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PROBLEMS IN IDENTIFYING RESTITE IN S-TYPE GRANITES OF SOUTHEASTERN AUSTRALIA, WITH SPECULATIONS ON SOURCES OF MAGMA AND ENCLAVES

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Abstract

Evaluation of the extent of high-level contamination in granites requires a determination of the amount of restite or resistate (resister) present. High-level S-type granites of the Lachlan Fold Belt (LFB) of southeastern Australia are the most consistently interpreted in terms of the "restite unmixing hypothesis." Therefore, they should provide the clearest evidence for restite in high-level plutons; however, microstructural and isotopic evidence either is against restite or is ambiguous. Metasedimentary enclaves appear to be accidental xenoliths collected from metamorphic rocks above unexposed granulite-facies source-rocks; microgranitoid (orthopyroxene microtonalite) enclaves have igneous microstructures and show structural and isotopic evidence of mixing of S-type felsic magma with more mafic magma before becoming mingled to form enclaves in the host pluton. The origin of the single grains and small aggregates that have been considered to be the main evidence for restite in the S-type granites of southeastern Australia is difficult to determine. Unobserved deep, low-pressure and high-temperature granulite-facies sources (possibly varieties of "MASH" zones), rather than typical migmatite complexes, may be applicable for large, high-level S-type granites. A possible example of the upper part of such a melting zone is the Hidaka Metamorphic Belt, Hokkaido, Japan.

Keywords: granite, Lachlan Fold Belt, magma mixing, magma mingling, metasedimentary enclave, microgranitoid enclave, partial melting, peraluminous granite, restite, resistate, S-type granite, xenolith, southeastern Australia.

Sommaire

Une évaluation de la portée de la contamination épizonale des granites repose sur la détermination de la quantité de restite (résistat) présente dans les roches. On interprète généralement les granites de type S de la ceinture plissée de Lachlan, dans le sud-est de l'Australie, en termes de l'hypothèse de la "démixion de restite". Ces granites devraient donc fournir les indices les plus clairs de la présence de restite dans les plutons épizonaux. Toutefois, l'évidence microstructurale et isotopique va à l'encontre de la présence de restite, ou bien est ambigüe. Les enclaves métasédimentaires semblent être des xénolites accidentels pris de roches métamorphiques situées au dessus du socle granulitique enfoui ayant servi de source. Les enclaves microgranitiques (microtonalite à orthopyroxène) possèdent des microstructures ignées et montrent des indices structuraux et isotopiques de mélanges entre un magma felsique de type S et un magma plus mafique, et d'une étape de mélange plus intime avant d'être incorporées pour former les enclaves du pluton hôte. L'origine des monocristaux et des aggrégats de cristaux, considérés comme évidence primaire en faveur de la présence de restite dans les granites de type S du sud-est de l'Australie, est difficile à déterminer. Il se peut que des lithologies granulitiques non observées, équilibrées à faible pression et à température élevée, possiblement des variantes de zones "MASH" plutôt que des complexes migmatitiques typiques, soient le cas général pour les massifs épizonaux de granite de type S. Un exemple possible de la partie supérieure d'une telle zone de fusion serait la ceinture métamorphique de Hidaka, à Hokkaido, au Japon.

(Traduit par la Rédaction)

Mots-clés: granite, ceinture plissée de Lachlan, mélange de magmas, enclave métasédimentaire, enclave microgranitique, fusion partielle, granite hyperalumineux, restite, résistat, granite de type S, enclave énallogène, Australie.

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Introduction

This review is concerned with the identification of restite and resistate in high-level S-type granites, in the context of the Symposium on Granite Contamination. By "contamination", I mean addition of foreign material during magma ascent or emplacement, not at the anatectic source. Evaluation of the extent of high-level contamination in granites first requires determination of the amount of solid-source material (restite or resistate).

The production of relatively leucocratic granites requires the separation of restite from melt. This can occur (1) in the source rock, (2) during movement of the magma, or (3) in the emplaced magma chamber, forming cumulates. The last two processes lead to the potential difficulty of distinguishing restite from solid products of high-level contamination with wallrock material. Therefore, the question in the present context is: how much restite survives to the emplacement level? I will emphasize microstructural and some mineral isotope evidence for restite and resistate. Bulk chemical and isotopic evidence for restite unmixing (with or without magma mixing) has been considered in detail by others.

S-type granites are those with "chemical, isotopic and other properties indicating derivation from sedimentary or supracrustal source rocks" (White *et al.* 1991, p. 493). The S-type granites of the Lachlan Fold Belt (LFB) of southeastern Australia (Fig. 1) are the most consistently interpreted in terms of the "restite fractionation (unmixing) hypothesis." Therefore, they would be expected to provide the clearest evidence for restite in high-level plutons. I will concentrate on these granites, though the general inferences may apply to S-type granites everywhere.

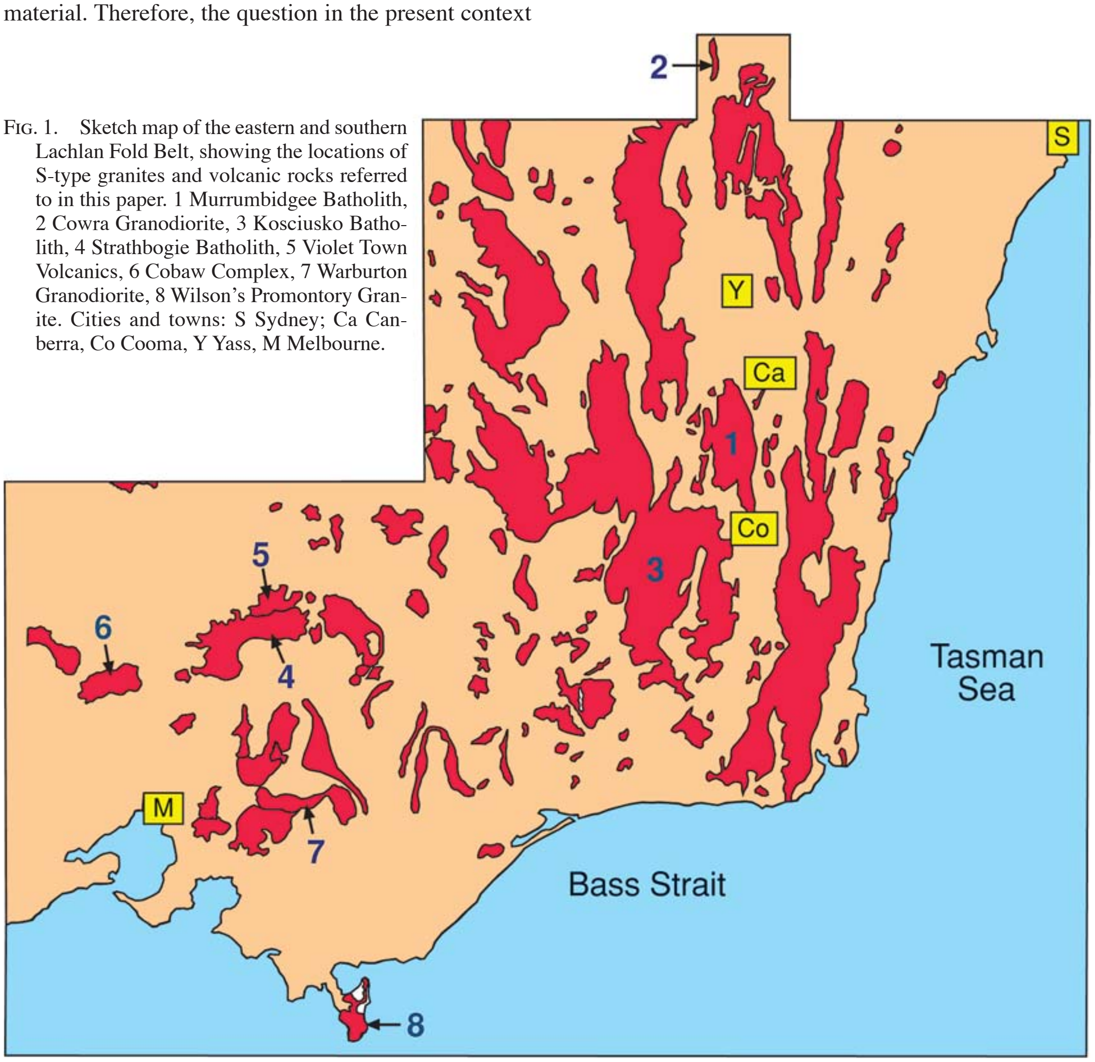

Fig. 1. Sketch map of the eastern and southern Lachlan Fold Belt, showing the locations of S-type granites and volcanic rocks referred to in this paper. 1 Murrumbidgee Batholith, 2 Cowra Granodiorite, 3 Kosciusko Batholith, 4 Strathbogie Batholith, 5 Violet Town Volcanics, 6 Cobaw Complex, 7 Warburton Granodiorite, 8 Wilson's Promontory Granite. Cities and towns: S Sydney; Ca Canberra, Co Cooma, Y Yass, M Melbourne.

Mineral abbreviations used in this paper are those recommended by Kretz (1983), extended by Bucher & Frey (1994).

Hypotheses

The restite-fractionation hypothesis has been championed for LFB granites by Chappell & White (1974, 1976, 1991, 1992, 2001), Chappell (1984, 2004), White & Chappell (1977, 1988), Griffin *et al*. (1978), Hine *et al*. (1978), Chappell *et al*. (1987, 1991, 1993, 1998, 2000, 2004), Wyborn *et al*. (1991), and Chappell & Wyborn (2004). For example, Chappell & White (1974) wrote: "S-type granitoids are considered to have been minimum melts or near-minimum melts with varying amounts of restite seen as metasedimentary rock xenoliths or as xenocrysts of biotite, cordierite and/or garnet. Some cordierites may contain telltale needles of sillimanite, indicating that they are xenocrysts from the source." They also stated that because the LFB S-type granites show no evidence of the former presence of kyanite, the melting must have occurred at <7 kbar (at 700°C) or <10 kbar (at 800°C).

On the other hand, Vernon (1983), Gray (1984), Elburg & Nicholls (1995), Elburg (1996b), Rossiter & Gray (1996), Keay *et al*. (1997), Maas *et al*. (1997, 2001a, b, c), Collins (1996, 1998), Collins *et al*. (2000, 2002), Collins & Richards (2001), Waight *et al*. (2000a, b, 2001), and Gray & Kemp (2001) inferred that the formation of various granitic magmas in the LFB involved both mantle and crustal magmatic contributions, as did Foden (2001) and Foden *et al*. (2002) for the Delamerian granites of southeastern South Australia, and Kemp (2001, 2004) and Kemp & Gray (1999) for the equivalent Glenelg River granites, occurring to the immediate west of the LFB. Estimates of the contribution of juvenile mafic magma to S-type granites range from none or minor amounts (Chappell & White 1992, Clemens & Watkins 2001) to 10–40% (Gray 1984, 1990, Collins 1996, Keay *et al*. 1997).

High-level S-type granites and related volcanic rocks in the LFB typically show evidence of some magma mixing and mingling. For example, they commonly contain peraluminous microgranitoid (typically microtonalite) enclaves that show evidence of hybridization between peraluminous felsic magma and more mafic magma, as discussed later. Therefore, mafic magma must have had some effect on the compositions of the S-type magmas involved in the mixing, though not necessarily the enclosing granitic magma. Whatever the proportion of added mafic magma, it was evidently not sufficient to remove the peraluminous character of LFB S-type granitic magmas (Clemens 2003).

The evaluation of these petrogenetic models is outside the scope of this review. However, it should be appreciated that magma-mixing models for S-type granites in the LFB actually combine magma mixing with restite or resistate fractionation. Because these models invoke restite or resistate-rich "regional aureole" granite (*e.g.,* the Cooma Granodiorite) as an end-member in the mixing process (Gray 1984, Collins 1996, Keay *et al*. 1997, Gray & Kemp 2001), restite or resistate fractionation necessarily would be involved in the differentiation responsible for more evolved rock-types according to these models.

Though much chemical evidence is consistent with the restite-fractionation hypothesis, unequivocal microstructural evidence is difficult to obtain. For example, Clemens (2001) stated that no clear evidence has been presented to show that the mafic minerals concentrated in the more mafic S-type granites are restite, rather than magmatic cumulates.

As well as single grains and small aggregates, both metasedimentary and microgranitoid enclaves have been regarded as restite by proponents of the restite hypothesis. Therefore, I will examine microstructural and some other evidence for the origin of single grains and small aggregates, as well as both classes of enclave (for both granites and related volcanic rocks), after which I will speculate on the nature of source areas for S-type granitic magmas.

Definitions of Restite and Resistate

White & Chappell stated that "at the source region, the product of partial melting is a mixture of melt and solid material, the latter sometimes being referred to as restite." A definition of *restite* presented by Chappell & White (1991, p. 376) is: "any solid material in a plutonic or volcanic rock that is residual from partial melting of the source." For me, the implication of both these definitions is that "source" means the specific parent rock that melted to produce the granitic magma under consideration. This view of restite is consistent with chemical arguments for the restite-segregation hypothesis applied to the explanation of the compositional variation of granite (Chappell & White 1974, White & Chappell 1977).

A clear, simple definition of *restite* is: "the residual material from which mobile material has been extracted" (Shelley 1993, p. 109). A more comprehensive definition is: "high-grade residual mineral phases or assemblages either in excess over dissolution or melting proportions or resulting from incongruent dissolution or melting" (Barbey 1991).

I suggest that enclave material that was restite for a melting reaction that occurred previously, and so did not give rise to the host granite, should be excluded from this definition, and should be included with *resister* (= *resistate*), which is a term applied to rocks in the source region that have not melted at all. A clear, simple definition of *resister* (= *resistate*) is: "rocks that resisted migmatization" (Shelley 1993, p. 109). If a rock in the source zone has partly melted previously and solidified before being incorporated as solid fragments in a magma, it can be considered to be resistate with respect to that magma. Distinguishing between restite and resistate in this sense would help to better understand processes involved at and near the source (see discussion of the Cooma Granodiorite below). I suggest that a more comprehensive definition of *resister* or *resistate* is: “a rock that resists melting during a particular anatectic event.”

In some publications (*e.g.,* White *et al*. 1991), the definition of restite appears to include resistate in this sense, which adds a fortuitous element to, and in that sense detracts from, the elegant simplicity of the original restite-fractionation hypothesis. Nonetheless, both restite and resistate theoretically could progressively unmix, as in the restite-fractionation model, or could be included in end-member compositions for mixing models that also involve the unmixing restite or resistate, as mentioned previously.

Single Grains and Small Aggregates of Inferred Restite

The main putative restite in the LFB S-type granites consists of individual grains (especially plagioclase cores and quartz) and small aggregates (especially of biotite), as emphasized by Chappell *et al*. (1987), who defined *primary restite* as residual crystals from the source and *secondary restite* as residual crystals that have “recrystallized”. Chappell *et al*. (1987, p. 1128) inferred that, apart from calcic plagioclase, primary restite minerals are generally destroyed during slow cooling.

Quartz restite would be expected from partial melting of quartzofeldspathic rocks (discussed below), but may be difficult to identify in granites; perhaps cathodoluminescence may help in this regard. Unzoned cores of calcic plagioclase conceivably could be restite, but may be formed in other ways. For example, the crystallization of plagioclase in felsic melts at small degrees of undercooling can produce large, homogeneous crystals (Swanson 1977), as pointed out by Wall *et al*. (1987), and magma mixing may also produce calcic plagioclase cores (*e.g.,* Gray & Kemp 2001). Moreover, near-liquidus plagioclase is calcic in hydrous felsic magma, especially at deep levels in the crust and moderate H_2O contents in the melt (Wall *et al*. 1987). Calcic (>An_{70}) cores in plagioclase of the Jindabyne Suite S-type granites, previously used as the strongest evidence of restite (*e.g.,* Chappell *et al*. 1987), have been shown, on the basis of detailed trace-element data, to more probably be the products of fractional crystallization with variation of $a(H_2O)$ (Allen 2001). The argument used by Chappell *et al*. (1987), that because plagioclase cores tend to have a constant composition in both mafic and felsic members of a granitic “suite” they must be restite, can be countered by the suggestion that they may equally well be the result of mixing with mafic magma.

Grains and cores of garnet and cordierite with sillimanite inclusions have been taken to indicate restite by Chappell & White (1974, 1991), but sillimanite inclusions in garnet and cordierite are common in high-grade metamorphic rocks generally, and so need not indicate restite; they could be accidental xenocrysts collected during ascent. Moreover, inclusions of magmatic sillimanite may occur in cordierite and muscovite phenocrysts (Zeck 1972, Roycroft 1991, Büttner 2005).

Inherited cores of zircon grains could be either refractory restite released from host grains into the melt during partial melting (especially of biotite) or xenocrysts accidentally collected during movement of the magma. Chappell *et al*. (1987) argued against a xenocrystic interpretation by stating that there is no mineralogical or chemical evidence for contamination in the S-type granites. However, the metasedimentary xenoliths in the S-type granites appear to be accidental, as discussed below, and so at least some zircon could be xenocrystic.

Many S-type granites contain grains and small aggregates of quartz (generally milky), which in the opinion of White *et al*. (1976) and Chappell & White (1991) come from disaggregated quartz veins in the metasedimentary source-rocks (*i.e.,* resistate, rather than restite). However, quartz veins are common in most metamorphic terranes (*e.g.,* metasediments of a mid- to upper-crust level of southeastern Australia), and so are not necessarily indicators of a source region. The Cooma Granodiorite (Fig. 1) contains scattered fragments of relatively coarse-grained quartz and K-feldspar, which result from physical disintegration of solidified leucosome in metapelite-derived migmatites, and consequently are resistate, not restite (Vernon *et al*. 2001, 2003), as discussed below.

Phenocrysts in S-Type Volcanic Rocks: Magmatic or Restitic?

High-temperature phenocrysts (especially garnet, orthopyroxene and calcic plagioclase) in S-type volcanic rocks are usually considered to be magmatic (true phenocrysts), though Wyborn *et al*. (1981) and Wyborn & Chappell (1986) suggested that they are restitic (pseudophenocrysts). The distinction is important for the evaluation of restite in S-type granites, because if it can be shown that the crystals of garnet, orthopyroxene and calcic plagioclase are magmatic, high temperatures of crystallization are implied for the volcanic rocks and hence for the equivalent granites, whereas if it can be shown that they are restite, lower temperatures of crystallization approaching those in the haplogranitic system may apply (*i.e.,* during crystallization of groundmass quartz and feldspar only). I will discuss relevant examples of phenocrysts in peraluminous felsic volcanic rocks from elsewhere, as well as the LFB.

Phenocrysts in S-type volcanic rocks of the Hawkins Suite

Wyborn *et al.* (1981) and Wyborn & Chappell (1986) inferred that euhedral phenocrysts of plagioclase, cordierite, orthopyroxene, biotite and quartz in the Hawkins Suite S-type felsic volcanic (ignimbritic) rocks, north of Canberra, Australian Capital Territory (Fig. 1), are erupted restite that underwent only limited or no magmatic reaction before quenching. These rocks are compositionally equivalent to the S-type granites of the Bullenbalong Suite of the Kosciusko Batholith (Fig. 1). The phenocrysts are commonly euhedral (with fragments of euhedra, owing to explosive eruption), some with oscillatory zoning (Fig. 2).

Maas *et al.* (1999, 2001a) examined phenocrysts in a dacite from the Hawkins Volcanics, and found that the phenocrysts of quartz, orthopyroxene and apatite contain inclusions of silicate melt (now glass), whereas those of cordierite and plagioclase contain what appear to be altered melt-inclusions, (Figs. 3, 4). For example, the quartz phenocrysts contain euhedral inclusions of plagioclase (An_{50-73}), orthopyroxene, apatite and biotite, indicating magmatic crystallization of these minerals. The silicate glass inclusions were inferred to represent melt droplets trapped in the growing crystals during rapid cooling, suggesting not only that the phenocrysts are igneous, but also that they crystallized at a shallow (volcanic or subvolcanic) level in the crust. The fact that the former inclusions of melt occur even in the cores of the phenocrysts was taken to indicate that the cores also are igneous crystals, not metamorphic restite (Maas *et al.* 1999, 2001a).

An alternative interpretation is that the glass inclusions represent melt formed during the anatexis that gave rise to the host magma; that is, the host phenocrysts represent solid peritectic products (restite) of the melting reaction that incorporated droplets of melt as they grew and preserved them as glass during ascent and solidification of the magma. However, this explanation cannot apply to quartz or intermediate plagioclase phenocrysts, because these minerals are reactants, not solid peritectic products of the reactions responsible for the felsic magmas under discussion (*e.g.,* Vielzeuf & Holloway 1988, Vernon *et al.* 2001, 2003). Therefore, the melt inclusions must have been incorporated as the quartz and plagioclase crystallized from the magma. If this explanation applies to glass inclusions in quartz and plagioclase, it may equally apply to glass inclusions in the other phenocrystic minerals in these rocks. Glass inclusions in quartz phenocrysts are common in felsic volcanic and pyroclastic rocks elsewhere (*e.g.,* Clocchiatti 1975, Sommer 1977, Beddoe-Stephens *et al.* 1983, Skirius *et al.* 1990).

A restite interpretation is inapplicable not only to melt inclusions concentrated in selected growth-zones of plagioclase phenocrysts (Fig. 2), but also to melt inclusions in the cores of the plagioclase grains (Maas *et al.* 1999, 2001a). If these cores were restite, they should not have melt inclusions, unless the cores were partly dissolved or otherwise broken down in a later reaction.

An additional argument in favor of a magmatic origin is the occurrence of abundant inclusions of plagioclase with wide compositional ranges in quartz and orthopyroxene phenocrysts (Maas *et al.* 2001a).

More may remain to be written about this problem, as more data on the inclusions become available, but one point can be made at this stage: glass inclusions in quartz and plagioclase phenocrysts cannot represent primary melt trapped during anatexis, for reasons outlined above.

Inclusions in phenocrysts of the Violet Town Volcanics, Victoria

The ignimbrites of the Violet Town Volcanics, Victoria (Fig. 1) have rare phenocrysts of garnet with a core containing sillimanite and quartz inclusions, and an inclusion-free rim inferred to represent magmatic overgrowths (Clemens & Wall 1984). Also present are rare cordierite grains with inclusions of sillimanite or green zincian hercynite or both. Both types of core were inferred to be of either restitic or unrelated xenocrystic origin by Clemens & Wall (1984). They make up less than 2 vol.% of the ignimbrites. The remainder of the garnet and cordierite phenocrysts, along with all the orthopyroxene phenocrysts in the ignimbrites, were inferred to be magmatic (Clemens & Wall 1984), as witnessed by their commonly euhedral shape, lack of metamorphic inclusions, distinctly different compositions, and the presence of devitrified rhyolitic glass in the orthopyroxene. The compositions of the garnet phenocrysts are similar to those produced by experimental crystallization of similar magma.

Phenocrysts in the Cerro del Hoyazo Dacite, southeastern Spain

Some phenocrysts in a peraluminous dacite at Cerro del Hoyazo, southeastern Spain, have been interpreted as magmatic, others as restitic (Zeck 1970). Phenocrysts inferred to have crystallized from the magma include euhedral biotite, labradorite and sector-twinned cordierite (Zeck 1970), trillings being very characteristic of cordierite precipitated from natural and synthetic melts (Zeck 1972). Ubiquitous oscillatory zoning also suggests that these crystals are magmatic. Despite these magmatic characteristics, the cordierite contains numerous small, acicular inclusions of sillimanite, most of which are arranged crystallographically (*e.g.,* concentrically) with regard to the host, though some appear to be random (Zeck 1972). The euhedral shape, oscillatory zoning and characteristic twinning of the cordierite, coupled with the crystallographic arrangement of the inclusions, indicate that both the cordierite

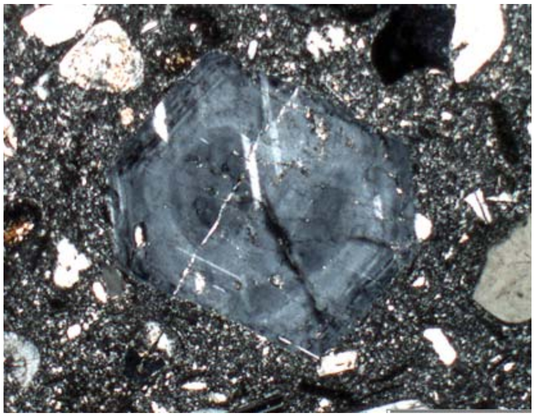

FIG. 2. Plagioclase phenocryst in S-type felsic volcanic rock, the Hawkins Dacite, north of Yass, New South Wales. The outer parts of the plagioclase show oscillatory zoning. Image by courtesy of Dima Kamenetsky and Roland Maas. Crossed polars; the field of view is 3.3 mm wide.

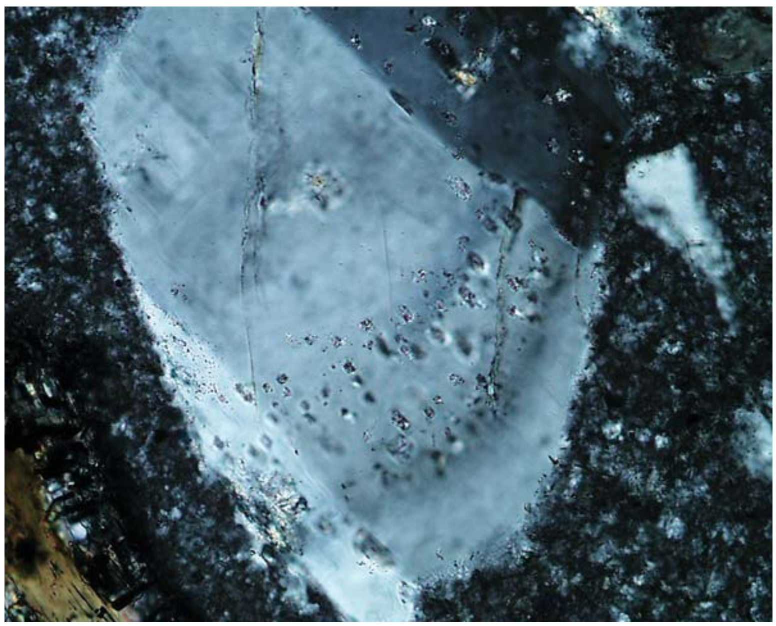

FIG. 3. Zones of altered glass inclusions in plagioclase phenocryst in S-type felsic volcanic rock, the Hawkins Dacite, north of Yass, New South Wales. Image by courtesy of Roland Maas. Crossed polars; base of photo 0.22 mm.

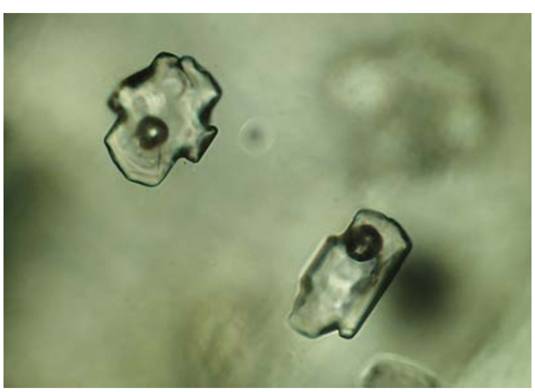

FIG. 4. Glass inclusions in orthopyroxene phenocryst in S-type felsic volcanic rock, the Hawkins Dacite, north of Yass, New South Wales. The inclusions have negative-crystal shapes. Image by courtesy of Roland Maas. Plane-polarized light; base of photo 0.44 mm.

and sillimanite precipitated simultaneously from the magma. This situation is similar to sillimanite needles aligned parallel to crystal faces in magmatic muscovite (Roycroft 1991). These occurrences show that sillimanite inclusions are not reliable indicators of a restitic or xenocrystic origin for muscovite or cordierite.

Individual crystals of almandine-rich garnet have been inferred to be solid fragments, not magmatic precipitates (Zeck 1970, Munksgaard 1985), owing to similarity in size, shape and compositional zoning to garnet grains in metasedimentary enclaves, though Munksgaard (1985, p. 80) was careful to point out that such an interpretation does not necessarily indicate that the garnet is restite for the enclosing dacite; a xenocrystic origin is equally probable on the available evidence. Moreover, garnet in the metasedimentary enclaves shows evidence of resorption and partial melting in the form of glass rims (Cesare 2000, Fig. 2), and glass–sillimanite aggregates occurring in replacement veins cutting evidently corroded porphyroblasts of garnet and in strain shadows against corroded garnet porphyroblasts (Cesare 2000, Fig. 1). Thus, the garnet is not a product of a peritectic reaction, but existed prior to melting.

Some isolated grains of biotite, hercynitic spinel, cordierite, sillimanite and quartz also have been inferred to represent physically disintegrated metasedimentary enclaves (Cesare *et al*. 1997). If this is so, it is not an argument for restite, because if the enclaves are xenoliths, the grains would be xenocrysts. The inferred fragmental biotite and cordierite have lower *mg* values than the inferred magmatic biotite and cordierite (Zeck 1970, Munksgaard 1985).

Metapelitic enclaves in the Cerro del Hoyazo Dacite, southeastern Spain: sources of phenocrysts?

The dacite at Cerro del Hoyazo contains a small proportion of aluminous metapelitic enclaves containing minerals with felsic glass inclusions, inferred to be of primary origin (Cesare *et al*. 1997, 2003). These enclaves are potential sources of phenocrysts with glass inclusions in S-type volcanic rocks, and so are discussed here. The enclaves are mainly garnet – biotite – sillimanite gneiss and spinel–cordierite rock, and have been interpreted as restite for the dacite (Zeck 1970, Cesare *et al*. 1997). Though the aluminous composition of the enclaves is consistent with a restite origin, microstructural evidence does not preclude a xenolithic origin, inasmuch as it does not indicate that the enclaves are specifically restite for the enclosing dacitic magma.

The garnet – biotite – sillimanite gneisses resemble typical metapelitic regional metamorphic rocks, being strongly foliated and microfolded, with garnet and former garnet porphyroblasts containing curved inclusion-trails, and with earlier microfolds preserved in strain shadows of porphyroblasts. However, they contain rhyolitic glass in films parallel to the foliation, patches in strain shadows adjacent to porphyroblasts, aggregates intergrown with sillimanite in pseudomorphs of inferred former porphyroblasts of garnet, thin veins perpendicular to the foliation and cutting the sillimanite–glass pseudomorphs, and small inclusions in most of the minerals (Cesare *et al*. 1997, 2003), though cordierite is generally devoid of glass inclusions and ilmenite has none (Cesare 2000, p. 274). The glass has been inferred to represent chilled anatectic melt formed during growth of the minerals (Cesare *et al*. 1997), mainly on the basis of the shapes of the glass inclusions being consistent with a primary origin (*e.g.,* Roedder 1979, 1984). However, though many primary inclusions have negative-crystal shapes (commonly with rounded corners), some are spherical (*e.g.,* Roedder, 1979, Fig. 14). Moreover, secondary melt-inclusions may "neck down" to form spherical and negative-crystal shapes (Frezzotti *et al*. 1999). In addition, some of the glass inclusions in andalusite shown by Cesare *et al*. (2003, Fig. 2, p. 574) have irregular to spherical, tubular and lenticular shapes; some are arranged crystallographically with regard to the host, whereas others are random. Moreover, melting in response to loss of stability of a mineral assemblage may conceivably produce melt droplets dispersed all through the grains of the unstable mineral if inclusions were originally present, which is a common situation in metamorphic assemblages.

The metasedimentary enclaves show abundant evidence of disequilibrium melting, in the form of extreme local variation in melt composition, such as spinel-bearing glass adjacent to partly resorbed biotite and cordierite, orthopyroxene in glass adjacent to partly resorbed biotite and quartz (Zeck 1970), sodic glass with calcic rims in partly dissolved plagioclase (Zeck 1970), contrasts in the composition of glass adjacent to and slightly further away from resorbed biotite (Cesare 2000), and variation of the composition of glass inclusions in andalusite (Cesare *et al*. 2003, p. 575). This evidence indicates rapid melting, without time for equilibration on the hand-specimen scale, or even for crystallization of the melt, apart from very small crystals of sanidine, orthopyroxene and spinel. As noted by Cesare (2000, p. 274), partial melt in migmatites should tend to equilibrate, whereas these rocks show extreme heterogeneity of the melt on the thin-section scale. This is an argument in favor of the inference that partial melting occurred after incorporation of the rocks as solid metamorphic xenoliths, and does not support the suggestion of anatexis in the source area. The melting may have occurred relatively rapidly during ascent of the enclaves, owing to superheating of H_2O-undersaturated granitic magma during rapid, adiabatic rise in the crust, as predicted by Clemens *et al*. (1997).

Cesare *et al*. (1997) inferred that all the minerals of the metapelitic gneisses grew in the presence of melt, during the formation of the foliation and microfolds, as supposedly indicated by the glass inclusions that occur even in the centers of grains. However, typical solid-state deformation-induced structures such as foliation or microfolding would not develop in the presence of melt, as ”stress transfer is inefficient within a liquid” (Cesare *et al*. 1997, p. 24). A more plausible interpretation is that the structures formed in the solid state and that melting occurred later, after incorporation of the xenoliths into the host magma. Concentrations of glass inclusions in garnet cores (Cesare *et al*. 1997) suggest greater chemical instability of inclusion-rich cores during the melting, rather than incorporation of melt droplets during growth.

In fact, garnet in the metasedimentary enclaves shows evidence of resorption and partial melting in the form of (1) glass rims (Cesare 2000, Fig. 2) and (2) glass–sillimanite aggregates occurring in replacement veins cutting corroded garnet porphyroblasts and in strain shadows against corroded porphyroblasts of garnet (Cesare 2000, Fig. 1). Thus, the garnet is not a product of a peritectic reaction, but existed prior to melting, implying that its glass inclusions must have formed later.

The presence in some metasedimentary enclaves of andalusite (Zeck 1970, Cesare *et al*. 2003) indicates metamorphism at pressures and temperatures too low for typical granulite-facies partial melting reactions, though Cesare *et al*. (2003) suggested H_2O-present melting on the basis of their inference that the andalusite grew in the presence of melt (forming inferred inclusions of primary melt). They suggested that “in low-*P* partial melting of metapelites, the coexistence of And + melt should be the rule”. However, melting responsible for granitic magmas in amphibolite facies, low-*P* migmatite terranes is typically the result of H_2O-absent reactions involving destruction, not growth, of andalusite (*e.g.,* Vernon *et al*. 2003). Moreover, H_2O-present melting cannot produce magmas capable of ascending in the crust; therefore, if H_2O-saturated melting were responsible for the melt in the enclaves, it could not have been responsible for the enclosing dacitic magma, and so the enclaves would not be restite for the dacite.

Melt inclusions in xenolithic quartz, Aeolian Arc

Quartz-rich xenoliths (inferred to have originally been quartzofeldspathic gneiss or mica schist) in volcanic and pyroclastic rocks (basalt to rhyolite) of the Aeolian volcanic Arc, Italy, have been partly melted (Frezzotti *et al*. 1999). The quartz grains have former melt along their boundaries, and also have silicate–melt (glass) inclusions. Some glass occurs in clusters or short trails leading from grain-boundary glass, suggesting that the inclusions were formed by secondary entrapment of melt formed by partial melting. The inclusions have rounded to negative-crystal shapes, and may contain incidentally trapped crystals, such as wollastonite, pyroxene, feldspar, sulfides or oxides. This occurrence indicates that melt inclusions in partly melted solid rock need not be incorporated during growth of the host mineral. Instead, they may be the result of secondary entrapment or to direct melting of the host plus impurity mineral(s).

Proportion of phenocrysts in LFB volcanic rocks

An argument advanced by Chappell *et al*. (1987) in favor of a restite origin for phenocrysts in LFB volcanic rocks is that they are too magnesian for liquids of their bulk composition and that the rocks contain too many phenocrysts. For example, they stated that the Hawkins Volcanics contain about 60 vol.% of phenocrysts. However, the possibility that the phenocrysts could be cumulative also needs to be considered. If magmas rich in putative restite can erupt volcanically, so presumably can magmas equally rich in magmatically accumulated crystals.

Regional- and Contact-Aureole Granites

White *et al*. (1974) distinguished between *regional-aureole S-type granites* and higher-level *contact-aureole S-type granites* in southeastern Australia. The contact-aureole S-type granites have also been called “batholithic S-types” by White & Chappell (1988) and Chappell & White (1991). The regional-aureole S-type granitic magmas are relatively cool, H_2O-rich and consequently relatively immobile, with the result that the granites typically occur in or close to parental migmatite terranes and contain abundant unmelted material (*i.e*., they were diatexites). In contrast, contact-aureole S-type granitic magmas are hot, H_2O-undersaturated and consequently mobile (Clemens 1984, 1988), with the result that the granites occur with contact aureoles in low-grade metamorphic areas.

In this paper, enclaves in regional-aureole LFB S-type granites are discussed first, followed by metasedimentary enclaves in contact-aureole LFB S-type granites and volcanic equivalents, and then by microgranitoid enclaves in contact-aureole granites.

Enclaves in Regional-Aureole Granites

General comments

Partial melting has produced migmatites in LFB sillimanite – K-feldspar zone (upper amphibolite facies) rocks, as well as relatively small, local “dirty granite” (diatexite) plutons, for example, in the Cooma Complex (see below), the Albury area (Joplin 1947) and the Omeo Complex (Fagan 1979, Morand 1990). Larger, higher-level bodies of synmetamorphic S-type granitic magma in these areas appear to have been derived from partial melting of metasedimentary crust beneath the exposed Ordovician rocks, according to Fagan (1979) and Morand (1990). Even the relatively small Cooma Granodiorite appears to consist partly of magma added from deeper levels (Collins & Richards 2001, Richards & Collins 2002, Vernon *et al*. 2003).

The Cooma granodiorite

The Cooma Granodiorite (Fig. 1) is typical of the regional-aureole type of diatexitic, S-type pluton in the LFB, and has been studied in the most detail. It is a small, peraluminous pluton in contact with high-grade (upper amphibolite facies) metasedimentary rocks, including migmatites (Joplin 1942, Hopwood 1976, Johnson *et al*. 1994, Vernon & Johnson 2000, Vernon *et al*. 2001, 2003, Collins & Richards 2001, Richards & Collins 2002). It is variably foliated, contains abundant metapelite enclaves, has no microgranitoid enclaves, and contains grains and small aggregates of milky quartz.

The metapelitic enclaves are widely inferred to be restite, and the Cooma Granodiorite is generally inferred to be a product of the metamorphism and partial melting of the metasediments of the Cooma Complex (Pidgeon & Compston 1965, White *et al*. 1974, 1991, Chappell & White 1976, Munksgaard 1988, White & Chappell 1988, Chappell *et al*. 1991, Ellis & Obata 1992, Williams 1998, 2001, Richards & Collins 2002).

In keeping with the interpretation that the Cooma Granodiorite was formed by *in situ* or practically *in situ* melting of the adjacent metasediments, Ellis & Obata (1992, p. 95) stated that "a progression can be seen in the field from unmelted sediments through migmatites to the granodiorite," and in several places, the main body of granodiorite grades into migmatites. However, it also contains sharply bordered enclaves identical to the adjacent migmatites, suggesting that these migmatite enclaves are xenoliths. Moreover, recent work (Vernon & Johnson 2000, Vernon *et al*. 2001, 2003) has shown that the popular view that melt extracted from the conspicuous stromatic, metapelite-derived migmatites was a major contributor to the Cooma Granodiorite magma (Ellis & Obata 1992, Richards & Collins 2002, p. 132, Williams 2001) is no longer tenable. The reason is that (1) metapelite melting produced relatively immobile neosomes that largely stayed in the metapelite beds (Vernon & Johnson 2000, Vernon *et al*. 2003), (2) the composition of the metapelite-derived neosome (mainly quartz, K-feldspar and cordierite, with very minor or no plagioclase) indicates that it is unsuitable as a source magma for the plagioclase-rich Cooma granodiorite (Vernon & Johnson 2000), and (3) mobile plagioclase-rich magma produced by melting of feldspathic metapsammite grades locally into the Cooma Granodiorite, and has disrupted, boudinaged and engulfed xenoliths and xenocrysts of the metapelitic neosome and leucosome, which solidified before the metapsammite melting (Vernon *et al*. 2001, 2003). Therefore, the contribution of metapelite-derived neosome to the Cooma Granodiorite is in the form of solid material, not melt. Furthermore, because it did not form as a residue from the partial melting that produced the Cooma Granodiorite, it is resistate, not restite. So too are the abundant enclaves of more mafic, metapelite-derived mesosome, which represent partial restite for the metapelite-derived neosome or leucosome, but not for the Cooma granodioritic magma.

Thus, though the Cooma Granodiorite can be regarded, chemically and isotopically, as an approximate mixture of metapelitic and metapsammitic sediments of the Cooma Complex, the only direct liquid contribution came from the feldspathic metapsammites, the remainder being contributed as solid material. This conclusion shows that chemical evidence alone may not give an accurate picture of events, and that careful distinction between restite and resistate can lead to increased understanding of processes in source terranes.

Though locally the metapsammite-derived neosome grades into massive Cooma Granodiorite (*e.g.,* Vernon *et al*. 2001, 2003), the main body of granodiorite appears to be too large for an origin entirely by local accumulation of neosome. Therefore, magma formed by similar partial melting of deeper feldspathic metapsammites may have ascended to its present position, probably accumulating local metapsammite-derived neosome on the way. Collins & Richards (2001) suggested that ascending magma responsible for the Murrumbidgee Batholith (Fig. 1) may have mixed with the locally derived magma to form the Cooma Granodiorite.

Recognition of restite in the Cooma Granodiorite is difficult, though some of the quartz-rich nature of the granite (*e.g.,* White *et al*. 1991) could be attributable to restitic quartz (*e.g.,* Ellis & Obata 1992). Thus, the Cooma Granodiorite appears to be a resistate-rich granite, with some restite quartz and possibly minor restite mafic and aluminous minerals from the partial melting of feldspathic metapsammite. However, some of this material may come from disaggregated metapelite laminae in the psammites, and therefore may be technically resistate.

METASEDIMENTARY ENCLAVES IN CONTACT-AUREOLE GRANITES

General comments

Metasedimentary enclaves are common in high-level LFB granites, though generally sparsely dispersed (<1% of the pluton), except for local concentrations of around 10%. They occur in both I- and S-type granites, but are more abundant in S-type granite (Anderson *et al*. 2001). They cannot reasonably be regarded as restite in I-type granites, but have been interpreted as restite in S-type granites (*e.g.,* White & Chappell 1977, Chappell *et al*. 1987, Chappell & White 1991).

The Cowra Granodiorite

The Cowra Granodiorite, about 250 km west of Sydney (Fig. 1), is a contact-aureole S-type granite containing roughly equal amounts of metasedimentary and microgranitoid enclaves (see later), as shown in Figure 5. The metasedimentary enclaves (Stevens 1952, White *et al.* 1991, Chappell *et al.* 1993, Waight *et al.* 2001) are mainly of metapelite, metapsammite and calc-silicate rock. The assemblages of minerals in calc-silicate rocks (epidote, actinolite, plagioclase, quartz, calcite, chlorite) reflect metamorphism at about 500°C, so that they cannot be from the source region of the magma; they appear to be xenoliths (Waight *et al.* 2001). Some vein quartz also is present; this could be from either the source region or higher-level metamorphic rocks.

The metapelite enclaves (Figs. 6, 7) are schistose, show microfolding, and consist of varying proportions of sillimanite, K-feldspar, quartz, plagioclase, cordierite, spinel, biotite, corundum, ilmenite and garnet. Inclusions of spinel and corundum are common in the cordierite, the corundum apparently having replaced spinel. Garnet occurs in some of the gneissic enclaves, as scattered grains in cordierite and irregularly shaped grains partly replaced by cordierite. Some garnet grains have sillimanite inclusions. A garnet-bearing metapelite enclave examined by Waight *et al.* (2001) shows evidence of partial melting; however, these enclaves are too aluminous to be restite related to the granitic magma, as discussed later.

Waight *et al.* (2001) have shown that the metasedimentary enclaves have lower ε_{Nd} than the host granodiorite, and so are not representative of the source of the granitic magma. Their isotopic compositions plot in or near the field of Paleozoic LFB sediments, and so they are probably xenoliths collected accidentally from the sedimentary succession above the source region (Waight *et al.* 2001).

The Jillamatong Granodiorite

The Jillamatong Granodiorite, which belongs to the Bullenbalong Suite of the Kosciusko Batholith (Fig. 1), contains sillimanite-bearing, biotite-rich, schistose enclaves of pelitic composition, as well as foliated quartzofeldspathic enclaves that probably represent metasandstones (Chen *et al.* 1989, White *et al.* 1991). Chen *et al.* (1989) found that the schistose enclaves are not restite for the host magma, because the plagioclase is less calcic and the cordierite less magnesian than in the host granite. They suggested that the enclaves are fragments of resistate that did not melt to an appreciable extent because of a deficiency in one or more of the chemical components needed to produce enough extractable melt.

The Strathbogie Batholith

Phillips *et al.* (1981) described high-grade regional metamorphic, foliated, fine- to medium-grained metapelitic enclaves in the S-type granites of the Strathbogie Batholith, Victoria, Australia (Fig. 1). They inferred that the enclave compositions are compatible with restite, on the basis of alkali and silica depletion, but noted that their low-pressure assemblages (inferred to have resulted from re-equilibration during ascent of the host magma) preclude estimation of the melting conditions. They stated that melting of metapelite, with high K/(Na + Ca), produces magma saturated with K-feldspar throughout its cooling history, whereas the Strathbogie magma became saturated in K-feldspar relatively late, and so was probably produced by melting of quartzofeldspathic rocks poor in K-feldspar. Therefore, they concluded that the metapelitic enclaves probably represent melt-depleted rocks that are not typical of the composition of the source for the Strathbogie granites. In other words, if these enclaves came from the source area, they are resistate, rather than restite, for the Strathbogie magma. Alternatively, they may be accidental xenoliths derived from metamorphic rocks outside the source.

The Deddick Granodiorite

The Deddick Granodiorite is an S-type pluton in the southern Kosciusko Batholith (Fig. 1), and has been described by Maas *et al.* (1997, 1998, 1999, 2001b) and Nicholls *et al.* (1999, 2001). Metasedimentary enclaves are abundant, and commonly contain sillimanite, cordierite, corundum and zincian hercynite (Ringwood 1955, Maas *et al.* 1997, 1998, 1999, 2001b, Nicholls *et al.* 1999, 2001). In gneissic enclaves, metapelitic layers rich in biotite, with cordierite, fibrous sillimanite and rare garnet, alternate with metapsammitic layers rich in quartz, with variable proportions of plagioclase and K-feldspar. Also present are enclaves of migmatite containing lenses and patches of leucosome that may be folded, and melanosome rich in cordierite, garnet and biotite. Some fragments inferred to be melanosome rich in cordierite, garnet, spinel and corundum appear to be restite from partial melting (Nicholls *et al.* 2001), though probably not involved in the formation of the host magma, because of isotopic evidence (see below).

Compositional and Sr–Nd isotopic data indicate that the metasedimentary enclaves are not in equilibrium with the host granite. For example, plagioclase in the enclaves is less calcic, and mafic minerals are less magnesian, than the equivalent minerals in the host granite (Maas *et al.* 1997). This observation suggests either that the enclaves were derived from a minor rock-type in the source area (*i.e.,* they are resistate), or that they are accidental xenoliths. However, if the enclaves represent a minor lithological heterogeneity in the protolith, restite enclaves from the dominant source-rock should also be present. Maas *et al.* (1997, 2001b) suggested that the enclaves are probably fragments of deeper equivalents of the locally exposed low-grade Ordovician metaturbidites, as their chemical and Nd–Sr

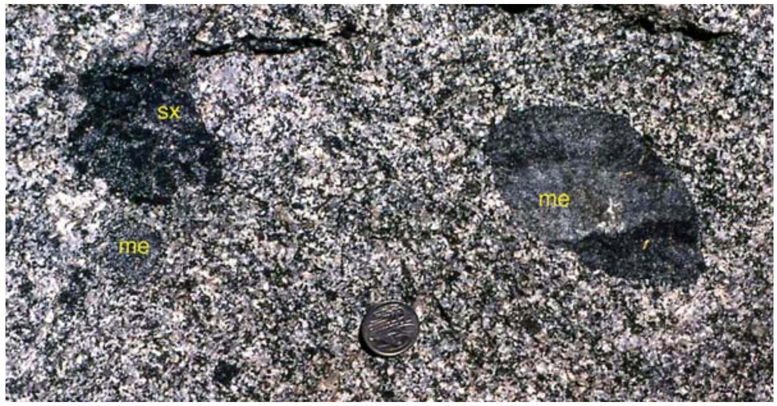

FIG. 5. Microgranitoid enclaves (me) and a metasedimentary enclave (sx) in the Cowra Granodiorite. Diameter of coin 28 mm.

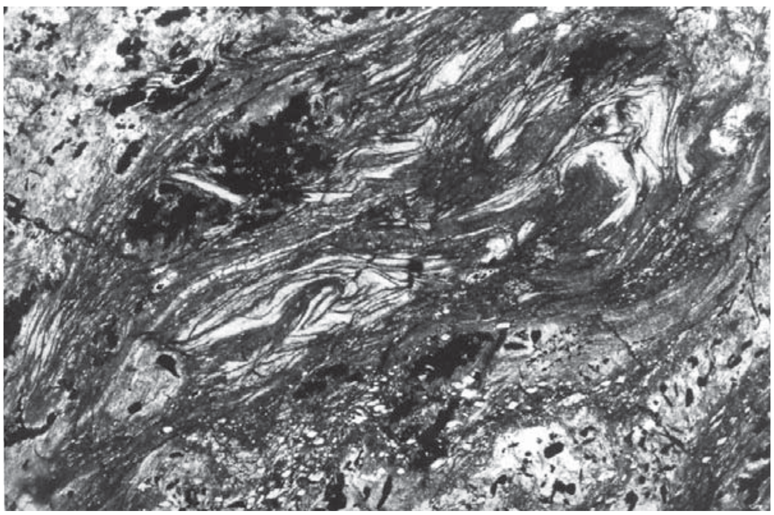

FIG. 6. Tightly folded fibrous sillimanite in metapelitic enclave, Cowra Granodiorite. Plane-polarized light; base of photo 4.4 mm.

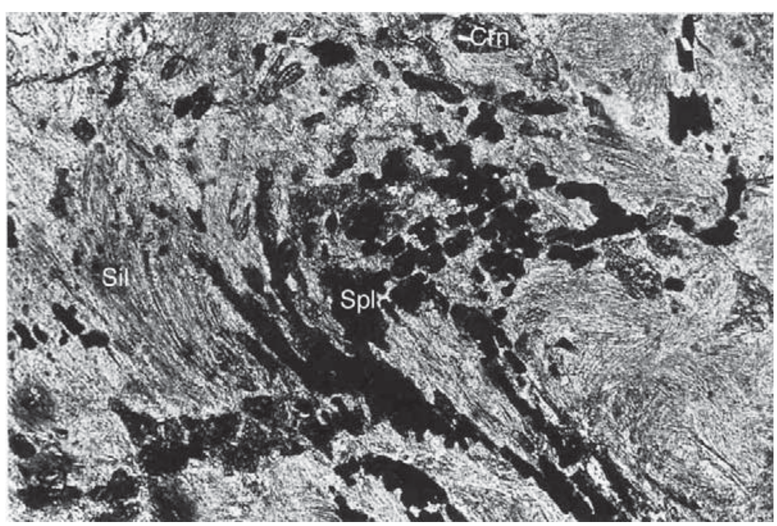

FIG. 7. Isoclinally folded aggregate of fibrous sillimanite (Sil), spinel (Spl) and corundum (Crn), interspersed with retrograde fine-grained white mica (lower relief) in metapelitic enclave, Cowra Granodiorite. Plane-polarized light; base of photo 4.4 mm.

isotopic compositions are broadly similar to the Ordovician–Silurian clastic sedimentary rocks of the LFB. SHRIMP U–Pb ages of inferred detrital zircon grains in the enclaves indicate Early Ordovician maximum ages of deposition for the psammite–pelite precursors, and U–Pb ages of zircon and Sm–Nd ages of garnet indicate that metamorphism of the metasedimentary enclave precursors was synchronous with the formation of the granite at about 430 Ma (Mass *et al.* 2001).

Thermobarometric estimates are 840°C and 0.55 GPa for early crystallized garnet–orthopyroxene aggregates in the Deddick Granodiorite (Maas *et al.* 2001b). In contrast, estimates of equilibration conditions for the metasedimentary enclaves are 820–750°C at similar or slightly shallower depths (Maas *et al.* 2001b), which indicates that the metasedimentary enclaves are xenoliths collected from above the melting zone for the granodioritic magma. This inference is supported by the large number of 600 Ma zircon ages (best represented in the Mid-Late Cambrian Kanmantoo Fold Belt of South Australia) in the granodiorite, compared with their poor representation in the metasedimentary enclaves. The implication is that the Deddick Granodiorite magma was formed in, or sampled during ascent through, crustal rocks older and probably deeper than the earliest Ordovician LFB rocks that were the source of the metasedimentary enclaves (Nicholls *et al.* 2001).

Because the Ca–Na contents are too low and the Nd–Sr isotopes too evolved for the metasedimentary enclaves to represent the only granite-source component, the general inference of Maas *et al.* (2001b) is that the S-type granitic magma received contributions from components richer in Ca and Na (as recognized by Wyborn & Chappell 1983), such as Cambrian volcanic-sedimentary "greenstone" successions, and from a more primitive Nd–Sr component (as recognized by McCulloch & Chappell 1982), such as contemporary mantle-derived magma.

The Pyalong Adamellite

Enclaves constitute <1% (locally >10%) of the total outcrop-area of the Pyalong Adamellite, which is part of the Cobaw Complex, Victoria, Australia (Fig. 1), as described by Anderson *et al.* (2001). Hornfels xenoliths are common near the contact, but rare elsewhere. Metasedimentary enclaves range from unmelted paragneisses to melt-depleted biotite-rich rocks (Maas *et al.* 2001b), and include quartz-rich and felsic schists, migmatites, quartzofeldspathic gneisses and calc-silicate rocks. Quartz fragments are also present. Biotite, sillimanite, cordierite, spinel and corundum occur in the dominant schistose varieties (Anderson *et al.* 2001). Their mineral assemblages are consistent with an origin in the middle crust at upper-amphibolite facies conditions.

The Sm–Nd isotopic data indicate that the metasedimentary enclaves do not represent the sole source-rocks of the host granite, as they have higher $^{87}Sr/^{86}Sr$ and lower ε_{Nd} than the host (Anderson 1997), though they conceivably could come from the source if the isotopic composition of the granite resulted from magma mixing. SHRIMP U–Pb data and the structures of the metasedimentary enclaves indicate that they were derived from unexposed and chemically distinctive Ordovician rocks that underwent a history of deformation different from that of the exposed country-rocks. They appear to be accidental xenoliths from an unexposed Ordovician terrane (Anderson *et al.* 2001).

General comment on metasedimentary enclaves in the southern LFB granites

In reviews of schistose metasedimentary enclaves in both S-type and I-type granites across the southern LFB, Fleming (1996), Anderson (1997) and Anderson *et al.* (1996, 1998) noted that the enclaves consist of fine- to medium-grained foliated biotite schist containing fibrous sillimanite and cordierite. Quartz, plagioclase, K-feldspar and garnet occur in some of the enclaves. Some are migmatitic, and deformational structures reflecting multiple deformation are common. Using ion-microprobe U–Pb isotopic data for zircon, Anderson *et al.* (1996, 1998) inferred that the enclaves are samples of the mid-crustal LFB Ordovician succession and, moreover, that such metasedimentary material contributed to the evolution of both the S-type and I-type granites, in view of their occurrence in both types.

Using chemical data for the associations Crd–Grt, Bt–Grt and Grt–Als–Pl–Qtz, Anderson (1997) calculated an approximate temperature of 750°C and a pressure of <500 MPa for the conditions of formation of the enclave assemblages. Anderson *et al.* (1998) inferred that the enclaves are from a mid-crustal source-region of granite, though they admitted the possibility of an accidental xenolithic origin. Fleming (1996) proposed that the metasedimentary enclaves preserve evidence of the structural history at the time of their incorporation in the granitic magma. However, this interpretation need not imply incorporation at the actual source of magma generation.

Metasedimentary Enclaves in Peraluminous Volcanic Rocks

Wyborn *et al.* (1981) mentioned the occurrence of gneissic "xenoliths" with garnet, cordierite, biotite, quartz and plagioclase, as well as quartz aggregates, in the Hawkins Volcanic suite, referred to previously. These enclaves were inferred to be restite.

Clemens & Wall (1984) described metasedimentary and microgranitoid enclaves in felsic ignimbrites of the Violet Town Volcanics, Victoria, Australia (Fig. 1). The metasedimentary enclaves are foliated with high-grade assemblages. Clemens & Wall (1984) inferred that they could be restite, but that they could also be xenoliths from the lower crust unrelated in origin to the ignimbrites.

Experimental Information on the Origin of Metasedimentary Enclaves

The inference of a restite-controlled origin for the chemical composition of LFB S-type granites (involving relatively low-temperature melting and variable retention of restite) is in conflict with abundant experimental evidence that volumetrically significant fluid-absent partial melting of appropriate crustal rocks occurs mainly at 850 to 950°C (Vielzeuf & Holloway 1988, Clemens & Wall 1981, 1984, Patiño-Douce & Beard 1995, Clemens 2003). This argument depends on the inference that euhedral phenocrysts in volcanic rocks are interpreted as products of magmatic crystallization, not restite; arguments in favor of a magmatic origin were presented previously.

Vielzeuf & Holloway (1988) showed that fluid-absent partial melting of metapelitic rocks at 850–875°C can produce 40% melt by the reaction: Bt + Als + Pl + Qtz = liquid + Grt (± Kfs), leaving a Qtz + Grt + Sil + Pl residue after segregation of the liquid. However, the experimental results of Green (1976), Clemens (1981), Clemens & Wall (1981) and Conrad *et al.* (1988) indicate that strongly peraluminous, pelitic enclaves are possible restite only for the most strongly peraluminous magmas, not for moderately peraluminous granites like the Cowra Granodiorite and other LFB contact-aureole S-type granites. Experiments show that these magmas are not in equilibrium with sillimanite in the ranges of 100 to 500 MPa, 700 to 900°C and 1 to 13% H_2O in the melt (Clemens & Wall 1981, Clemens 2003).

Experimental results of Clemens (1981) and Clemens & Wall (1981, 1984) on melt compositions typical of southeastern Australian S-type granites (*e.g.,* Strathbogie Batholith) and volcanic equivalents (*e.g.,* Violet Town Volcanics, Lake Mountain Rhyodacite) indicate temperatures of early crystallization of up to 950°C at 400 to 500 MPa and H_2O activities of 0.2 to 0.3. Thus, the magmas were hot and markedly H_2O-deficient (Clemens & Wall 1984). Inferred near-liquidus assemblages (850 to 950°C) in these rocks, together with the compositions of the rocks, indicate that the source was dominated by H_2O-poor, weakly to mildly peraluminous quartzofeldspathic, granulite-facies rocks, as suggested by Clemens & Wall (1981).

Clemens & Wall (1984) concluded that the markedly H_2O-deficient nature of the magma suggests that the melting reactions were fluid-absent, and involved the breakdown of biotite in quartz-saturated rocks to produce hydrous granitic melt plus granulite-facies residues containing minerals such as garnet, orthopyroxene, calcic plagioclase, quartz and possibly some biotite. Therefore, restite produced by partial melting of feldspathic metapsammite at experimentally determined temperatures suitable for producing volumetrically sufficient magma, namely 850 to 950°C, should consist of quartz, plagioclase, garnet and orthopyroxene, enclaves of which are typically absent from LFB S- type granites (Clemens 2001, 2003). Conceivably, these minerals could be mainly converted to lower-temperature minerals in the crystallizing magma if they survive the ascent. Quartz and calcic plagioclase restite, if present (possibly as cores in magmatic grains), may be difficult to recognize. Moreover, as mentioned previously, cores of calcic plagioclase may result from magma mixing or fractional crystallization, especially at high activity of H_2O (Allen 2001).

Experiments on granitic compositions (Winkler 1974, Winkler *et al.* 1975, Wyllie 1977, Clemens & Wall 1981) have indicated that with increasing temperature, Mg–Fe minerals become more soluble in the liquid, and that more mafic granitic melts form only at high temperatures. High-temperature fluid-absent melting produces melts with appreciable amounts of dissolved Mg, Fe and Ca, as mafic as granodiorite (Clemens 1984). Questions about whether realistic compositions of granite can crystallize from liquids are answered by experiments of Vielzeuf & Holloway (1988), which showed that liquids with 66.6 to 73.2% SiO_2, 1.4 to 7.1% FeO+MgO and 3.2 to 3.7% H_2O are produced by melting of pelitic compositions at 1 GPa and 875 to 1050°C, and by experiments of Rutherford *et al.* (1985), which showed that liquids with 62 to 76% SiO_2, 2.22 to 6.05% FeO+MgO and 1 to 8% H_2O are produced in 5 to 55 hours by melting of felsic compositions at 100 to 320 MPa and 875 to 1090°C. Patiño-Douce *et al.* (1995) showed that melts may vary widely in composition, depending on the pressure and H_2O activity.

The general conclusion derived from experimental information is that the LFB contact-aureole S-type granites crystallized from hot, relatively dry magmas formed at granulite-facies conditions.

Origin of Metasedmentary Enclaves in LFB Contact-Aureole S-Type Granites

A critical point relevant to the origin of the metasedimentary enclaves is the experimental evidence that melting responsible for the contact-aureole S-type granites occurred at about 850–950°C and 400 to 600 MPa. If this evidence is accepted, it means that any restite or resistate from the source rocks must also have been at these temperatures. However, assemblages in the metasedimentary enclaves reflect temperatures of about 750–800°C. These assemblages delineate strongly foliated, complexly microfolded and refolded microstructures developed at prograde amphibolite-facies conditions (*e.g.,* Clemens 2001). They do not resemble coarse-grained, granulite-facies aggregates that have undergone retrograde metamorphism. Furthermore, some of the metasedimentary enclaves contain andalusite (Chen *et al.* 1989, Maas *et al.* 1997). Therefore, a reasonable interpretation is that high-temperature restite or resistate is rare to absent, and that the metasedimentary enclaves come from a relatively high-grade region in the middle crust, possibly in the upper parts of the source terrane. The fact that some of them have compositions appropriate for restite (*e.g.,* Mass *et al.* 1997) does not mean that they are restite for the host granite. They may be from a higher-level source-volume that lost melt before being intruded and incorporated by the host magma. For example, the Cooma Granodiorite physically incorporated metasedimentary resistate xenoliths that had partly melted in an earlier event, and some of this material also found its way into plutons of the Murrumbidgee Batholith (Collins & Richards 2001). In addition, some metasedimentary enclaves contain domains of quartz–feldspar leucosome inferred to be the result of partial melting (Anderson *et al.* 1998, p. 120); this episode of melting must have occurred prior to incorporation of the enclaves, which consequently are xenoliths, not restite.

As noted by Clemens & Wall (1981), high-level S-type felsic magmas are typically not saturated with sillimanite, whereas the metasedimentary xenoliths commonly contain sillimanite, confirming that the xenoliths are not restite for the enclosing magma.

Also, the S-type granites may or may not contain other metasedimentary enclaves (*e.g.,* calc-silicate rocks or metaquartzites), which are resistate if they come from the source area, as suggested by White *et al.* (1977), or xenoliths if collected accidentally during ascent of the magma. Biotite-rich ("surmicaceous") and sillimanite-rich enclaves are consistent with restite (*e.g.,* White *et al.* 1991), but a resistate or xenolithic origin is difficult to exclude, especially as aggregates of these types are common in high-grade metamorphic rocks everywhere (*e.g.,* Vernon 1987).

Moreover, as pointed out by Clemens (2001, 2003), the metapelitic enclaves are typically biotite-rich, whereas biotite would have largely broken down in partial melting reactions if the enclaves represented restite. The biotite should also have mostly broken down if the metapelitic enclaves represent resistate from a high-temperature granulite-facies terrane.

An argument advanced by White *et al.* (1991) is that the metasedimentary enclaves must come from the source region, because they are absent from adjacent I-type granites. However, though this may be true for some areas, Anderson (1997) and Anderson *et al.* (1996) sampled metasedimentary enclaves from both I-type and S-type granites across the southern LFB. Moreover, even where the generalization of White *et al.* (1991) is applicable, it does not demand that the metasedimentary enclaves are from the actual source of melting, as they could be accumulated by ascending magma in parts of a metasedimentary terrane that may not have been accessed by the I-type granites.

Microgranitoid Enclaves in LFB S-Type Granites and Volcanic Rocks

General comments

Contact-aureole S-type granites in the LFB typically contain microgranitoid enclaves ("microgranular enclaves", "mafic enclaves") in small amounts (generally less than one or a few per cent of the pluton, except for local concentrations). Similar enclaves occur in some LFB S-type volcanic rocks, such as the Violet Town Volcanics, Victoria (Clemens & Wall 1984, Elburg 1996b, Clemens 2003). The microgranitoid enclaves mostly consist of microtonalite, with igneous microstructures, such as plagioclase laths projecting into quartz (White *et al.* 1991, p. 495-497), and a mineral assemblage appropriate to a peraluminous composition, namely orthopyroxene, biotite, cordierite, plagioclase and quartz (Figs. 8 to 20). Examples are discussed below.

Chen *et al.* (1989), Wyborn *et al.* (1991), White *et al.* (1991), Chappell & White (1991) and Chappell *et al.* (1993) have proposed that the microgranitoid enclaves are restite. This interpretation must be examined in the context of the present review.

I apply the term "microgranitoid" to this general class of igneous enclaves because it reflects their relatively small grain-size, igneous microstructure, and general range of granitoid compositions (Vernon 1983, 1984), embodying useful terms such as "microtonalite", "microadamellite" or "microgranodiorite" (*e.g.,* Phillips *et al.* 1981, Clemens & Wall 1984, Maas *et al.* 1997). White *et al.* (1991) preferred "microgranular", stating that "the fabrics of these enclaves are neither metamorphic nor are they granitic." However, the term "microgranitoid" was never intended to refer to the structures of coarse-grained granites or metamorphic rocks. Instead, it describes the microstructures of *micro*granites and related rocks. Even those who use "microgranular" in preference to "microgranitoid" are commonly obliged to use terms such as "microtonalite" and "microdiorite" (*i.e., microgranitoid* terms) when being more specific about the compositions of these enclaves (*e.g.,* Maas *et al.* 1997, Waight *et al.* 2000a, p. 1108, Clemens 2003). The term "microgranular" suggests a granoblastic or sugary microstructure without igneous connotations, and the term "mafic" is inappropriate, as the enclaves are commonly felsic to intermediate, and only rarely mafic.

Chappell & White (1991, p. 380) stated that microgranitoid enclaves in S-type granites "cannot be of magmatic origin." Chen *et al.* (1989, p.1214) stated that the peraluminous composition of the microgranitoid enclaves "rules out the possibility that they could represent crystallized blobs of an extraneous mingling magma such as basalt or andesite." Of course, this interpretation would hardly be reasonable, as the enclaves are typically quartz-bearing, not mafic. However, their peraluminous composition does not rule out the likelihood that the enclaves represent crystallized blobs of mingling *hybrid* magma, formed by the mixing of peraluminous felsic magma with more mafic magma, as discussed below.

Chappell *et al*. (2000) stated that microgranitoid enclaves in S-type granites "share distinctive compositional features with calcareous shales. They are therefore probably metamorphosed equivalents of such rocks." However, metamorphosed calcareous shales have been well studied, and typically consist of aggregates of calc-silicate minerals (such as diopside, grossular, epidote, actinolite and wollastonite) that have metamorphic microstructures. For example, calc-silicate xenoliths in the Cowra Granodiorite consist mainly of epidote, actinolite, plagioclase, quartz and calcite (Waight *et al*. 2001).

Chen *et al*. (1989), Chappell & White (1991) and White *et al*. (1999) speculated that these supposed "restite" enclaves contained a small amount of melt, which somehow converted a metamorphic into an igneous microstructure, forming plagioclase phenocrysts with oscillatory zoning, zoned plagioclase laths, interstitial quartz, highly elongate orthopyroxene, thin platy biotite, acicular apatite and ilmenite, skeletal zircon, and features characteristic of magma mixing, such as mantled quartz and feldspar xenocrysts, as discussed below. The mechanisms of this unlikely transformation were not explained.

The Cowra Granodiorite

As mentioned previously, the Cowra Granodiorite (Fig. 1) contains roughly equal proportions of microgranitoid enclaves and metasedimentary enclaves (Fig. 5). The microgranitoid enclaves (Figs. 8 to 15) are mainly peraluminous orthopyroxene microtonalite, with less common orthopyroxene microdiorite and biotite microtonalite. They have igneous microstructures, characterized by zoned plagioclase phenocrysts (Figs. 8, 9), orthopyroxene phenocrysts (Figs. 10, 11), local subophitic intergrowths of orthopyroxene and plagioclase, indicating simultaneous growth of these two minerals, elongate laths of plagioclase with oscillatory zoning in the groundmass (Figs. 8, 9, 11, 12, 14), very elongate crystals of orthopyroxene (Figs. 12, 13), acicular apatite, and a poikilitic microstructure of plagioclase, biotite and orthopyroxene crystals with respect to interstitial quartz (Figs. 8, 9, 11, 12, 14). This poikilitic relationship is very common in microgranitoid enclaves of undoubted magmatic origin in I-type granites (Vernon 1983, 1984, 1991, Figs. 1, 2; Collins *et al*. 2000, Fig. 3f). Electron-microprobe analyses (R.H. Vernon, R.H. Flood and W.F. D'Arcy, unpubl. data) show a compositional range of An_{73} to An_{40} in plagioclase, and Waight *et al*. (2001) reported a range of An_{63} to An_{35}. Extremely calcic compositions of groundmass plagioclase (An_{80} or even more calcic) were reported by White *et al*. (1991, p. 501), and up to An_{94} was reported by Wyborn *et al*. (1991). Biotite appears to have partly to completely replaced orthopyroxene in

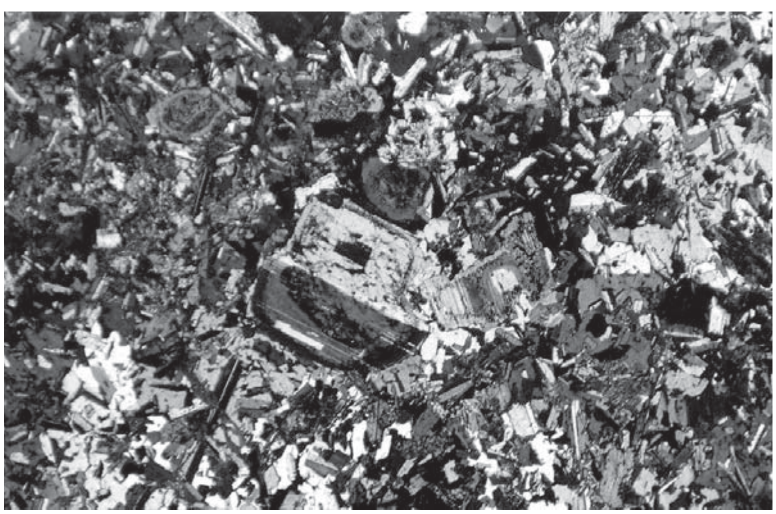

FIG. 8. Microgranitoid (microtonalite) enclave in the Cowra Granodiorite, showing an igneous microstructure, consisting of phenocrysts of zoned plagioclase, commonly with a corroded and altered core, scattered through a groundmass consisting of elongate crystals of plagioclase and orthopyroxene set poikilitically in quartz. Crossed polars; base of photo 4.4 mm.

FIG. 9. Microgranitoid (microtonalite) enclave in the Cowra Granodiorite, consisting of phenocrysts of zoned plagioclase scattered through a groundmass consisting of elongate crystals of plagioclase, orthopyroxene and biotite set poikilitically in quartz. The plagioclase phenocrysts show a corroded core, with magmatic fillings and an overgrowth in optical continuity (lighter grey). This "patchy zoning" (Vance 1965) can be explained by magma mixing (Hibbard 1981, Vernon 2004). Small inclusions of orthopyroxene in the plagioclase filling probably crystallized from melt trapped in the corroded cores (Vance 1962). Many of the orthopyroxene crystals are very elongate. Plane-polarized light; base of photo 4.4 mm.

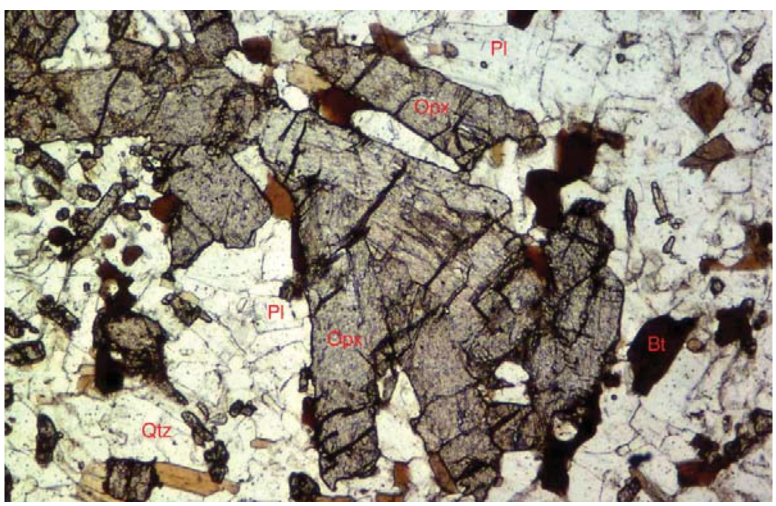

FIG. 10. Phenocrystic aggregate of orthopyroxene (Opx) in a microtonalite enclave in the Cowra Granodiorite. Plane-polarized light; base of photo 1.75 mm.

some enclaves, especially in the outer parts, apparently as a result of reaction of the orthopyroxene with the host magma. Some cordierite occurs as interstitial grains intergrown with quartz, plagioclase and orthopyroxene (Fig. 14), and so evidently can be a member of the primary mineral assemblage; it also occurs as scattered large grains of uncertain origin (Fig. 15).

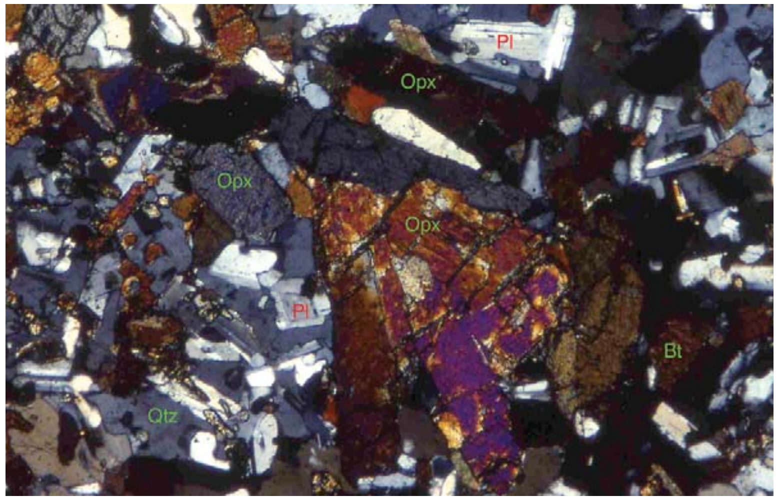

FIG. 11. Same field of view as for Figure 10, showing laths of plagioclase (Pl) in quartz (Qtz). Crossed polars; base of photo 1.75 mm.

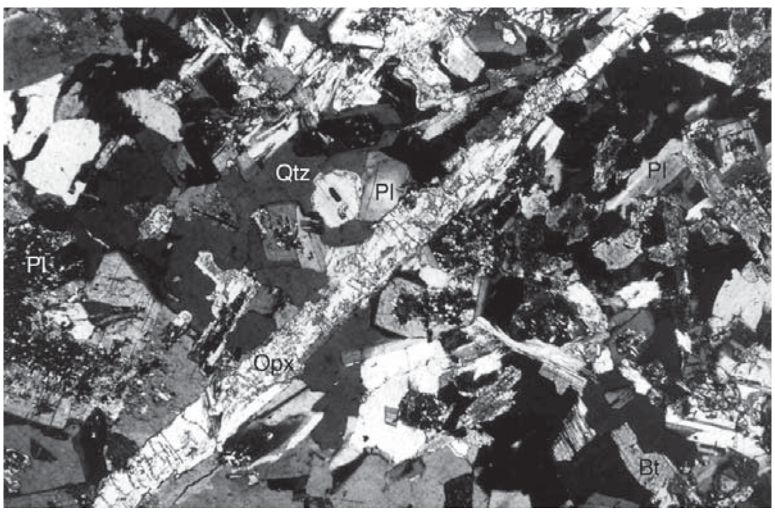

FIG. 12. Euhedral phenocryst of plagioclase (Pl; bottom-left) and a very elongate phenocryst of orthopyroxene (Opx) in a microtonalite enclave in the Cowra Granodiorite. Also shown are euhedral to subhedral laths of zoned plagioclase, with some biotite, in interstitial quartz (Qtz). Crossed polars; base of photo 4.4 mm.

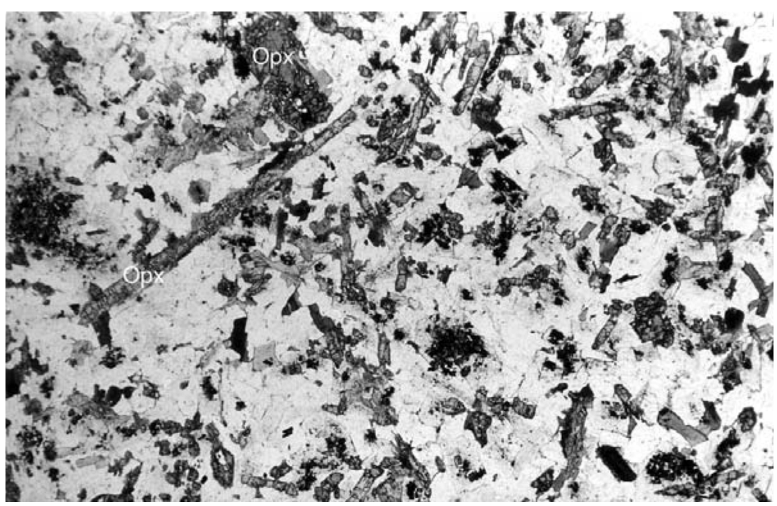

FIG. 13. Microtonalite enclave in the Cowra Granodiorite, showing abundant, very elongate crystals of orthopyroxene (Opx). Plane-polarized light; base of photo 12 mm.

The microgranitoid enclaves also show strong evidence of having been composed of mixed (hybrid) magma before mingling with the present host magma to form the enclaves. This evidence includes phenocrysts of mantled plagioclase, some of which have a corroded core (Fig. 9), corroded xenocrysts of cordierite from the peraluminous felsic magma, and xenocrysts ("ocelli") of quartz mantled with orthopyroxene and biotite precipitated from the hybrid magma, in response to dissolution of the quartz (Figs. 16, 17), as described by Vernon (1983, 1990, 1991). These "ocelli" are identical in appearance to orthopyroxene-mantled quartz xenocrysts in glassy (and hence magmatic) orthopyroxene-bearing microgranitoid enclaves in S-type volcanic rocks (Vernon 1983, 1991, 2004). They are the peraluminous equivalents of quartz xenocrysts mantled with hornblende in I-type microgranitoid enclaves (Fig. 18), and are characteristic products of magma mixing (Vernon 1983, 1990, 1991, 2004, Wiebe *et al.* 2002).

Waight *et al.* (2001) and Nicholls *et al.* (2001) showed that the microgranitoid enclaves have isotopic compositions ($^{87}Sr/^{86}Sr$ and ε_{Nd}) generally more primitive than, or similar to, those of the host granite, and that the high CaO contents and relatively high ε_{Nd} of the microgranitoid enclaves are not found in either the Palaeozoic LFB turbidites or the metapsammitic–pelitic enclaves in LFB granites. Thus, the Cowra microgranitoid enclaves are not metasedimentary, and the isotopic evidence supports the suggestion of Vernon (1983) that they result from the mingling of hybrid magma formed by mixing of peraluminous felsic magma and more mafic magma. The mixing must have occurred deeper in the crust than the level of granite emplacement, as no *in situ* mixing effects have been observed, although some minor chemical diffusion between enclave and host commonly has occurred, *e.g.,* biotite replacing orthopyroxene. The isotopic variability, lower $^{87}Sr/^{86}Sr$ and higher ε_{Nd} of the microgranitoid enclaves, compared with the host granite, are probably the result of isotopic equilibration with an initially isotopically more primitive mafic magma (Waight *et al.* 2001).

Wyborn *et al.* (1991) stated that the Cowra microgranitoid enclaves do not have igneous compositions and are probably restitic. White *et al.* (1991) acknowledged that the microgranitoid enclaves "have fabrics suggestive of igneous rocks (pseudo-doleritic fabric)", but asserted that they are restite (resistate in my terminology), in the form of partially melted calcareous metasediments. The melt was inferred to have remained in the calc-silicate enclaves, converting the original metamorphic microstructure to a wholly igneous one, which is an improbable process, as mentioned above. A partially melted calc-silicate origin is particularly unlikely, in view of the occurrence in the Cowra Granodiorite of pale green calc-silicate xenoliths that show no evidence of melting, but instead contain normal calc-silicate minerals, such as calcite, clinozoisite, epidote, actinolite and titanite (White *et al.* 1991, p. 500, Waight *et al.* 2001). Moreover, the presence of some double enclaves (Fig. 19), consisting of either sharply bordered cores of dolerite in microtonalite or of gneissic or hornfelsic metasedimentary xenoliths in microtonalite, confirm that the microtonalite is not restite. Furthermore, Waight *et al.* (2000a) showed that a Cowra calc-silicate enclave has the low Rb/Sr value expected of a Ca-rich marl, and $^{87}Sr/^{86}Sr$ and ε_{Nd} in the range of LFB metapsammitic–metapelitic enclaves, contrary to the situation in the microgranitoid enclaves, as noted above.

The Deddick Granodiorite

Maas *et al.* (1997) and Nicholls *et al.* (2001) described microgranitoid (mainly microtonalite) enclaves in the S-type Deddick Granodiorite, Kosciusko batholith (Fig. 1), noting their igneous microstructure, apatite needles, mantled quartz xenocrysts ("ocelli"), xenocrysts of plagioclase and cordierite, and magnesian orthoyproxene phenocrysts compatible with crystallization from a high-Mg mafic to intermediate magma. Plagioclase phenocrysts are of similar size and composition (An_{50}) to plagioclase in the host granodiorite. Plagioclase laths in the groundmass were reported to have a compositional range of An_{80} to An_{50} (Maas *et al.* 1997). Orthopyroxene occurs as zoned phenocrysts, glomeroporphyritic aggregates and euhedral crystals in the groundmass. Acicular apatite also is present in the groundmass.

Large grains of plagioclase in the microgranitoid enclaves are commonly rimmed by more calcic plagioclase ($An_{80–85}$). Some plagioclase grains have core-to-rim transition zones riddled with melt inclusions, of the type produced by experimental reaction of sodic plagioclase with calcic melt (Tsuchiyama 1985). Corroded grains of quartz up to 5 mm across, with prominent orthopyroxene–plagioclase reaction rims (quartz "ocelli") occur in most of the microgranitoid enclaves. All these features indicate that the enclave magmas were hybrids formed by mixing of peraluminous felsic magma with more mafic magma, before final mingling with the host granodiorite to form enclaves.

Isotope data (Sr and Nd) are consistent with an origin of the enclaves as globules of hybrid magma, preserving evidence of arrested hybridization (Maas *et al.* 1997, Nicholls *et al.* 2001). Maas *et al.* (2001) also inferred that in view of the occurrence of microgranitoid enclaves in many S-type granites, along with the regional-scale isotope heterogeneity of the granites, a contribution from mafic magma may be of major importance in the generation of S-type granitic magmas of the LFB. Maas *et al.* (1997) noted that magnesian–orthopyroxene-bearing andesites have been reported from Japan.

The Strathbogie Batholith

The Strathbogie Batholith, Victoria, Australia (Fig. 1), contains microgranitoid (microadamellite) enclaves with igneous microstructures, composed of quartz and K-feldspar anhedra with inclusions of euhedral laths of plagioclase (An_{43}) and biotite plates (Phillips *et al*. 1981). Phillips *et al*. inferred that the enclaves formed as "accumulations of small, early-formed biotite, plagioclase and cordierite crystals in interstitial liquid that crystallized as quartz and orthoclase", ostensibly representing a chilled phase of the granite dislodged from the walls or roof. However, though such quench cumulates may exist, especially in the absence of evidence for the presence of mafic magmas coeval with granites and of isotopic contrast between host and microgranitoid enclaves (Donaire *et al*. 2005), arguments presented below counter this suggestion for the LFB granites.

Wilson's Promontory Batholith

Microgranitoid enclaves occur in the S-type Wilson's Promontory Batholith, southeastern Victoria (Fig. 1), as described by Elburg & Nicholls (1995) and Elburg (2001). They occupy less than 1 vol.% of the pluton, but locally form swarms of enclaves, together with some hornfels xenoliths. They have poikilitic to equigranular microstructures, are finer-grained than the host granite, and contain more biotite and less garnet. Cordierite is rare. Some have interlocking networks of platy biotite and tabular crystals of plagioclase in a groundmass of quartz or, more rarely, K-feldspar, forming a typical igneous microstructure. Most of the microgranitoid enclaves contain megacrysts of K-feldspar, plagioclase or quartz, many of which have an overgrowth of higher-temperature minerals. For example, K-feldspar megacrysts generally have an overgrowth of plagioclase, rounded quartz xenocrysts have a rim of fine-grained biotite, and plagioclase megacrysts have a more calcic overgrowth, such as a core of An_{45} with an overgrowth of $An_{70–80}$ (Elburg 1996c). These features are characteristic of magma mixing.

Acicular apatite is common (Elburg & Nicholls 1995), elongate crystals of zircon are present, and some zircon overgrowths have a skeletal habit (Elburg 1996c). All these microstructures indicate rapid growth in a supersaturated liquid (Wyllie *et al*. 1962, Vernon 1983, 2004, Bossart *et al*. 1986). The microgranitoid enclaves mostly have lower initial Sr and higher Nd isotopic values than the host granite (Elburg 2001). The microstructural, chemical and isotopic data are best explained by mixing of felsic magma with a more mafic magma (Elburg & Nicholls 1995).

Some of the microgranitoid enclaves have larger crystals of zircon with inherited ages, similar to those in the host granite. These crystals were evidently introduced during the magma mixing that was responsible for introducing xenolithic quartz and feldspar into the enclaves. This mixing occurred before crystallization of most of the magmatic zircon crystals in the host granite (Elburg 1996c). Rare microgranitoid enclaves have zircon with no inheritance, confirming that these enclaves cannot be restitic (Elburg 1996c).

The microgranitoid enclaves have undergone varying degrees of chemical and isotopic interaction with the host magma, but it is uncertain whether the least re-equilibrated enclaves represent original compositions of enclave magma (Elburg 1996a).

The Warburton Granodiorite

Elburg (1996a, 2001) described microgranitoid enclaves with zoned plagioclase crystals, acicular to skeletal apatite, elongate zircon and quartz oikocrysts enclosing plagioclase and biotite crystals in the Warburton Granodiorite, Victoria, Australia (Fig. 1). Some of the enclaves contain plagioclase megacrysts with an anhedral core similar in composition to plagioclase in the host granodiorite (An_{45}), the core being mantled by calcic plagioclase ($An_{70–80}$), which was also reported to be the composition of plagioclase in the enclave groundmass. Rare enclaves have rounded megacrysts of quartz ("ocelli") rimmed by fine-grained mafic minerals. Both the mantled plagioclase and quartz megacrysts are characteristic of magma mixing, as discussed previously.

Many of the enclaves contain biotite as the only mafic mineral, but some have orthopyroxene in their core, amphibole in a surrounding mantle, and biotite in the rim, owing to reaction between the enclave and host magma (involving diffusion between residual liquids in both enclave and host). Three out of four microgranitoid enclaves analyzed have lower initial $^{87}Sr/^{86}Sr$ and higher ε_{Nd} values than the host granite. Some microgranitoid enclaves contain low-grade country-rock xenoliths ("double enclaves"), as shown in Figure 19. This observation, together with the similarity in age between the crystals of magmatic zircon in the enclaves and host, also imply that the enclaves are not restitic (Elburg 2001). Moreover, the lack of isotopic equilibrium between the enclaves and the host granodiorite indicates that the enclaves do not represent accumulations of early-crystallized minerals from the granodiorite (Elburg 1996a). The high Cr content of some enclaves suggests that the parental magma of the enclave material is mantle-derived. This inference, together with the rimmed plagioclase and quartz xenocrysts, as well as the isotopic evidence, suggest that the enclave magmas were hybrids formed by mixing of S-type felsic magma with more mafic magma.

The Violet Town Volcanics

The ignimbrites of the Violet Town Volcanics, Victoria (Fig. 1), contain microgranitoid enclaves

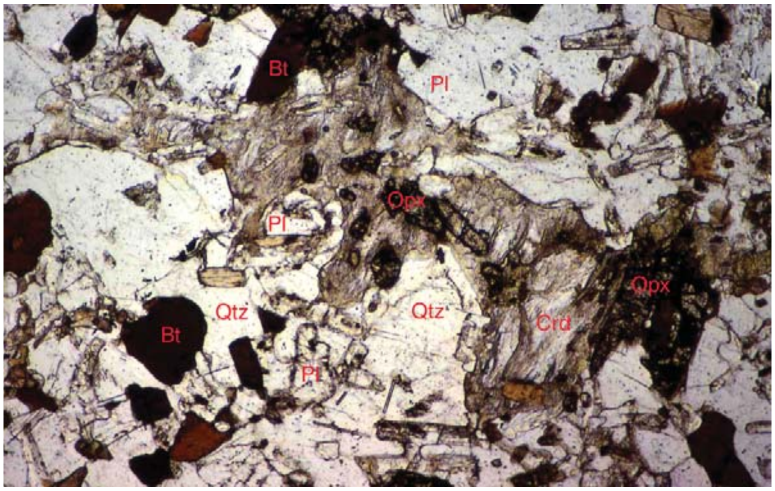

FIG. 14. Interstitial retrograded cordierite (Crd) with inclusions of plagioclase (Pl), orthopyroxene (Opx) and biotite (Bt) in a microtonalite enclave in the Cowra Granodiorite. Also present are plagioclase inclusions in quartz (Qtz) and needles of apatite. Plane-polarized light; base of photo 4.4 mm.

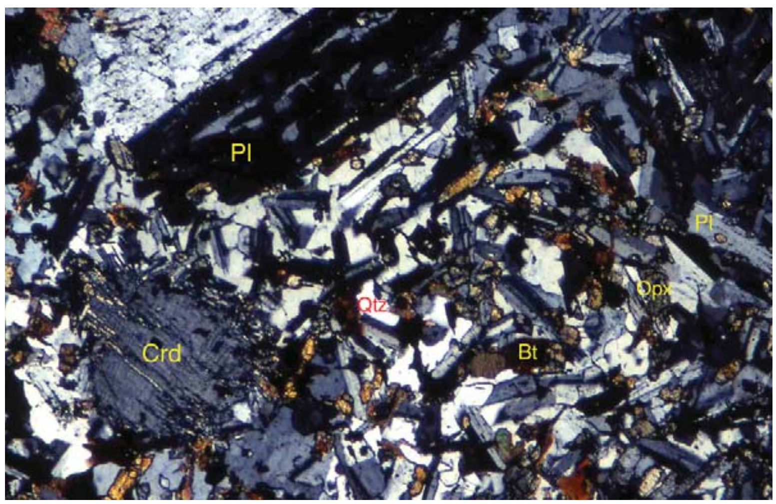

FIG. 15. Microtonalite enclave in the Cowra Granodiorite, showing a phenocryst of plagioclase (Pl) and a large grain (phenocryst or xenocryst) of anhedral cordierite (Crd) in a groundmass consisting of elongate crystals of plagioclase (Pl), biotite (Bt), orthopyroxene (Opx) and interstitial quartz (Qtz). Crossed polars; base of photo 4.4 mm.

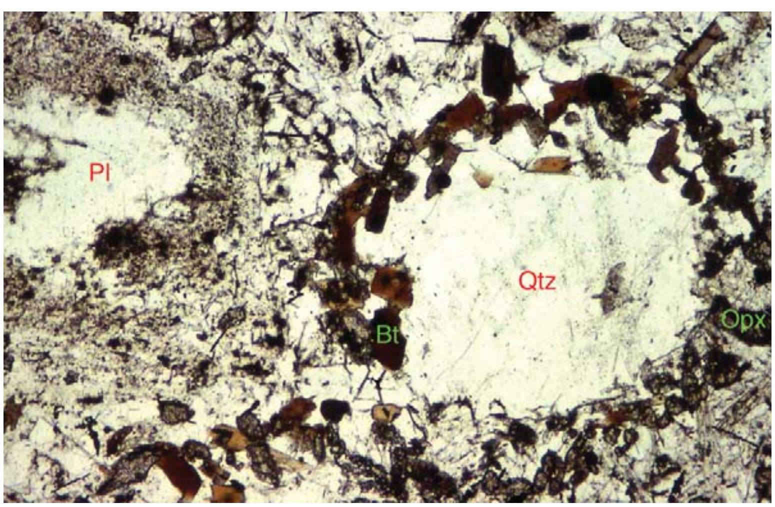

FIG. 16. Microtonalite enclave in the Cowra Granodiorite, showing a phenocryst of corroded plagioclase (Pl) and a fragment of quartz (Qtz) mantled with biotite (Bt) and orthopyroxene (Opx), constituting a mantled xenocryst or "ocellus" formed by magma mixing. Plane-polarized light; base of photo 4.4 mm.

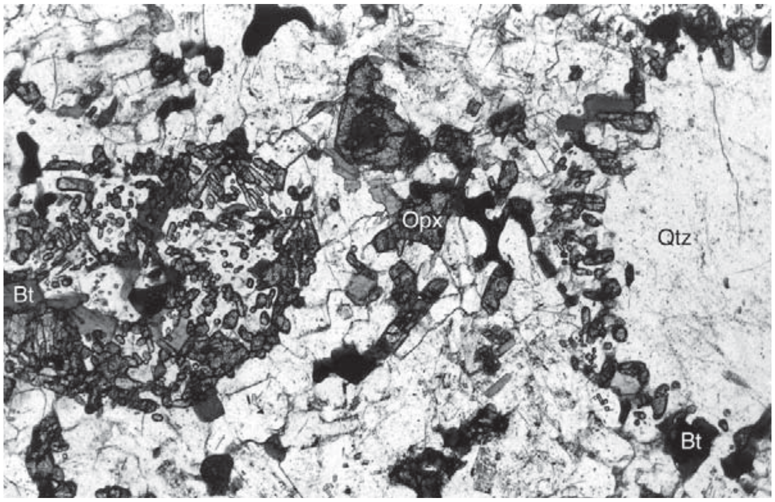

FIG. 17. Microgranitoid (microtonalite) enclave in the Cowra Granodiorite, showing evidence of magma mixing, in the form of two quartz xenocrysts ("ocelli") rimmed by fine-grained orthopyroxene-rich aggregates with some biotite. The example at right shows a xenocryst and rim, whereas the example at left shows a rim that has been intersected by the section. These are analogous to quartz xenocrysts ("ocelli") rimmed by hornblende in microgranitoid enclaves in I-type granites. Plane-polarized light; base of photo 4.4 mm.

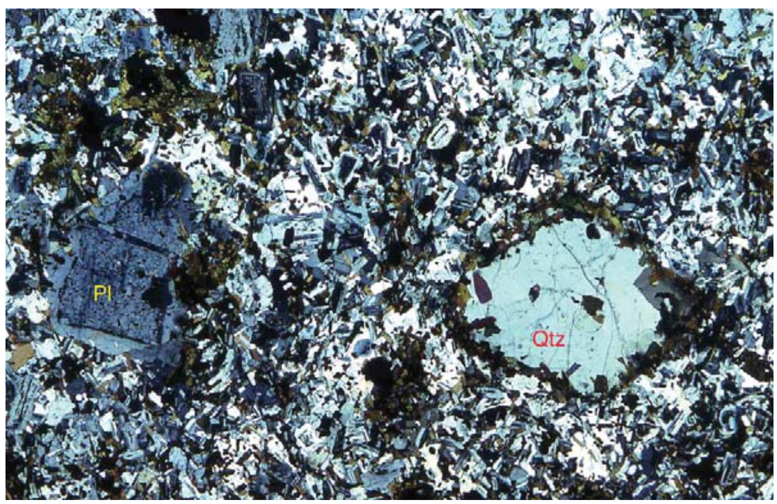

FIG. 18. Microgranitoid enclave in an I-type granite, New England Batholith, New South Wales, showing a plagioclase phenocryst (Pl) and a quartz (Qtz) xenocryst ("ocellus") rimmed by hornblende and biotite. The groundmass consists of elongate crystals of zoned plagioclase, biotite and hornblende, with interstitial quartz. Crossed polars; base of photo 8.5 mm.

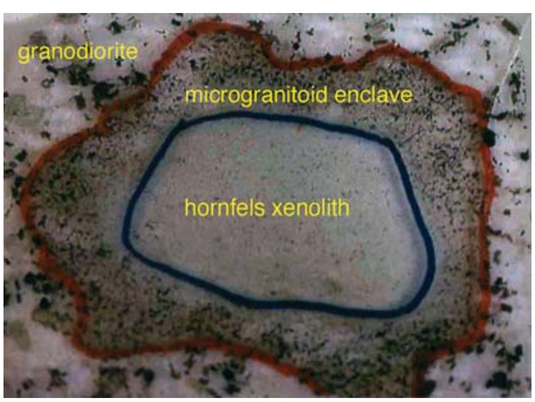

FIG. 19. Double enclave in the Cowra Granodiorite, consisting of a low-grade hornfels core and a microtonalite rim. Base of photo 4 cm.

(Clemens & Wall 1984, Elburg 1996b). These enclaves have igneous microstructures (Fig. 20), and are peraluminous microtonalites (with high orthopyroxene : biotite ratio) and microadamellites (with a low orthopyroxene : biotite ratio). The plagioclase, orthopyroxene and biotite crystals are typically elongate, and are enclosed poikilitically in quartz (Fig. 20). The enclaves resemble microgranitoid enclaves in the S-type granites, and the plagioclase has a similar range in zoning ($An_{60–44}$), rarely as calcic as An_{84} (Elburg 1996b). The plagioclase laths show oscillatory zoning (Elburg 1996b). Rare megacrysts of plagioclase, orthopyroxene and biotite are present. The plagioclase phenocrysts commonly have a corroded core with a composition similar to the plagioclase in the host ignimbrite, surrounded by a more calcic overgrowth with oscillatory zoning (Elburg 1996b), which is a typical result of magma mixing. As in the Cowra Granodiorite, zoned microgranitoid enclaves have more biotite in the rim, owing to reaction with the host magma.

Clemens & Wall (1984) inferred that the microgranitoid enclaves are accumulations of early-crystallizing minerals detached from the walls of the magma chamber. However, the enclaves have lower initial $^{87}Sr/^{86}Sr$ and higher ε_{Nd} values than the host ignimbrite, and also have corroded xenocrysts of plagioclase with a more calcic overgrowth, which is a typical feature of magma mixing (*e.g.*, Vernon 1990, 1991, 2004). Therefore, the microgranitoid enclaves appear to represent mingled globules of mafic, mantle-derived magma hybridized by mixing and diffusional exchange with felsic magma (Elburg 1996b). Trace-element contents of the plagioclase and orthopyroxene in microgranitoid enclaves that have reacted to a relatively small extent with the host magma suggest that the enclave magmas were similar to the magma that formed the basaltic andesite xenoliths that also are present in these rocks, giving some clue as to the nature of the mafic parent (Elburg 1996b).

Origin of Microgranitoid Enclaves in LFB S-Type Granites

As discussed above, the igneous microstructures, evidence of magma mixing in the form of mantled quartz and feldspar xenocrysts, isotopic characteristics, and double enclaves with a metasedimentary xenolith core, indicate that the microgranitoid enclaves in the southeastern Australian S-type granites are neither restite nor resistate. They are best interpreted as resulting from magma mixing and subsequent mingling, on the basis of microstructural, chemical and isotopic evidence (*e.g.*, Vernon 1983, 1984, 1990, 1991, Elburg & Nicholls 1995, Elburg 1996a, c, Maas *et al.* 1997, Waight *et al.* 2000a, b).

Peraluminous microtonalitic enclaves with plagioclase phenocrysts, elongate crystals of orthopyroxene and orthopyroxene-mantled quartz xenocrysts in S-type volcanic rocks (Fig. 20) are similar to microtonalite enclaves in the S-type granites. However, they may be glassy in young volcanic rocks (*e.g.*, in the Hoyazo area, Spain), which proves that they were magma globules, the glass having been preserved by chilling in contact with the atmosphere (Vernon 1983, 1991).

Some investigators have interpreted the microgranitoid enclaves as being disrupted chilled margins or accumulations of early-formed crystals that originally formed at the margins of the pluton (*e.g.*, Phillips *et al.* 1981, Clemens *et al.* 1984). However, Chappell (1996) pointed out that such a mechanism is inconsistent with the observation that microgranitoid enclaves have different patterns of zircon inheritance from that of their host granite. In addition, Elburg & Nicholls (1995) and Waight *et al.* (2001) have shown that the microgranitoid enclaves are distinct from the host rocks with respect to other isotopes. Furthermore, the evidence of magma mixing also is inconsistent with this mechanism, as is the local occurrence of enclaves with chilled margins. Moreover, as pointed out by Elburg & Nicholls (1995), the experiments of Clemens (1981) have shown that neither changes in pressure nor H_2O content can change the crystallization sequence of an S-type magma in such a way that it reproduces the various observed disequilibrium microstructural relationships between plagioclase and K-feldspar in the microgranitoid enclaves.

A logical conclusion is that the microgranitoid enclaves in the southeastern Australian S-type granites are former globules of hybrid magma that were mingled with felsic magma, in the same way as equivalent microgranitoid enclaves in I-type magmas worldwide (*e.g.*, Vernon 1983, 1984, 1990, 1991, 2004, Reid *et al.* 1983, Vernon *et al.* 1988, Waight *et al.* 2000a). They show similar hybrid features, such as mantled feldspar and quartz xenocrysts. For example, quartz xenocrysts in microgranitoid enclaves in I-type granites have a rim of hornblende (Fig. 18), whereas quartz xenocrysts in microgranitoid enclaves in S-type granites have a rim of orthopyroxene or biotite (Figs. 16, 17).

An important question remains: where did the magma mixing responsible for the enclave magmas occur? The occurrence of double enclaves consisting of hornfels or other metasedimentary xenoliths enclosed in microgranitoid enclaves (*e.g.*, Elburg 2001, Waight *et al.* 2000b) indicates that the mixing occurred below the level of emplacement of the host pluton, and that the enclave magma incorporated the xenoliths during its ascent or emplacement. Local chilled margins, as well as intricate shapes, such as thin tails and indented boundaries (*e.g.*, Elburg & Nicholls (1995, p. 432), reflecting viscous flow at high temperatures (Williams & Tobisch 1994), indicate that the enclaves were in a magmatic state after their incorporation as magma globules (mingling). Therefore, many of the enclaves were incorporated as already mixed (hybrid) magmas, independently of the host felsic magma, after which mingling and some chemical interaction with the host magma took place. However, the presence of K-feldspar megacrysts in microgranitoid enclaves in some granites is consistent with some mixing at a relatively late stage, when the felsic magma had cooled to 750 to 770°C after ~40% crystallization, according to experimental results of Clemens (1981), as discussed by Elburg & Nicholls (1995); this situation implies relatively high-level mixing for these particular enclaves.

Though some mixing between locally derived felsic magma and more mafic magma can occur in migmatite terranes (*e.g.,* Kemp & Gray 1999, Kemp 2001, 2004), the most effective places for such mixing to occur would be in deeper, hotter zones, in which mafic magmas are more abundant (Vernon 2005). Possible, though somewhat speculative, sources are zones of melting, assimilation storage and homogenization ("MASH zones") at the base of the crust, as suggested by Hildreth & Moorbath (1988) on the basis of extensive isotopic evidence (see below).

Source of High-Level S-Type Granitic Magmas

As discussed previously, experimental, chemical and isotopic evidence indicates that exposed Ordovician metapelites are unsuitable as source rocks for the LFB contact-aureole S-type granites (Clemens & Wall 1981, 1984, Price 1983, Chappell 1984, Clemens 1988), although the Ordovician feldspathic metapsammites are the evident source for LFB regional-aureole granites, such as the Cooma Granodiorite. More Ca and Na are needed for the contact-aureole S-type granites, implying either that quartzofeldspathic rocks form a basement to the Ordovician succession (*e.g.,* Clemens 1988) or that mafic magma was added. Thermobarometry based on phenocrysts in S-type volcanic rocks (Clemens & Wall 1984) indicates early crystallization of LFB S-type granitic magma at 12 to 16 km, whereas metasedimentary enclaves in the same rocks contain mineral assemblages equilibrated at 19 to 23 km. This evidence suggests that after collecting the xenoliths, the crystal-poor magma rapidly ascended without further appreciable crystallization.

Large, more mafic high-level granites may require deep, hot sources, involving mafic magma, rather than mid-crustal migmatite complexes (*e.g.,* Vernon 2005). Even the small, relatively mafic Cooma Granodiorite, derived partly from local metapsammite-derived neosome contaminated with disrupted, solid metapelite migmatite, probably required addition of a similar deeper magma (Vernon *et al.* 2003); this magma may have been similar to magmas that formed the contiguous, large, closely related Murrumbidgee Batholith, which required major transfer of magma from a deeper source than the Cooma migmatites (Collins & Richards 2001, Collins *et al.* 2002). Similarly, migmatitic Delamerian granites in southeastern South Australia are intruded by I–S granites formed by basalt–sediment interaction at greater depth (Foden *et al.* 2002). In the Arunta Block, central Australia, some high-level plutons were fed by observed migmatites, but the largest, hottest plutons have a deeper, unobserved source (Collins & Sawyer 1996).

Phillips *et al.* (1981), Clemens (1988, 2003, 2005), Vielzeuf *et al.* (1990), Vernon *et al.* (1990, 1993), Clarke *et al.* (1990), Collins & Vernon (1991, 1992, 1994), Finger & Clemens (1995), Keay *et al.* (1997) and Vigneresse (2004) suggested that mantle-derived mafic magma is necessary as a heat source for partial melting in low-pressure – high-temperature metamorphic terranes, such as the LFB. Some of these authors suggested that the mafic magma contributed to hybrid magmas (*e.g.,* Keay *et al.* 1997, Collins 1996, 1998). On the other hand, in some areas elsewhere in the world, the inferred mafic magma shows no evidence of having been mixed or mingled with felsic magma (Finger & Clemens 1995). Moreover, some other areas show no clear evidence of mafic magma as a heat source, despite the formation of granulite-facies assemblages (Villaseca *et al.* 1999).

The unobserved deep sources that are typical of most larger high-level plutons may not be conventional migmatite complexes, which tend to produce relatively leucocratic granites, and which may not be capable of supplying magma continuously enough to keep feeder dykes active. Perhaps these sources could be "MASH" zones, as suggested by Collins (1996) and Vernon (2005). A considerable amount of isotopic evidence for such zones exists in some magmatic arcs (Hildreth & Moorbath 1988). Though exposures of undoubted MASH zones have not yet been found, Hopson & Mattinson (1994) suggested that the Chelan Complex, Washington, U.S.A., may be one. These complex zones may be expected to show: (1) a high proportion of mafic rock, (2) a high proportion of melting, (3) rocks representing hot felsic magmas, (4) major depletion in melt, (5) evidence of repeated injection of mafic magma, and (6) evidence of repeated magma-mixing. An important question is: can a MASH zone produce and store enough suitable magma to eventually form a dyke of critical width that can drain the zone rapidly enough to form a major pluton in the upper crust?

The upper part of a possible source-region for S-type granitic magmas may be represented by the Hidaka Metamorphic Belt, Hokkaido, northern Japan (Fig. 21), where sheets of high-temperature S-type, garnet–orthopyroxene tonalite have intruded and been deformed in granulite-facies rocks (orthogneisses and paragneisses) in island-arc crust at the base of a tilted crustal section (Shimura *et al.* 1992, 2004). The deepest exposed rocks are garnet – cordierite – biotite gneiss, garnet – orthopyroxene gneiss, orthopyroxene – cordierite gneiss, orthopyroxene – plagioclase gneiss and orthopyroxene amphibolite, all cut by anatectic leucosome with euhedral orthopyroxene. Garnet – cordierite – sillimanite – biotite gneisses contain domains of biotite-free leucosome with euhedral orthopyroxene, euhedral to subhedral plagioclase and quartz, and local cordierite (Owada *et al*. 2003, Fig. 20); mafic rocks contain domains of orthopyroxene – plagioclase – quartz leucosome (Shimura *et al*. 1999). The pressure estimated from the garnet–cordierite geobarometer is 500–700 MPa, and the temperature estimate is 870°C (Osanai *et al*. 1992).

In the Niikappu River part of the Hidaka Metamorphic Belt, homogeneous S-type tonalite (consisting of garnet, orthopyroxene, cordierite, biotite, plagioclase and quartz with minor K-feldspar) has been inferred to represent original magma formed by partial melting of granulite-facies metasedimentary rocks (Shimura *et al*. 2004). Heterogeneous, S-type tonalite consisting of leucocratic patches (K-feldspar, plagioclase and quartz) interspersed with melanocratic aggregates (garnet, orthopyroxene, biotite, plagioclase, quartz), occurring around, and as veins in, metasedimentary enclaves, appear to be the products of assimilation processes. Furthermore, S-type tonalite composed mainly of garnet, orthopyroxene, biotite, plagioclase and quartz

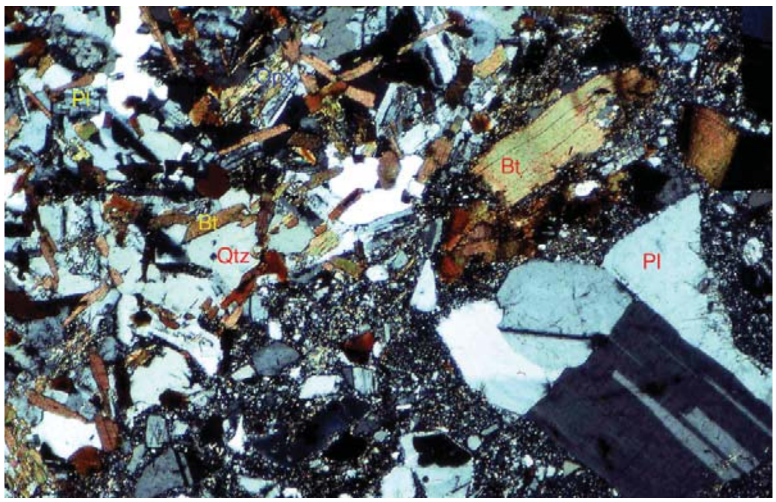

FIG. 20. Microgranitoid (microtonalite) enclave (top left half of photo) in an S-type ignimbrite of the Violet Town Volcanics, Victoria. The microtonalite consists of elongate crystals of plagioclase (Pl), biotite (Bt) and orthopyroxene (Opx) set in quartz (Qtz). Crossed polars; base of photo 4.4 mm.

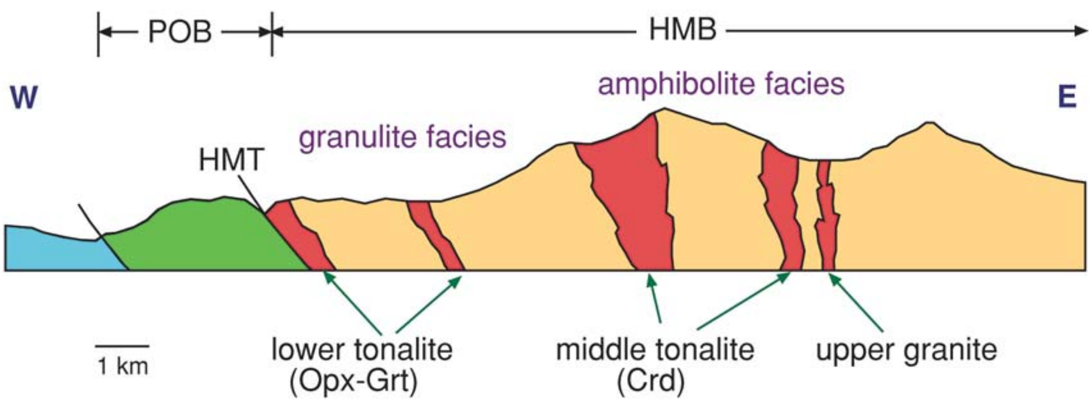

FIG. 21. Diagrammatic cross-section (after Shimura *et al*. 1992, 2004) of the Hidaka Metamorphic Belt, Hokkaido, Japan, showing sheets of S-type tonalite, with orthopyroxene – garnet in the granulite facies and cordierite in the amphibolite facies. POB: Poroshiri Ophiolite Belt, HMB: Hidaka Metamorphic Belt, HMT: Hidaka Main Thrust (base of the exposed HMB).

(cordierite and K-feldspar being rare) has been interpreted as containing cumulus plagioclase, garnet and some orthopyroxene; small melanocratic clots of orthopyroxene or garnet (or both), possibly restitic, are also present (Shimura *et al.* 2004).

As well as garnet–orthopyroxene tonalites, abundant gabbros and diorites also intrude the metamorphic rocks in the Hidaka Metamorphic Belt, and complex field relationships indicate mixing and mingling of mafic and felsic magmas (Owada *et al.* 2003), as well as mixing between I- and S-type tonalites.

Experiments show that the tonalitic magma began to crystallize above 900°C (930°C for orthopyroxene crystallization, higher, but undetermined for garnet crystallization) for 3–4% H_2O at 600 MPa. Model calculations suggest that restite for the S-type tonalitic magmas would be garnet–orthopyroxene aluminous granulite (Shimura *et al.* 2004).

Because it is intrusive, the garnet–orthopyroxene tonalitic magma must have been produced at even greater depth, possibly a few km deeper (Owada *et al.* 2003, p. 93). Moreover, the estimated temperature of the highest-grade Hidaka gneisses (870°C) is less than the experimentally determined temperature of initial crystallization of garnet and orthopyroxene in the garnet–orthopyroxene tonalitic magma (Osanai *et al.* 1992). Therefore, the melting conditions have been inferred to be >900°C at about 800 MPa (Osanai *et al.* 1997, Owada *et al.* 2003, Shimura *et al.* 2004). The unexposed lowermost part of the crust for the Hidaka Metamorphic Belt is probably garnet – orthopyroxene – clinopyroxene mafic granulite, garnet – orthopyroxene aluminous granulite, and gabbro (Shimura *et al.* 2004).

The inferred pressure and tonalite liquidus are consistent with the anticipated situation in the upper parts of a MASH zone in magmatically heated crust, as are the anhydrous granulite-facies assemblages and the abundance of intrusive mafic and ultramafic rocks, which presumably represent at least the local heat-source. Experimental studies by McCarthy & Patiño-Douce (1997) have shown how interaction of hot mafic (aluminous olivine tholeiitic) magma with pelitic material can produce felsic (dacitic to rhyolitic) magmas and garnet granulite as a residuum.

Chappell *et al.* (1987) inferred liquidus temperatures of 1,000°C for the Hawkins Volcanics, on the basis of the experimental results of Clemens & Wall (1981) for a slightly more felsic, less magnesian composition. They stated that such high temperatures are at odds with metamorphic geothermal gradients. However, though these temperatures are not attained in normal amphibolite–granulite-facies terranes, they may well be reached in deep crust heated by basaltic magma. Because temperatures of well over 900°C are indicated by the pyroxene–garnet tonalites of the lowest exposed part of the Hidaka Metamorphic Belt, the temperatures were probably higher for the deeper, unexposed rocks (intensely heated by mafic magma) from which the pyroxene–garnet-bearing tonalitic magmas were generated.

Some xenoliths in basalt confirm the extensive existence of high-temperature granulite-facies paragneisses (0.9–1.4 GPa) in the deep crust of central Mexico (Hayob *et al.* 1989). The rocks were heated to about 1100°C by mantle-derived, underplated mafic magma, cooled to 900°C (as indicated by mesoperthite) and then collected by the basaltic magma. Partial melting is indicated by rims of inferred former melt around garnet, showing evidence of devitrification in the form of intergrowths of euhedral orthopyroxene, hercynitic spinel and either interstitial glass or calcic plagioclase. The rims and fine-grained intergrowths suggest rapid melting and quenching, presumably during ascent.

Source of Enclaves in S-Type Granitic Magmas

Enclaves of metapsammitic–metapelitic gneiss (garnet – orthopyroxene – biotite granulite) in the garnet–orthopyroxene tonalite of the Hidaka Metamorphic Belt would seem to be the most likely potential candidates for restite in high-level S-type granites. However, these enclaves show evidence for fluid-absent partial melting reactions (Shimura *et al.* 1999). Therefore, they are not restite for the host tonalite, but "accidental" restite formed by heating of country-rock xenoliths by the host tonalitic magma. Small clots consisting mainly of orthopyroxene, garnet and plagioclase dispersed through the tonalite may be primary restite (Shimura *et al.* 1999). However, only some of the tonalite contains these clots, so that much of the magma was able to free itself of restite, even at that depth.

As pointed out by Clemens (2003), enclaves with orthopyroxene and garnet typically are absent from S-type granites of the LFB, suggesting that restite is either removed close to the source (which is likely for hot, mobile magma in a strongly deformed setting) or reacts completely with the magma during ascent. Effective removal of restite close to the source is suggested by detailed studies of diatexites (Sawyer 1996, 1998, 2001). Moreover, Clemens *et al.* (1997) have shown that restite may be dissolved during magma transport, owing to superheating of H_2O-undersaturated granitic magma during rapid, adiabatic rise in the crust. Minerals that become unstable and tend to be resorbed during ascent of S-type granitic magmas are garnet and quartz, possibly with some biotite, as well as cordierite below 400 MPa; orthopyroxene would remain stable (Clemens *et al.* 1997). I suggest that examples of this kind of arrested melting could be felsic glass in metasedimentary xenoliths and xenocrysts in the Hoyazo dacite (discussed previously) and devitrified glass rims on garnet in granulite-facies metasedimentary xenoliths in basalt in central Mexico (Hayob *et al.* 1989).

In the Hidaka Metamorphic Belt, tonalites intrusive into amphibolite-facies rocks above the garnet–orthopyroxene tonalite are cordierite-bearing, and contain less garnet and orthopyroxene. Shimura *et al*. (1992) inferred that these higher-level tonalites are differentiates of the garnet–orthopyroxene-bearing tonalitic magma. They contain xenoliths of amphibolite-facies country rocks, such as biotite–muscovite gneiss and schist, garnet–biotite gneiss, and amphibolite (Osanai *et al*. 1991, Shimura *et al*. 1999). This kind of source would be suitable for amphibolite-facies metapelitic xenoliths in the LFB S-type granites, whereas a granulite-facies source similar to the deeper (unexposed) parts of the Hidaka Metamorphic Belt would be appropriate for the LFB S-type granitic magmas.

Field relationships consistent with local mixing involving mafic rock and tonalitic magma in the Hidaka Metamorphic Belt suggest that magmas suitable for producing the microtonalite enclaves in the Cowra Granodiorite may have been formed close to the source of the felsic magma. The mixing may have occurred in a MASH zone similar to the one that I speculate may have existed beneath the Hidaka Metamorphic Belt. This interpretation would be consistent with the composition of orthopyroxene–cordierite-bearing tonalite and evidence of magma mixing in the enclaves.

Conclusions

Detailed evidence indicates that microgranitoid enclaves in the LFB high-level S-type granites are quenched globules of magma, not restite. The evidence is less clear for metasedimentary xenoliths, but nevertheless suggests that they are neither restite nor resistate, but are probably xenoliths accidentally incorporated into ascending magma, possibly from higher-level parts of the source terrane. The evidence concerning single grains and small aggregates is least convincing, possibly apart from some zircon; quartz cores could be hard to detect, and calcic plagioclase cores can be formed in several ways, as can cordierite and garnet grains with sillimanite inclusions. Phenocrysts with felsic glass inclusions in S-type volcanic rocks, which have been interpreted as restite by some investigators, are more likely to be magmatic, especially because quartz and at least some feldspar phenocrysts with glass inclusions cannot be products of a peritectic reaction, implying that this interpretation is probably also true for the other phenocrysts with glass inclusions (cordierite, orthopyroxene and garnet).

A granulite-facies source similar to the deeper (unexposed) parts of the Hidaka Metamorphic Belt would be appropriate for the S-type granites of the LFB. An amphibolite-facies source similar to the higher parts of the Hidaka Metamorphic Belt would be suitable for metapelitic xenoliths in the LFB S-type granites. Microtonalite enclaves in the LFB S-type granites are former globules of magma generated by the mixing of peraluminous felsic magma with more mafic magma, possibly in or close to the melting zone, before mingling in the host magma.

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