	809	832	816	792
T^a	10.25	11.1	10.7	11.0	11.4
Q_M	46.1	49.9	48.2	49.3	51.3
Q_C	57.7	29.0	48.0	28.4	17.8
Q_S	103.8	78.9	96.2	77.7	69.1

Parameters as for Table 3.

These calculations indicate that observed heat-production rates, together with reasonable estimates for the thickness of the lithosphere after early Benambran deformation, are insufficient to produce temperatures where the crust can melt within 10–20 million years of initial thickening. A thermal time-scale to reach steady state after 10–20 million years is associated with a length scale of approximately 20–25 km for conductive heating, whereas a thermal time-scale to reach steady state of 160 million years corresponds to a length scale of about half the Benambran lithospheric thickness. Thus, the extra heat needed to produce crustal melting within 10–20 million years of initial thickening must have been supplied by injecting a new heat source into the crust itself. The conclusion is that this additional heat source must have involved intrusion of mantle-derived melts into the crust. Such material itself must have been produced on a time-scale that did not involve conductive heating due to lithospheric thickening and a likely process is decompression melting of the mantle arising either from thickening of the crust followed by rapid erosion or more likely from rifting during formation of the Early–Middle Silurian basins, such as the Tumut Trough. These possibilities will be evaluated elsewhere.

Gray and Cull (1992) generated 1-D thermal models for the west Lachlan Fold Belt and indicated that shear heating could have provided significant heat to cause melting of the lower crust and generation of granitoids, provided that thermal equilibrium of the crust was attained. They concluded that the west Lachlan Fold Belt granitoids were a consequence of crustal thickening rather than the cause. However, this requires time intervals of ~30–50 million years following shortening, which cannot apply to the east and central Lachlan Fold Belt, where magmatism followed early Benambran deformation by 10–20 million years. They also modelled for generation of only 10% granite, but ~36% of the east Lachlan Fold Belt is represented by granite (Chappell *et al*. 1988) and the central Lachlan Fold Belt has a greater percentage, at present exposure levels (original depths equivalent to 200–400 MPa). Also, the shear-heating model implies that the granites are entirely crustally derived, but typical S-types have at least 10% mantle component and associated I-types may have up to 50% mantle component (Collins 1996, 1998).

SOURCE OF S-TYPE GRANITES

The results discussed above indicate that the heat necessary to cause partial melting of the Lachlan Fold Belt middle crust at 430 Ma, forming migmatitic complexes such as Cooma and Omeo, must ultimately be derived from the mantle, rather than solely by crustal overthickening. If these complexes represent the top of the melt zone of S-type granite generation (Collins 1996), then the ultimate source of heat for the S-type granites was also the mantle. The 500 MPa and ~800°C P/T estimates for phenocrysts inferred to be restite from the source of the S-type Hawkins Suite (Wyborn *et al*. 1981) suggest derivation of the S-type magmas from within the (present) upper ~20 km of the east Lachlan Fold Belt, just below the exposed migmatite complexes. Thus, the diatexites (or ‘dirty granites’) at Cooma and Omeo are likely to be representative of partially molten sediment that was incorporated into S-type granitic magmas (Collins 1996).

Collins (1996, 1998) has demonstrated that the S-type granites can be chemically modelled as I-type granites heavily contaminated by Ordovician sediment, which accords strongly with the Nd–Sr isotopic data for east Lachlan Fold Belt granitoids (Keay *et al*. 1997). Particularly convincing are the Early to Middle Ordovician ages of inherited zircon in Lachlan Fold Belt S-type granites (Anderson *et al*. 1998; Keay *et al*. 1999), which are within error of the detrital zircon populations in high-grade rocks of the migmatitic complexes (Williams 1995; Keay *et al*. 1999). The combined field, petrological, geochemical, seismic and heat-flow data are compelling evidence that the S-type batholiths of the east and central Lachlan Fold Belt contain a major component of Ordovician sediment (see also Collins 1999a).

The thermal model outlined above which best accords with present heat flow in the east Lachlan Fold Belt requires an exponential decrease in radiogenic components through the top 22 km of the crust and a mafic granulite lower crust with virtually zero intrinsic heat output. Therefore, the heat-flow data render it unlikely that fragments of Proterozoic 'continental crust' material exist in the lower crust of the east Lachlan Fold Belt, such as the hypothetical 'Kosciuszko basement terrane' (Chappell *et al*. 1988). For the same reasons, the speculated Wagga basement terrane may not exist beneath the central Lachlan Fold Belt. The estimated ~20 km depth of generation of the S-type magmas is consistent with our preferred heat-flow model, which invokes an exponential decrease in heat-producing elements with depth, caused by mid-crustal anatexis of Ordovician sediment.

TWO MAGMATIC ARCS AND TWO SUBDUCTION COMPLEXES IN THE EARLY SILURIAN

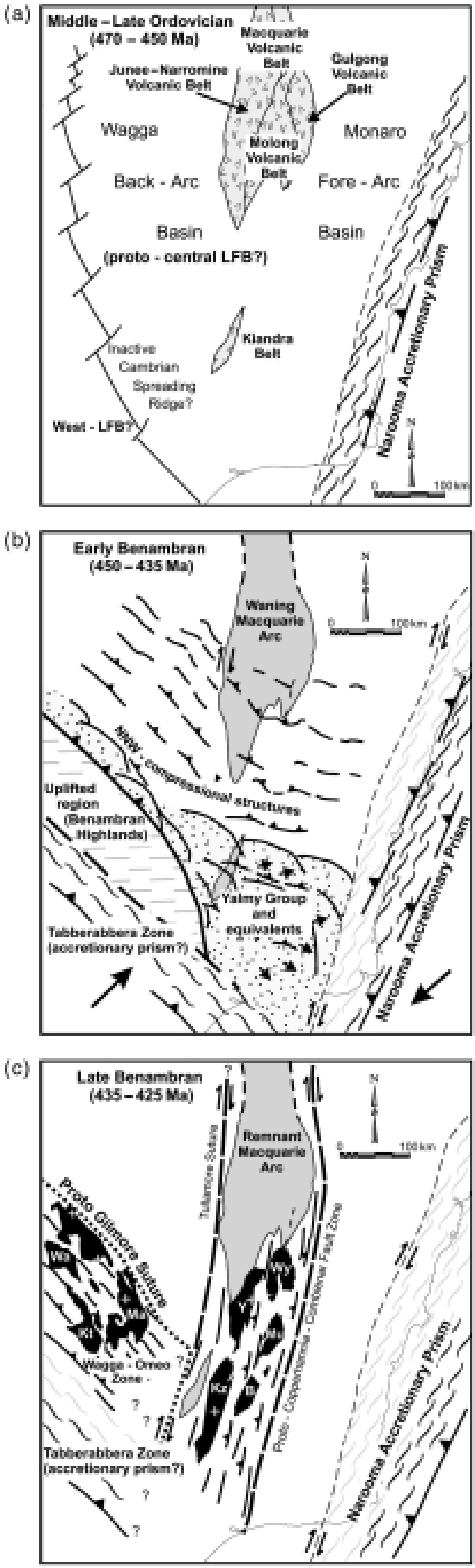
A major conclusion from the thermal modelling is that heat from crustal thickening alone is insufficient to cause widespread mid-crustal melting and that a mantle heat source is required. Evidence for contemporaneous mantle activity during S-type magma generation is best illustrated by the extensive *ca* 430 Ma Micalong Swamp Igneous Complex in the Young Batholith (Owen & Wyborn 1979), but most of the east Lachlan Fold Belt S-type batholiths contain coeval gabbroic complexes. The geochemistry of primitive basalts from the Micalong Swamp Igneous Complex closely resembles those from parts of the early Lau Basin (Collins 1999b), which implies an extensional arc or backarc environment for S-type magma generation. That the S-type magmatic belt developed over and along strike from the pre-existing Ordovician intraoceanic Macquarie Volcanic Arc (Glen *et al*. 1998) implies that the change from mafic to silicic volcanism in the east Lachlan Fold Belt simply involved burial of the arc by overthrusting of Ordovician sediments, rather than a change of tectonic setting (Figure 4).

The Middle–Late Ordovician tectonic setting of the east Lachlan Fold Belt is commonly regarded as a west-dipping intraoceanic arc (Macquarie Volcanic Arc), with a wide backarc basin (Wagga Marginal Basin) to the west and a forearc (Monaro Forearc) and accretionary prism (Narooma) complex to the east (Figure 4a) (Scheibner 1973; Powell 1983; Glen *et al*. 1998). However, the subsequent evolution during the Silurian is controversial. Based on the above considerations and the synthesis presented below, we support a double convergent subduction model for the east and central Lachlan Fold Belt, similar to that initially proposed by Collins and Vernon (1992 figure 3d), and subsequently by Gray *et al*. (1997) and Soesoo *et al*. (1997).

The consensus tectonic model for Benambran deformation in the east and central Lachlan Fold Belt has been one of dextral transpression during plate convergence (Powell 1983; Fergusson & Coney 1992), possibly associated with wholesale southward tectonic transport of the central Lachlan Fold Belt toward the east Lachlan Fold Belt (Morand & Gray 1991; Glen 1992). Fergusson (1998a) considered the deformation as 'intraplate', and Glen *et al*. (1992) suggested that the Melbourne Zone (west Lachlan Fold Belt) was thrust northward against the Tabberabbera Zone in the Late Silurian – Early Devonian.

None of these models explains the existence of two 430 Ma 'magmatic belts' of S-type batholiths in the east and central Lachlan Fold Belt, with structural trends and batholith orientations up to 30–40° oblique to one another (Figure 1). These two belts constitute the majority of granitoids in the Lachlan Fold Belt (Chappell *et al*. 1991a) and are comparable to abundances in subduction-related magmatic arcs, such as the Mesozoic Cordilleran batholiths of western North and South America. However, these batholiths usually lack S-type granites, a feature that has been used by Chappell (1984) and coworkers to argue against subduction in the Lachlan Fold Belt. Nonetheless, S-type granites do exist in the US Cordillera (Todd & Shaw 1985; Ague & Brimhall 1987), despite assertions to the contrary (White *et al*. 1986). The major difference is the amount of sediment contamination in the magmas. In the Lachlan Fold Belt, S-type granites are heavily contaminated I-types (Collins 1996, 1998; Keay *et al*. 1997), whereas most of the Cordilleran granites are direct derivatives from meta-igneous (I-type) sources (Gromet & Silver 1987; Atherton 1990; Tate *et al*. 1999), with minor sediment contamination (Ague & Brimhall 1987). This compositional difference largely reflects granite generation either within pre-arc crust thickened dominantly by igneous underplating (US Cordillera and Andes) or within pre-arc crust thickened by tectonic processes (east Lachlan Fold Belt). This explanation removes a major criticism of subduction-related models of the Lachlan Fold Belt (Chappell 1984; O'Halloran & Bryan 1998), which implied that 'calc-alkaline' rocks are the only magmatic type produced at active continental margins. The presence of mafic magmas associated with S-type granites, such as the Micalong Swamp Igneous Complex in the central Lachlan Fold Belt, and the Devonian mafic complexes in the Tabberabbera Zone, each with magmatic-arc affinities (Collins 1999b; Soesoo & Nicholls 1999), confirms that the Lachlan Fold Belt granitoid magmatism was subduction related.

The two oblique 430 Ma magmatic belts are juxtaposed against turbidites, each separated by an 'I–S' line (Hendrickx 1999), with the adjacent sedimentary packages inferred to be remnants of accretionary complexes (Figure 4c). The Narooma complex and equivalents along



the New South Wales south coast are inferred to be part of a west-dipping, north-northeast-verging, east-migrating Late Ordovician accretionary prism (Miller & Gray 1996; Offler *et al*. 1998), which is subparallel to the trend of the east Lachlan Fold Belt S-type batholiths and the I–S line of White *et al*. (1976). Similarly, the Tabberabbera Zone was part of an asymmetric northeast-dipping, southwest-verging, south-migrating zone of deformation, which has also been interpreted as an accretionary prism (Gray *et al*. 1997; Soesoo *et al*. 1997; Gray & Foster 1998). The presence of blueschist facies metamorphic rocks in the southernmost leading fault is strong support for a subduction–accretion complex for the zone (Spaggiari *et al*. 1999). The Tabberabbera Zone is separated from the Wagga–Omeo Zone by a 'western' I–S line marked by the Kiewa–Kancoona Fault System (Hendrickx 1999). Collins (1998) suggested that the original, or 'eastern', I–S line represents the boundary along which Ordovician sediments were underthrust to sufficient depths (~20 km) to be involved in crustal melting. This is consistent with a subduction model involving thickening of Ordovician sediments in two simultaneously developing, but independent, accretionary prisms during the latest Ordovician – Early Silurian.

In this tectonic context, it is considered that regional early Benambran northwest–southeast-trending folds and thrusts were associated with crustal thickening during the early stages of convergent-margin tectonism in the central Lachlan Fold Belt, between 450 and 440 Ma (Figure 4b). This northeast–southwest shortening uplifted the central Lachlan Fold Belt to produce the inferred widespread early Llandovery carbonate shelf (Benambran highlands) of that region (VandenBerg 1998), with consequent shedding of detritus into the east Lachlan Fold Belt producing the Yalmy Group and equivalents (Crook *et al*. 1973); these are concentrated within the southern part of the east Lachlan Fold Belt. Deformation occurred before the widespread high-grade metamorphism and S-type plutonism at 430 Ma (Owen & Wyborn 1979; VandenBerg 1998). Based on these considerations, this early Benambran deformation may have initiated northeast-directed subduction beneath the central Lachlan Fold Belt, as no evidence exists for an earlier, northwest-trending magmatic arc in that region. Location of the subduction zone was possibly controlled by the Cambrian(?) spreading centre (Figure 4a) that produced the backarc basin in which Lower Ordovician turbidites of the Girilambone Group and central Lachlan Fold Belt were deposited (the Wagga Marginal Basin of Powell 1983).

Figure 4 Inferred Late Ordovician to Early Silurian tectonic evolution of the east and central Lachlan Fold Belt (LFB). (a) Pre-deformation, intraoceanic arc arrangement of major tectonic elements. Note the inferred east-facing extensional arc associated with west-dipping subduction. (b) Early Benambran northeast–southwest-directed shortening associated with initiation of Tabberabbera Zone subduction, causing oblique folds and thrusts in east Lachlan Fold Belt, uplift of central Lachlan Fold Belt (Benambran highlands), and sediment dispersal to the north and east. (c) Late Benambran deformation in the east and central Lachlan Fold Belt: two S-type magmatic arcs associated with regional shortening and two opposed accretionary complexes. Original relative disposition of each belt unknown, but possibly located several hundred kilometres apart.

The northern limit of early Benambran oblique structures in the east Lachlan Fold Belt broadly coincides with the northern extent of S-type granites, which can be traced northward through the east Lachlan Fold Belt into the Wyangala Batholith (Figure 4c). No other batholith shows a north–south variation in S- and I-type granites, and the change is coincident with the northern limit of observed east–west-trending folds and thrusts. The most northerly recognised regional-scale east–west fault separates Ordovician turbidites from the coeval Macquarie Volcanic Arc to the north (Glen & Wyborn 1997). Therefore, the generation of S-type granites, intensity of east–west deformation, and degree of uplift, all decrease toward the north in the east Lachlan Fold Belt, suggesting a causal relationship between all features. The coincidence of S-type granites with areas of 'early Benambran' crustal thickening suggests that it was a prerequisite for S-type magma generation.

The eastern limit of early Benambran oblique structures in the east Lachlan Fold Belt broadly coincides with the original or 'eastern' I–S line, which is also the eastern limit of the Early Silurian magmatic arc. East of the Wyangala Batholith, the boundary is the Copperhannia Thrust (Paterson *et al.* 1990 p. 648), but farther south, the I–S line progressively jumps westward, crossing upper crustal structural trends to coincide with the Yalmy Fault (Glen & VandenBerg 1987), where it corresponds to the boundary between thin- and thick-skinned late Benambran deformation. However, the eastern limit of provable early Benambran structures exists at least 20 km farther eastward, and must be beyond the eastern limit of the 'Benambran influx' of Early Silurian (Yalmy Group) detritus, located near Delegate (Figure 1). The Yalmy Group extends at least 90 km northward from Delegate in fault-bound slivers along the western margin of the Bega Batholith (Lewis *et al.* 1994), coinciding with the southern extension of the Narongo Fault. The Narongo Fault in turn links northward into the Lake George Fault, which links with the Copperhannia Thrust (Figure 1). Therefore, a major tectonic boundary can be established that coincides with the eastern limit of Benambran highlands influx, but it is not the I–S line.

Tracing the boundary farther south into Victoria is problematic. The southern extension of the Narongo fault system can be interpreted as a series of north- to northeast-trending, anastomosing faults broadly coinciding with the western margin of the Bega Batholith. Southward in Victoria, the major structural break and geophysical anomaly is the Combienbar Fault (Simpson *et al.* 1997), which separates the Buchan and Mallacoota Zones (Gray 1988). Simpson *et al.* (1997) considered that the Combienbar Fault might link with the northeast-trending Burragate Fault in southern New South Wales, but our mapping shows that the latter swings southward into the Cann Valley Shear Zone, some 10 km eastward of the Combienbar Fault. The Burragate is also much younger, mainly of Devonian age (Simpson *et al.* 1997). We therefore consider that the major tectonic break coincides with the Combienbar Fault in the south and the Copperhannia Thrust in the north, and term this structure the Copperhannia–Combienbar Fault System. It represents the eastern boundary of the Molong–Wyangala–Jerangle–Kuark Zone of Scheibner and Basden (1996) and appears to separate the Early Silurian arc from forearc and accretionary prism rocks farther east.

Given the absence of Yalmy Group detritus and lack of oblique early Benambran structures eastward of the Copperhannia–Combienbar Fault System, it appears that considerable translation may have occurred along this boundary. It is possible that strain was partitioned along the boundary during regional dextral transpression associated with late Benambran deformation (Powell 1983) (Figure 4b), or it may have developed during south to south-southeast thrusting of the central Lachlan Fold Belt over the eastern Lachlan Fold Belt during the Late Silurian – Early Devonian (Morand & Gray 1991). The whole fault system was reactivated in the Carboniferous (Paterson *et al.* 1990; Simpson *et al.* 1997).

CONCLUSIONS

Early (latest Ordovician – earliest Silurian) and late (late Llandovery) Benambran events, which we consider relate to two separate, but coeval, subduction–accretion/magmatic arc complexes, have been identified in the east and central Lachlan Fold Belt. The two events involved deformation along two active continental margins and facilitated burial and melting of the Ordovician sedimentary rocks to produce the migmatitic complexes and S-type granite batholiths of the region. The north-northwest-trending magmatic arc of the central Lachlan Fold Belt developed at *ca* 430 Ma following northeast-directed subduction that probably began ~20 million years earlier during early Benambran deformation. In contrast, the north-trending magmatic arc of the east Lachlan Fold Belt developed over a pre-existing Ordovician intraoceanic arc associated with west-dipping subduction during late Benambran deformation. Initiation of the central Lachlan Fold Belt subduction system in the Late Ordovician is considered responsible for the widespread underthrusting of Ordovician sediment, and was associated with regional uplift of the central Lachlan Fold Belt (Benambran highlands), and with dispersal of quartzose Llandoverian sediments across the southern part of the east Lachlan Fold Belt.

The two Late Ordovician accretionary prisms, each with opposing thrust vergence, support the subduction-related model for S-type magma genesis, and the I–S lines of both provinces generally reflect the boundary between arc and prism. The two convergent margins developed on either side of a coherent structural block comprising the east and central Lachlan Fold Belt. Correlation of the Lower Ordovician turbidites between the two provinces (VandenBerg & Stewart 1992; Glen 1992) and similarity of detrital zircon ages in the Ordovician sediments from the Cooma (Williams 1995) and Omeo (Keay *et al.* 1999) migmatitic complexes strongly imply that the overriding plate of both subduction systems was originally contiguous. Accordingly, two magmatic arcs operated simultaneously as part of a double convergent subduction system.

ACKNOWLEDGEMENTS

This work was supported by ARC Large Grant A39702018. Simon Richards, Mark Pawley and a host of third-year undergraduates helped map the Cooma Complex. Sue Keay and Dave Gray are thanked for pre-print access, Dick Glen, Mike Rickard, John Walshe and Alison Ord for comments, and Steven Dick for drafting. We also acknowledge the helpful reviews of Chris Fergusson and Mike Sandiford. BEH publishes with the permission of the Director of the AGCRC.

REFERENCES

AGUE J. J. & BRIMHALL G. H. 1987. Granites of the batholiths of California: products of local assimilation and regional-scale crustal contamination. *Geology* **15**, 63–66.

ANDERSON J. A. C., PRICE R. C. & FLEMING P. D. 1996. Structural analysis of metasedimentary enclaves: implications for tectonic evolution and granite petrogenesis in the southern Lachlan Fold Belt, Australia. *Geology* **26**, 119–122.

ANDERSON J. A. C., WILLIAMS I. S., PRICE R. C. & FLEMING P. D. 1998. U–Pb zircon ages from the Koetong adamellite: implications for granite genesis and the local basement in NE Victoria. *Geological Society of Australia Abstracts* **42**, 1–2.

ATHERTON M. P. 1990. The Coastal Batholith of Peru: the product of rapid recycling of 'new' crust formed within rifted continental margin. *Geological Journal* **25**, 337–349.

CARR G. R., DEAN J. A., SUPPEL D. W. & HEITHERSAY P. S. 1995. Precise lead isotopic fingerprinting of hydrothermal activity associated with Ordovician to Carboniferous metallogenic events in the Lachlan Fold Belt of New South Wales. *Economic Geology* **90**, 1467–1505.

CARSON L. J. & RICKARD M. J. 1998. Early recumbent fold hinge in the Bradleys Creek Metamorphic complex, near Canberra, Australia. *Australian Journal of Earth Sciences* **45**, 501–508.

CHAPPELL B. W. 1984. Source rocks of S- and I-type granites in the Lachlan Fold Belt, southeastern Australia. *Philosophical Transactions of the Royal Society of London* **A310**, 693–707.

CHAPPELL B. W., ENGLISH P. M., KING P. L., WHITE A. J. R. & WYBORN D. 1991a. *Granites and Related Rocks of the Lachlan Fold Belt. 1:1 250 000 Geological Map*. Bureau of Mineral Resources, Canberra.

CHAPPELL B. W., WHITE A. J. R. & HINE R. 1988. Granite provinces and basement terranes in the Lachlan Fold Belt, southeastern Australia. *Australian Journal of Earth Sciences* **35**, 505–521.

CHAPPELL B. W., WHITE A. J. R. & WILLIAMS I. S. 1991b. A transverse section through granites of the Lachlan Fold Belt: Second Hutton Symposium Excursion Guide. *Bureau of Mineral Resources Record* **1991/22**.

COLLINS W. J. 1996. S- and I-type granitoids of the eastern Lachlan fold belt: products of three-component mixing. *Transactions of the Royal Society of Edinburgh: Earth Sciences* **88**, 171–179.

COLLINS W. J. 1998. An evaluation of petrogenetic models for Lachlan Fold Belt granitoids: implications for crustal architecture and tectonic models. *Australian Journal of Earth Sciences* **45**, 483–500.

COLLINS W. J. 1999a. Reply. An evaluation of petrogenetic models for Lachlan Fold Belt granitoids: implications for crustal architecture and tectonic models. *Australian Journal of Earth Sciences* **46**, 831–836.

COLLINS W. J. 1999b. Heat sources and tectonic models for the Lachlan Fold Belt: the basalt story. *Geological Society of Australia Abstracts* **53**, 33–34.

COLLINS W. J. & VERNON R. H. 1992. Palaeozoic arc growth, deformation and migration across the Lachlan Fold Belt, southeastern Australia. *Tectonophysics* **214**, 381–400.

CROOK K. A. W., BEIN J., HUGHES R. J. & SCOTT P. A. 1973. Ordovician and Silurian history of the southeastern part of the Lachlan Geosyncline. *Journal of the Geological Society of Australia* **20**, 113–138.

CULL J. P. 1982. An appraisal of Australian heat-flow data. *BMR Journal of Australian Geology & Geophysics* **7**, 11–21.

ELLIS D. J. & OBATA M. 1992. Migmatite and melt segregation at Cooma, New South Wales. *Transactions of the Royal Society of Edinburgh: Earth Sciences* **83**, 95–106.

FAGAN R. K. 1979. Deformation, metamorphism and anatexis of an Early Palaeozoic flysch sequence in northeastern Victoria. PhD thesis, University of New England, Armidale (unpubl.).

FERGUSSON C. L. 1987. Early Paleozoic back-arc deformation in the Lachlan Fold Belt, southeastern Australia; implications for terrane translations in eastern Gondwanaland. *American Geophysical Union Geodynamic Series* **19**, 39–56.

FERGUSSON C. L. 1998a. Cambrian–Silurian oceanic rocks, upper Howqua River, eastern Victoria: tectonic implications. *Australian Journal of Earth Sciences* **45**, 633–644.

FERGUSSON C. L. 1998b. Thick-skinned folding in the eastern Lachlan Fold Belt, Shoalhaven Gorge, New South Wales. *Australian Journal of Earth Sciences* **45**, 677–686.

FERGUSSON C. L. & CONEY P. J. 1992. Convergence and intraplate deformation in the Lachlan Fold Belt of southeastern Australia. *In:* Fergusson C. L. & Glen R. A. eds. *The Palaeozoic eastern margin of Gondwanaland: Tectonics of the Lachlan Fold Belt, southeastern Australia, and related orogens*, pp. 417–439. Tectonophysics **214**.

FERGUSSON C. L. & VANDENBERG A. H. M. 1990. Middle Palaeozoic thrusting in the eastern Lachlan Fold Belt, southeastern Australia. *Journal of Structural Geology* **12**, 577–589.

FINLAYSON D. M., LEVEN J. H., JOHNSTONE D. W., KORSCH R. J. & GLEN R. A. 1998. Crustal architecture in the eastern Lachlan Orogen from deep seismic profiling at wide-angle and near-vertical incidence. *Geological Society of Australia Abstracts* **49**, 144.

FOSTER D. A., GRAY D. R. & BUCHER M. B. 1999. Chronology of deformation within the turbidite-dominated, Lachlan Orogen: implications for the tectonic evolution of eastern Australia and Gondwana. *Tectonics* **18**, 452–485.

FOSTER D. A., GRAY D. R., KWAK T. A. P. & BUCHER M. B. 1998. Chronologic and orogenic framework of turbidite-hosted gold deposits in the western Lachlan Fold Belt, Victoria. *Ore Geology Reviews* **13**, 229–250.

GLEN R. A. 1992. Thrust, extensional and strike-slip tectonics in an evolving Palaeozoic orogen—a structural synthesis of the Lachlan Orogen of southeastern Australia. *In:* Fergusson C. L. & Glen R. A. eds. *The Palaeozoic eastern margin of Gondwanaland: Tectonics of the Lachlan Fold Belt, southeastern Australia, and related orogens*, pp. 341–380. Tectonophysics **214**.

GLEN R. A., SCHEIBNER E. & VANDENBERG A. H. M. 1992. Paleozoic intraplate escape tectonics in Gondwanaland and major strike-slip duplication in the Lachlan orogen of southeastern Australia. *Geology* **20**, 795–798.

GLEN R. A. & VANDENBERG A. H. M. 1987. Thin-skinned tectonics in part of the Lachlan Fold Belt near Delegate, southeastern Australia. *Geology* **15**, 1070–1073.

GLEN R. A., WALSHE J. L., BARRON L. M. & WATKINS J. J. 1998. Ordovician convergent-margin volcanism and tectonism in the Lachlan sector of east Gondwana. *Geology* **26**, 751–754.

GLEN R. A. & WYBORN D. 1997. Inferred thrust imbrication, deformation gradients and the Lachlan Transverse Zone in the Eastern Belt of the Lachlan Orogen, New South Wales. *Australian Journal of Earth Sciences* **44**, 49–68.

GRANATH J. W. 1980. Strain, metamorphism, and the development of differentiated crenulation cleavages at Cooma, Australia. *Journal of Geology* **88**, 589–601.

GRAY C. M. 1984. An isotopic mixing model for the origin of granitic rocks in southeastern Australia. *Earth and Planetary Science Letters* **70**, 47–60.

GRAY C. M. 1990. A strontium isotopic traverse across the granitic rocks of southeastern Australia: petrogenetic and tectonic implications. *Australian Journal of Earth Sciences* **37**, 331–349.

GRAY D. R. 1988. Structure and tectonics. *In:* Douglas J. G. & Ferguson J. A. eds. *Geology of Victoria* (2nd edition), pp. 1–36. Geological Society of Australia, Victorian Division, Melbourne.

GRAY D. R. & CULL J. P. 1992. Thermal regimes, anatexis, and orogenesis: relations in the western Lachlan Fold Belt, southeastern Australia. *Tectonophysics* **214**, 441–461.

GRAY D. R. & FOSTER D. A. 1998. Character and kinematics of faults within the turbidite-dominated Lachlan Orogen: implications for tectonic evolution of eastern Australia. *Journal of Structural Geology* **20**, 1691–1720.

GRAY D. R., FOSTER D. A. & BUCHER M. B. 1997. Recognition and definition of orogenic events in the Lachlan Fold Belt. *Australian Journal of Earth Sciences* **44**, 489–501.

GROMET L. P. & SILVER L. T. 1987. REE variations across the Peninsular Ranges Batholith: implications for batholithic petrogenesis and crustal growth in magmatic arcs. *Journal of Petrology* **28**, 75–125.

HENDRICKX M. 1999. The I–S line: a thrust sheet boundary in eastern Victoria. *Geological Society of Australia Abstracts* **53**, 101–102.

HOBBS B. E. 1965. Structural analysis of the rocks between the Wyangala Batholith and the Copperhannia Thrust. *Journal of the Geological Society of Australia* **12**, 1–24.

HOPWOOD T. P. 1976. Stratigraphy and structural summary of the Cooma metamorphic complex. *Journal of the Geological Society of Australia* **23**, 345–360.

IRELAND T. R. & GIBSON G. M. 1998. SHRIMP monazite and zircon geochronology of high-grade metamorphism in New Zealand. *Journal of Metamorphic Geology* **16**, 149–167.

ISAACS D. R. L., BLEVIN P. & ARMSTRONG R. A. 2000. Geochronological re-evaluation of the Nymagee 1:250 000 sheet. *Geological Society of Australia Abstracts* **59**, 247.

IWATA K., SCHMIDT B. L., LEITCH E. C., ALLAN A. D. & WATANABE T. 1995. Ordovician microfossils from the Ballast Formation (Girilambone Group) of New South Wales. *Australian Journal of Earth Sciences* **42**, 371–376.

JOHNSON S. & VERNON R. H. 1995. Stepping stones and pitfalls in the determination of an anticlockwise P–T–t-deformation path: the low-P, high-T Cooma Complex, Australia. *Journal of Metamorphic Geology* **13**, 165–183.

KEAY S. M., COLLINS W. J. & MCCULLOCH M. T. 1997. A three-component mixing model for granitoid genesis: Lachlan Fold Belt, eastern Australia. *Geology* **25**, 307–310.

KEAY S. M., STEELE D. & COMPSTON W. 1999. Identifying granite sources by SHRIMP U–Pb zircon geochronology: an application to the Lachlan fold belt. *Contributions to Mineralogy and Petrology* **137**, 323–341.

LAWRIE K. C., MERNAGH T., BLACK L. P. & WYBORN D. 1998. Au–Cu mineralization at the Bald Hill Prospect, Lachlan Fold Belt, New South Wales. *Geological Society of Australia Abstracts* **49**, 262.

LEWIS P. C., GLEN R. A., PRATT G. W. & CLARKE I. 1994. *Bega–Mallacoota 1:250 000 Geological Sheet SJ/55-4, Explanatory Notes*. Geological Survey of New South Wales, Sydney.

MAAS R., NICHOLLS I. A., GREIG A. & NEMCHIN A. A. 1998. U–Pb geochronology of metasedimentary enclaves from the S-type Deddick granodiorite, southern Kosciuszko Batholith, eastern Victoria. *Geological Society of Australia Abstracts* **49**, 286.

MCCULLOCH M. T. & CHAPPELL B. W. 1982. Nd isotopic characteristics of S- and I-type granites. *Earth and Planetary Science Letters* **58**, 51–64.

MCCULLOCH M. T. & WOODHEAD J. D. 1993. Lead isotopic evidence for deep crustal-scale fluid transport during granite petrogenesis. *Geochimica et Cosmochimica Acta* **57**, 659–674.

MILLER J. M. & GRAY D. R. 1996. Structural signature of sediment subduction-accretion in a Palaeozoic accretionary complex, southeastern Australia. *Journal of Structural Geology* **18**, 1245–1258.

MORAND V. J. 1987. Structure of the Abercrombie Beds south of Reids Flat, New South Wales. *Australian Journal of Earth Sciences* **34**, 119–134.

MORAND V. J. 1988. Omeo Zone. *In:* Douglas J. G. & Ferguson J. A. eds. *Geology of Victoria* (2nd edition), pp. 18–20. Geological Society of Australia, Victorian Division, Melbourne.

MORAND V. J. 1990. Low-pressure regional metamorphism in the Omeo Metamorphic Complex, Victoria, Australia. *Journal of Metamorphic Geology* **8**, 1–12.

MORAND V. J. & GRAY D. R. 1991. Major fault zones related to the Omeo Metamorphic Complex, northeastern Victoria. *Australian Journal of Earth Sciences* **38**, 203–221.

MUIRHEAD K. J. & DRUMMOND B. J. 1991. The base of the lithosphere under Australia. *In:* Drummond B. J. ed. *The Australian Lithosphere*, pp. 23–40. Geological Society of Australia Special Publication **17**.

MUNKSGAARD N. C. 1988. Source of the Cooma Granodiorite, New South Wales—a possible role of fluid–rock interactions. *Australian Journal of Earth Sciences* **35**, 263–378.

OFFLER R., MILLER J. M., GRAY D. R., FOSTER D. A. & BALE R. 1998. Crystallinity and b_0 spacing in a Paleozoic accretionary complex, eastern Australia: Metamorphism, paleogeotherms, and structural style of an underplated sequence. *Journal of Geology* **106**, 495–509.

O'HALLORAN G. J. & BRYAN S. E. 1998. Comment. Divergent double subduction: tectonic and petrologic consequences. *Geology* **26**, 1051.

OVERSBY B. 1971. Palaeozoic plate tectonics in the southern Tasman Geosyncline. *Nature* **234**, 45–48.

OWEN M. & WYBORN D. 1979. Geology and geochemistry of the Tantangara and Brindabella area. *Bureau of Mineral Resources Bulletin* **204**.

PATERSON S. R., TOBISCH O. T. & MORAND V. J. 1990. The influence of large ductile shear zones on the emplacement and deformation of the Wyangala Batholith, SE Australia. *Journal of Structural Geology* **12**, 639–650.

POGSON D. J. & WATKINS J. J. 1998. *Bathurst 1:250 000 Geological Sheet SI/55-8, Explanatory Notes*. Geological Survey of New South Wales, Sydney.

POLLACK H. N. & CHAPMAN D. S. 1977. On the regional variation of heat flow, geotherms and lithospheric thickness. *Tectonophysics* **38**, 279–396.

POWELL C. MCA. 1983. Tectonic relationship between the Late Ordovician and Late Silurian palaeogeographies of southeastern Australia. *Journal of the Geological Society of Australia* **30**, 353–373.

RODDICK J. C. & COMPSTON W. 1976. Radiometric evidence for the age of emplacement and cooling of the Murrumbidgee Batholith. *Journal of the Geological Society of Australia* **23**, 223–233.

SCHEIBNER E. 1973. A plate tectonic model of the Palaeozoic tectonic history of New South Wales. *Journal of the Geological Society of Australia* **20**, 405–426.

SCHEIBNER E. & BASDEN H. (Editors) 1996. Geology of New South Wales—Synthesis. Vol. 2, Geological Evolution. *Geological Survey New South Wales Memoir* **13(2)**.

SHERWIN L. 1996. Narromine *1:250 000 Geological Sheet SI/55-3, Explanatory Notes*. Geological Survey of New South Wales, Sydney.

SIMPSON C. J., FERGUSSON C. L. & ORANSKAIA A. 1997. Craigie 1:100 000 map area report. *Geological Survey of Victoria Report* **111**.

SOESOO A., BONS P. D., GRAY D. R. & FOSTER D. A. 1997. Divergent double subduction: petrologic and tectonic consequences. *Geology* **25**, 755–758.

SOESOO A. & NICHOLLS I. A. 1999. Mafic rocks and spatially associated Devonian felsic intrusions of the southern Lachlan fold Belt: a possible mantle contribution to crustal evolution processes. *Australian Journal of Earth Sciences* **46**, 725–734.

SPAGGIARI C. V., GRAY D. R. & FOSTER D. A. 1999. Occurrences and significance of Franciscan-like melange and blueschist metamorphism in Lachlan orogen fault zones. *Geological Society of Australia Abstracts* **53**, 248–249.

STACEY F. D. 1992. *Physics of the Earth*. Brookfield Press, Brisbane.

STUART-SMITH P. G., HILL R. I., RICKARD M. J. & ETHERIDGE M. A. 1992. The stratigraphy and deformation history of the Tumut region: implications for the development of the Lachlan fold belt. *Tectonophysics* **214**, 211–238.

TALENT J. A. 1965. The stratigraphic and diastrophic evolution of central and eastern Victoria in Middle Palaeozoic times. *Proceedings of the Royal Society of Victoria* **79**, 179–195.

TATE M. C., NORMAN M. D., JOHNSON S. E., FANNING C. M. & ANDERSON J. L. 1999. Generation of tonalite and trondhjemite by subvolcanic fractionation and partial melting in the Zarza Intrusive Complex, western Peninsula Ranges Batholith, northwestern Mexico. *Journal of Petrology* **40**, 983–1010.

TODD V. R. & SHAW S. E. 1985. S-type granitoids and an I–S line in the Peninsula Ranges batholith, southern California. *Geology* **13**, 231–233.

TUCKER R. D. & MCKERROW W. S. 1995. Early Paleozoic chronology: a review in light of new U–Pb zircon ages from Newfoundland and Britain. *Canadian Journal of Earth Sciences* **32**, 368–379.

VandenBerg A. H. M. 1998. The Benambran deformation, eastern Lachlan Fold Belt: some observations of a complex event. *Geological Society of Australia Abstracts* **49**, 449.

VandenBerg A. H. M. 1999. Timing of orogenic events in the Lachlan Orogen. *Australian Journal of Earth Sciences* **46**, 691–701.

VandenBerg A. H. M. & Stewart I. R. 1992. Ordovician terranes of the southeastern Lachlan Fold Belt: stratigraphy, structure and palaeogeographic reconstruction. *In:* Fergusson C. L. & Glen R. A. eds. *The Palaeozoic eastern margin of Gondwanaland: Tectonics of the Lachlan Fold Belt, southeastern Australia, and related orogens*, pp. 159–176. Tectonophysics **214**.

VandenBerg A. H. M., Hendrickx M. A., Willman C. E., Magart A. P. M., Simons B. A. & Ryan S. M. 1998. Benambra 1:100 000 map area report. *Geological Survey of Victoria Report* **114**.

van der Hilst R. D., Kennett B. L. N. & Shibutani T. 1998. Upper mantle structure beneath Australia from portable array deployments. *In:* Braun J., Dooley J., Goleby B., van der Hilst R. & Klootwijk C. eds. *Structure and Evolution of the Australian Continent*, pp. 39–57. American Geophysical Union Geodynamic Series **26**.

White A. J. R. & Chappell B. W. 1989. *Geology of the Numbla 1:100 000 sheet 8624*. Geological Survey of New South Wales, Sydney.

White A. J. R., Clemens J. D., Holloway J. R., Silver L. T., Chappell B. W. & Wall V. J. 1986. S-type granites and their probable absence in southwestern North America. *Geology* **14**, 115–118.

White A. J. R., Williams I. S. & Chappell B. W. 1976. The Jindabyne Thrust and its tectonic, physiographic and petrogenetic significance. *Journal of the Geological Society of Australia* **23**, 105–112.

White A. J. R., Williams I. S. & Chappell B. W. 1977. *Geology of the Berridale 1:100 000 sheet 8625*. Geological Survey of New South Wales, Sydney.

Williams I. S. 1995. Zircon analysis by ion microprobe: the case of the eastern Australian granites. *In: Leon T. Silver 70th Birthday Symposium and Celebration*, pp. 27–31. California Institute of Technology, Pasadena.

Williams P. F. 1971. Structural analysis of the Bermagui area, NSW. *Journal of the Geological Society of Australia* **18**, 279–316.

Willman C. E., Morand V. J., Hendrickx M. A., VandenBerg A. H. M., Haydon S. J. & Carney C. 1999. Omeo 1:100 000 map area report. *Geological Survey of Victoria Report* **118**.

Wyborn D., Chappell B. W. & Johnston R. W. 1981. Three S-type volcanic suites from the Lachlan Fold Belt, southeast Australia. *Journal of Geophysical Research* **86(B11)**, 10335–10348.

Wyborn D. & Sun S-S. 1993. Nd-isotopic 'fingerprinting' of Cu/Au mineralization in the Lachlan Fold Belt. *AGSO Research Newsletter* **19**, 13–14.

Received 18 January 2000; accepted 31 July 2000