Middle Carboniferous-Early Triassic eclogite-blueschist blocks within a serpentinite mélangé at Port Macquarie, eastern Australia: implications for the evolution of Gondwana's eastern margin

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Keywords
blueschist, blocks, within, serpentinite, m, large, port, macquarie, eastern, eclogite, australia, middle, implications, evolution, gondwana, 039, margin, triassic, early, carboniferous, GeoQuest

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Middle Carboniferous - Early Triassic eclogite-blueschist blocks within a serpentinite mélange at Port Macquarie, eastern Australia: Implications for the evolution of Gondwana's eastern margin

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Abstract

The New England Orogen of easternmost Australia is dominated by suites of Palaeozoic to earliest Mesozoic rocks that formed in supra-subduction zone settings at Gondwana’s eastern margin. On the northern New South Wales coast at Rocky Beach, Port Macquarie, a serpentinite mélange carries rare tectonic blocks of low-grade, high-pressure, metamorphic rocks derived from sedimentary and igneous protoliths. Dominant assemblages are glaucophane + phengite ± garnet ± lawsonite ± calcite ± albite blueschists and lawsonite-bearing retrogressed garnet + omphacite eclogites. In some blocks with sedimentary protoliths, eclogite forms folded layers within the blueschists, which is interpreted as Mn/(Mn+Fe) compositional control on the development of blueschist versus eclogite assemblages. Review of previous studies indicates pressure-temperature conditions of 0.7-0.5 GPa and ≤450°C. Three samples of high-pressure metasedimentary rocks contain Archaean to 251 ± 6 Ma (Permo-Triassic) zircons, with the majority of the grains being Middle Devonian to Middle Carboniferous in age (380-340 Ma). Regardless of age, all grains show pitting and variable rounding of their exteriors. This morphology is attributed to abrasion in sedimentary systems, suggesting that they are all detrital grains. New \textit{in situ} metamorphic zircon growth did not develop because of the low temperature (≤450°C) of metamorphism. The Permo-Triassic, Devonian and Carboniferous zircons show strong heavy rare earth element enrichment and negative europium anomalies, indicating that they grew in low pressure igneous systems, not in a garnet-rich plagioclase-absent high pressure metamorphic environment. Therefore the youngest of these detrital zircons provide the \textit{maximum} age of the metamorphism. A titanite + rutile porphyroblast within an eclogite has a U-Pb age of 332±140 Ma (poor precision due to very low U abundances of mostly <1 ppm.) and provides an imprecise direct age for metamorphism. In the south of the Port Macquarie area, the Lorne Basin ≥220 Ma Triassic sedimentary and volcanic rocks unconformably overlie serpentinite mélange, and provide the \textit{minimum} age of the high-pressure metamorphism. Our preferred interpretation is that the 251
Ma zircons are detrital and thus the Port Macquarie high-pressure metamorphism is constrained to the end of the Permian – Early Triassic. Emplacement of the serpentinite mélange carrying the Rocky Beach high-pressure rocks might have been due to docking of a Permian oceanic island arc (represented by the Gympie terrane in southern Queensland?) and an Andean-style arc at the eastern Australian margin (expressed in the New England Orogen by 260-230 Ma north-south orientated magmatic belts). Alternatively, if the 251 Ma grains are regarded as having grown in thin pegmatites, then the dominant Devonian-Carboniferous detrital population still indicate a maximum age for the high pressure metamorphism of ca. 340 Ma. A ≤340 Ma age of metamorphism would still be much younger than the previously suggested ca. 470 Ma (Ordovician) age, which was based on Ar-Ar dating of phengites.

Keywords: New England Orogen; Blueschist-eclogite; Serpentinites; Gondwana; Port Macquarie
1. Introduction

Throughout the Palaeozoic, eastern Australia was at the margin of Gondwana, and was growing eastwards via the addition of granitoids and their volcanic and sedimentary carapaces, largely formed in supra-subduction zone settings (e.g., Aitchison et al., 1994, Cawood, 2005; Glen, 2005; Cawood et al., 2011; Rosenbaum et al., 2012). Lineaments carrying slivers of tectonised, serpentinised, mantle are found intertwined between the granitoids and their carapaces (e.g., Spaggiari et al., 2004; Glen; 2005). The serpentinites carry sedimentary and igneous rocks as mélanges of exotic blocks, including rare ones carrying eclogite and blueschist facies assemblages. They are interpreted as vestiges of subduction zone assemblages modified by younger tectonic events (e.g., Glen, 2005).

In the New England Orogen (NEO) of northeastern New South Wales and south-eastern Queensland (Fig. 1), two notable occurrences of serpentinite mélange are the Great Serpentine Belt in the Peel-Manning fault system in the southwest (Benson, 1913, 1914a, b, 1915a,b; Offler and Williams, 1987; Sano et al., 2004; Phillips and Offler, 2011) and Port Macquarie in the southeast (Benson, 1918; Barron et al., 1976; Leitch, 1980; Och et al., 2003, 2007; Lennox and Offler, 2009). Exotic blocks within the Great Serpentine Belt have so far all yielded latest Neoproterozoic to early Palaeozoic U-Pb zircon protolith ages (Aitchison and Ireland, 1995; Watanabe et al., 1997). The NEO serpentinites have been interpreted as exhumed vestiges of early Palaeozoic subduction zone complexes, caught-up in latest Palaeozoic to earliest Mesozoic crust-forming events (e.g., Och et al., 2003). However, the protolith ages of the included blocks in the serpentinites can give only maximum, rather than direct, ages for the high-pressure metamorphic events and the incorporation of the blocks within the serpentinite belts. Age determinations on minerals grown or recrystallised during high-pressure metamorphism will give the sought-after direct
metamorphic ages. Particularly at medium to high metamorphic grade, this can be achieved by employing U-Pb dating of metamorphic zircon growths or Sm-Nd determinations on coexisting garnet and omphacite (e.g., Hermann et al., 2001; Gilotti et al., 2004; Nutman et al., 2008). At low grade, direct determination of metamorphic ages of high pressure assemblages becomes difficult, because under these low temperature conditions the growth of metamorphic zircon is uncommon and because of the issue of excess Ar in metamorphic phengites derived from older detrital grains resulting in incorrect, older ages in K-Ar and Ar-Ar geochronology (e.g., Arnauld and Kelley, 1995; Li et al., 1999; Sherlock and Kelley, 2002).

In this paper we present field relationships, zircon U-Pb geochronology and rare earth element (REE) chemistry and metamorphic rutile and titanite U-Pb dating from metasedimentary blueschist and eclogite facies blocks in the NEO’s easternmost serpentinite mélange at Port Macquarie (Fig. 1). Our results show that these rocks formed at the end of the Palaeozoic and that high-pressure metamorphism was most likely at ≤251 Ma, i.e., ca. 200 million years later than the previously proposed Ordovician timing for this event at Port Macquarie (Fukui et al., 1995; Och et al., 2003, 2007). We discuss the implications of this Permo-Triassic event for the evolution of Gondwana’s eastern margin.

2. Regional geological setting – the New England Orogen

The New England Orogen (NEO; Fig. 1) is the easternmost, youngest, component of the Tasmanides orogenic belt of eastern Australia (Flood and Aitchison, 1988; Harrington and Korsch, 1985; Leitch, 1974, 1975; Murray et al., 1987; Aitchison et al., 1994; Cawood, 2005; Glen, 2005; Cawood et al., 2011; Rosenbaum, 2012). The NEO contains latest Neoproterozoic to Permo-Triassic arc-related
magmatic rocks (Leitch, 1974; Murray et al., 1987; Aitchison et al., 1994; Cawood et al., 2011; Rosenbaum et al., 2012). Devonian-Carboniferous supra-subduction zone components are the Tamworth Belt forearc basin and Tablelands Complex accretionary wedge rocks (Fig. 1) which developed on top of previously-accreted early Palaeozoic oceanic terranes (Flood and Aitchison, 1988). The core of a Devonian-Carboniferous volcanic arc lies farther to the west, but it is now largely obscured by post-Devonian sedimentary rocks (Leitch, 1974, 1975; McPhie and Fergusson, 1987; Roberts and Engel, 1987; Roberts and James, 2010; Roberts et al., 1995, 2004, 2006). It is only evident from intermediate to felsic volcanic rocks of the Currabubula Arc on the western edge of the Tamworth Belt. Southwards, tectonically-displaced forearc basin rocks in the Hastings Block (Figure 1 inset) are correlated with the Tamworth Belt. The Tablelands Complex rocks are mostly Carboniferous to early Permian deep marine volcaniclastic turbidites and cherts, mafic volcanic rocks and olistostromal deposits (Leitch & Cawood, 1980; Cawood, 1982; Fergusson, 1984; Aitchison et al., 1994). These rocks show penetrative steeply dipping fabrics and display prehnite-pumpellyite/lower greenschist to amphibolite-facies metamorphic assemblages (Binns et al., 1966; Korsch and Harrington, 1981; Dirks et al., 1992). The Peel-Manning fault system (Fig. 1a) forms the tectonic boundary between the Tamworth Belt and the Tablelands Complex. It is marked by serpentinite belts that contain exotic tectonic blocks of early Phanerozoic to Neoproterozoic arc-related rocks and rare blocks of eclogite and blueschist facies rocks (e.g., Shaw and Flood, 1974; Fukui et al., 1995; Aitchison and Ireland, 1995). Generally, an early Palaeozoic age has been ascribed for the Peel-Manning fault system assemblage and the age of high-pressure metamorphic blocks within it (e.g., Fukui et al., 1995; Phillips and Offler, 2011), albeit it is recognised that the Peel-Manning serpentinites were
deformed and metamorphosed at the end of the Permian (Aitchison and Flood, 1992; Aitchison et al., 1997).

The Early Permian history of the NEO is marked by development of sedimentary basins dominated by proximal seep marine mass flow deposits (Aitchison and Flood, 1992; Aitchison et al., 1997; Korsch et al., 2009). The Early Permian granitoids are the S-type Bundarra granite and the Hillgrove suite, and several other small S- and I-type plutons; Figure 1a). U-Pb zircon ages (Cawood et al., 2011; Rosenbaum et al., 2012) show that these were all emplaced between 296 and 267 Ma. Thereafter, from 250 to 225 Ma, subduction resumed, and there was the emplacement of eastward-stepping north-south trending arc-related plutons. The 250–225 Ma magmatism produced I-type granitoids and calc-alkaline volcanic rocks, which are interpreted to have been emplaced in an Andean arc setting (Cawood, 1984; Flood & Shaw, 1975; Bryant et al., 1997).

One proposed geodynamic setting for 296-230 Ma crustal evolution is an eastwards, asymmetric, subduction rollback, that between ca. 300 and 267 Ma lead to the evolution of an orocline in the southern end of the NEO (Jenkins et al., 2002; Collins and Richards, 2008; Cawood et al., 2011; Rosenbaum et al., 2012). After 267 Ma the proposed orocline in the southern NEO continued to tighten (Murray et al., 1987; Cawood et al., 2011; Rosenbaum et al., 2012) and hence there was a hiatus in arc magmatism between 267-250 Ma. Subduction with arc magmatism resumed from 250 Ma. Thereafter, there was the onset of the Hunter-Bowen orogeny – a phase of E-W contractional deformation (Collins, 1991; Holcombe et al., 1997; Korsch et al., 2009). This orogeny involved westward-propagating thrusting of the early Permian basins, with their transformation into foreland systems (Fergusson, 1991; Korsch et al., 2009). The east-west compression of the Hunter-Bowen orogeny has been attributed to
coupling in the subducted and over-riding plates, following further eastward migration of subduction systems located at the edge of Gondwana (Rosenbaum et al., 2012).

However, another alternative is the docking of a Permian intra-oceanic island arc terrane against the eastern edge of Gondwana (Harrington, 1983). Following the Hunter-Bowen orogeny, Triassic to Jurassic depositional basins developed over the NEO.

3. The Port Macquarie serpentinite mélangé

3.1. Geological background

The Port Macquarie area is mostly underlain by a ‘broken formation’ of Palaeozoic units intercalated with the Port Macquarie serpentinite mélangé (Leitch, 1974; Barron et al., 1976; Och et al., 2007). Common lithologies are metabasalts, with preservation of pillow structures quite widespread, cherts bearing radiolarians and conodonts, and clastic sedimentary rocks that are mostly volcanic lithic arenites (e.g., Barron et al., 1976; Leitch, 1980). The easternmost of these along the coast are well exposed, and have been named the Watonga Formation (Leitch, 1980). Based on radiolarian and conodont assemblages, Watonga Formation metacherts have been ascribed Ordovician and Devonian ages (Ishiga et al., 1988; Och et al., 2007a). The Tacking Point Gabbro (31°28′24″S 152°56′14″E; Och et al., 2007b) was intruded into Watonga Formation metabasalts and cherts, and based on U-Pb zircon dating, is Devonian in age, (Buckman et al., in preparation). Zircon U-Pb dating on detrital/volcanic zircons from a Watonga Formation volcanic lithic arenite from Green Point (31°25′49″S 152°55′26″E; Parker, 2010) indicates a maximum mid-Carboniferous depositional age (Buckman et al., in preparation). Thus the Watonga Formation contains rocks of different age and origin. Similar
metasedimentary units inland are poorly exposed. They are either intruded or
intercalated with metadolerite, and were assigned to the ‘middle’ Palaeozoic
Touchwood Formation and the Lower Permian Thrumster slate by Leitch (1980).

Cropping out sporadically along the shoreline from Town Beach (31°25’47”S
152°55’20”E) to just north of Tacking point (around 31°28’13”S 152°56’05”E) is the
Port Macquarie serpentinite mélange (Barron et al., 1976; Och et al., 2007b). Recent
mapping suggests that the mélange structurally overlies the parts of the Watonga
Formation that consist of lithic arenites with Carboniferous detrital zircons (Buckman
et al., in preparation). However, inland, because of poor exposure, the relationship of
the mélange to the Touchwood Formation and the Thrumster slate has not been
elucidated. These Ordovician to Lower Permian tectonic assemblages, including the
serpentinites, are overlain unconformably by Early Triassic alluvial deposits and
volcanic rocks of the Lorne Basin (Leitch, 1980; Och et al., 2007b; Pratt, 2010). Thus
the last major tectonic event involved the intercalation of Devonian to Lower Permian
rocks, and must have occurred after deposition of the Early Permian Thrumster slate
and before the deposition of the unconformable Lorne Basin Early Triassic rocks
(timescale of Gradstein et al., 2012). Some deformation continued after deposition of
the Lorne basin, as shown by folding in its eastern parts (Pratt, 2010).

3.3. Rocky Beach blueschist and eclogite blocks

At the northern end of Rocky Beach (31°26’12”S 152°55’33”E) several blocks
of high-pressure metamorphic rocks up to 10 m diameter occur within the Port
Macquarie serpentinite mélange (Barron et al., 1976; Och et al., 2003, 2007b). One ca.
10 m block of massive blueschist preserves relict pillow structures in its centre, which
towards its margin becomes obliterated by higher strain with the development of a
glaucophane schistosity. Such rocks are clearly derived from basaltic protoliths. Given
the protolith nature and the low temperature of metamorphism, such rocks are unlikely
to carry zircon and were not investigated in this study. Other blocks consist of
blueschist with thin, folded layers of eclogite (Fig. 2a; also illustrated by Och et al.
2003 as their Fig. 2). This block was the source of our samples eclogite 21-06 and
blueschist RB-2. There are also blocks of widely retrogressed, high-pressure
metamorphic rocks that also have compositional layering (Fig. 2b; our sample RB-3 is
from this block). These retrogressed rocks still contain some relict garnet, but the
omphacite and/or glaucophane have been entirely replaced by chlorite. Consequently,
it is uncertain whether they originally carried blueschist or eclogite facies assemblages.
The blocks with compositional layering are likely to have been derived from
sedimentary or volcano-sedimentary protoliths. Thus we targeted them for zircon U-Pb
dating, because even if it transpired that the metamorphic grade was too low to grow
zircon in situ during high-pressure metamorphism, their youngest detrital zircons
would define a maximum possible age for that metamorphism.
The whole rock geochemical analyses for eclogite 21-06, blueschist RB-2 and
retrogressed high-pressure rock RB-3 are presented in Table 1. Their high MgO +
Fe₂O₃, Cr (400-206 ppm) and Ni (68-64 ppm) contents show dominance of mafic
material. However, high K₂O (3.63-1.02 wt%), Ba (1102-401 ppm) and overall low
SiO₂ contents (down to 43.88 wt%) suggest that the protoliths were not fresh mafic
rocks. Furthermore Zr content is high (275-103 ppm). Retrogressed sample RB-3 has
an anomalously high TiO₂ content (3.37 wt%). The block from where RB-3 was
sampled contains ilmenite + rutile segregations (sample RB-4, see below).
Over-representation of such segregation material in the bulk sample is a likely reason
for the high TiO₂ content. Combining the interlayered nature of eclogite and blueschist
(21-06 and RB-2), with the whole rock chemistry, we propose that the protoliths of the
dated samples are dominated by altered mafic volcanic material, with an additional
detrital component.

In the Rocky Beach blueschist facies rocks, the main assemblage is
glaucophane + garnet + phengite ± lawsonite ± albite + titanite + pyrite. In the eclogites
lawsonite is present, but there has been extensive retrogression of the garnet and
omphacite, with the development of chlorite. Both Barron et al. (1976) and Och et al.
(2003) recorded complex metamorphic textures, with several generations of
high-pressure mineral growth, during the development syn-kinematic fabrics. The
blueschists are traversed by thin albite-bearing veins. For the Rocky Beach
lawsonite-bearing blueschist rocks Price (1991) estimated pressure-temperature
conditions of ca. 0.75-0.6 GPa and 270-200°C, whereas Och et al. (2003) suggested
somewhat higher temperatures of 450-350°C, in a similar pressure regime. Och et al.
(2003) suggested appreciably high temperatures and pressures for the Rocky Beach
eclogites (see Discussion).

4. Zircon petrography, U-Pb dating and rare earth element chemistry

4.1. Analytical methods

Zircon concentrates were prepared by standard heavy liquid and isodynamic
separation techniques at the mineral separation laboratory of the Research School of
Earth Sciences, the Australian National University (ANU). The location of the three
samples investigated are shown in Figs. 2a,b). The concentrates were hand picked
under a binocular microscope, and the selected grains were cast in epoxy resin discs
together with the zircon Temora 1 reference material (Black et al., 2003). The discs
were ground to reveal mid-sections through the grains and then polished. The grains
were documented with reflected and transmitted light photomicrographs (Fig. 3), and
also by cathodoluminescence (CL) imaging. CL images of representative zircons are shown in Fig. 4.

Zircons were U-Pb dated on the SHRIMP II instruments at the University of Hiroshima and Geoscience Australia (Canberra). Analytical protocols followed Williams (1998), and reduction of the raw data used the ANU software ‘PRAWN’ and ‘Llead’. \(^{206}\text{Pb}/^{238}\text{U}\) of the unknowns was calibrated using measurements of the Temora 2 reference material (U-Pb ages concordant at 417 Ma; Black et al., 2003) interleaved with analyses of the unknowns. U and Th abundance was calibrated using measurement of the reference zircon SL13 (U=238 p.p.m.) located in a set-up mount. The zircon U-Th-Pb data are summarised in Table 2. The reduced and calibrated data were assessed and plotted using the ISOPLOT program of Ludwig (2003).

Rare earth element (REE) analyses of some zircons were undertaken using the Hiroshima University SHRIMP II instrument, and for others the REE and other trace element abundances were acquired by LA-ICP-MS at ANU. The SHRIMP analyses were undertaken in the normal mass-resolution mode (~6000), and according to the method outlined by Maas et al. (1992) and Hidaka et al. (2002). Abundances were calibrated using count rates obtained from analyses of the zircon standard 91500 (Wiedenbeck et al., 2004). The zircon LA-ICP-MS data was acquired according to the methodology given by Eggins et al. (1998). It was calibrated using interspersed analyses of NIST610 glass. As a check on the accuracy of elemental abundance, SL13 zircons were analysed twice as unknowns, and yielded an average U abundance of 230 p.p.m., compared to the accepted isotope dilution value of 238 ppm. REE analyses are summarised in Table 3.

3.2. Zircon morphology and U-Pb dating
Three blueschist-eclogite samples from Rocky Beach (samples 21-06, RB-2 and RB-3; Fig. 2a,b) all gave low yields of zircons (<50 grains per kilogram), and display the same range of morphologies. The exteriors of all grains are rounded, and regardless of their age show pitting of their surface, which is attributed to abrasion within sedimentary systems (Fig. 3a-c). Their rounded and pitted form contrasts strongly with igneous zircons in Port Macquarie rocks, such as from the Tacking Point gabbro (31°28’24”S 152°56’14”E; Och et al., 2007b), at the southern end of the Port Macquarie coastal tract. In such rocks, the zircons are not pitted and sharp clear boundaries are preserved between the grain facets (Fig. 3d). Furthermore, the exterior form of the zircons in Rocky Beach samples differs from definite syn-
high-temperature eclogite facies metamorphic zircon, for example from the Franklin metamorphics of Tasmania (Fig. 3e). In such eclogites, metamorphic zircons have clean, exterior surfaces, with common development of multifaceted habits. Therefore, combined with evidence from CL images for the lack of overgrowths developed in situ, we conclude that the zircons from both Rocky Beach blueschist and eclogite facies metamorphic rocks are best interpreted as detrital grains, largely devoid of coatings of new, syn-metamorphic growth.

CL images of selected zircons are shown in Fig. 4. A few rare grains (2 in sample 21-06, 1 in sample RB-3; Figs. 4a,b) are structurally simple, slightly rounded euhedral prisms with well-preserved oscillatory zoning. Multiple analyses on these three grains yielded a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 251 ± 6 Ma (MSWD=0.16; Fig. 5a; Induan = earliest Triassic, according to the timescale of Gradstein et al., 2012). These grains are interpreted to be detrital in origin and derived from an Induan igneous source. Their age falls at the start of the youngest arc magmatism in the New England Orogen. Seven 380-340 Ma grains (Middle Devonian to Middle Carboniferous) are somewhat more rounded, devoid of inherited cores and with well developed oscillatory
zoning (Figs. 4c and 5a). Older grains form more complex populations, with many
grains being composite, with magmatic cores and younger magmatic overgrowths, or
more commonly rounded cores of oscillatory-zoned zircon with overgrowths that
appear homogeneous or sector-zoned in CL images. In all such cases our interpretation
is that the metamorphic overgrowths have been abraded and partly removed during
sedimentary transport, showing that they carry a record of distal metamorphism in their
source region, and not \textit{in situ} metamorphism in the Rocky Beach blueschist-eclogite
facies blocks. This is illustrated by RB-2 grains 2 and 4 and RB-3 grain 3 (Figs. 4d,e,f).
In Neoproterozoic RB-2 grain 2, a 735 Ma core has a 594 Ma overgrowth, partly
removed by abrasion such that the core occurs at half of the grain’s surface. For RB-3
grain 3, a rounded oscillatory core (not dated) has a sector-zoned ca. 405 Ma (early
Devonian) rim, which is partly abraded. At the right hand grain margin (Fig. 4f), the
inclusion of another mineral previously entirely enclosed, is now at the grain surface.
Thus the eclogite and blueschist samples all yielded essentially the same zircon
populations, both in terms of habit, structure and age distribution.

4.3. Zircon REE chemistry

REE analyses of zircons from the blueschist and eclogite facies rocks were
undertaken to ascertain which phases they equilibrated with (chondrite-normalised
data are plotted in Fig. 5b). 251 Ma oscillatory zoned prism grains 4 and 7 from sample
21-06 display strong enrichment of the heavy over the light REE, with a negative Eu
anomaly. This indicates the youngest grains grew in the presence of plagioclase and
absence of garnet, showing that they did not grow during high-pressure
metamorphism. A similar REE pattern is found for sample 21-06 Carboniferous grains
10, 11 and 14 (Fig. 5b). Other REE analyses focussed on metamorphic (albeit abraded)
overgrowths, to ascertain if they provide evidence of blueschist-eclogite facies metamorphism at an earlier time, i.e. in accord with suggestion that it occurred in the early Palaeozoic (Och et al., 2003, 2007b, 2010 and references therein). Most of these rims, ranging in age from ca. 405 Ma on RB-3 grain 3 to ca. 595 Ma Neoproterozoic ones such as on RB-2 grains 2 and 4 (Figs. 4d and e) show similar heavy REE enriched patterns with negative Eu anomalies, demonstrating that during metamorphism they grew in a low pressure plagioclase-present, garnet-absent environment (Rubatto, 2002). The only exceptions are the ca. 560 Ma rim on RB-2 grain 6 and the entire 1100 Ma 21-06 grain 15, which both show depressed heavy REE abundances but still carry negative Eu anomalies (Fig. 5b). These indicate source region metamorphisms at medium pressures with both garnet and plagioclase present (Rubatto, 2002). Thus none of the dated zircons from the blueschist-eclogite blocks carry a depressed heavy REE without a Eu anomaly signature, which would be indicative of high-pressure metamorphism (example from an eclogite facies metamorphic zircon from the Franklin metamorphics is shown in Fig. 5b for comparison).

5. Titanite-rutile U-Pb dating

5.1. Sampling and analytical method

Rocky beach eclogites locally contain porphyroblasts of titanium rich phases which were identified by use of a Niton hand-held portable XRF analyser. One >1 cm diameter porphyroblast (sample RB-4; Fig. 2b,c) was sawn from a retrogressed high pressure metamorphic block, cast in an epoxy resin disc along with the ANU Khan titanite standard and polished for phase identification via EDS spectral analysis with a scanning electron microscope. The phases were then dated by the LA-ICP-MS method.
LA-ICP-MS analysis was undertaken at GEMOC at Macquarie University, Sydney using a New Wave UP213 Nd:YAG 213 nm laser ablation system coupled to an Agilent 7700 quadrupole ICP-MS. The setup for LA-ICP-MS analysis and analytical procedure were described by Jackson et al. (2004) and R19 rutile (489.5 ± 0.9 Ma; Zack et al., 2011) was used as a calibration standard.

5.2. Titanite-rutile segregation petrography and U-Pb dating

A backscattered electron image of part of the segregation is shown in Fig. 6. The main phases present are titanite and rutile, seemingly in equilibrium with each other. Sulphides, chlorite, glaucophane and ilmenite also occur.

The U-Pb data are summarised in Table 4. The high \( \frac{^{207}\text{Pb}}{^{206}\text{Pb}} \) combined with low \( \frac{^{238}\text{U}}{^{206}\text{Pb}} \) in all the analyses shows that Pb in the rutiles and titanites are dominated by ‘common’ non-radiogenic Pb. In order to obtain a radiogenic age avoiding the problem of isobaric mercury (\( ^{204}\text{Hg} \)) interference at the \( ^{204}\text{Pb} \) mass when undertaking a correction for common Pb, the data was regressed in \( \frac{^{207}\text{Pb}}{^{206}\text{Pb}} \) versus \( \frac{^{238}\text{U}}{^{206}\text{Pb}} \) space, as shown in Fig. 7. In this approach, where the regression line intercepts the \( \frac{^{207}\text{Pb}}{^{206}\text{Pb}} \) axis (\( \frac{^{238}\text{U}}{^{206}\text{Pb}}=0 \)), it gives the \( \frac{^{207}\text{Pb}}{^{206}\text{Pb}} \) composition of the common Pb. Where the regression line gives a lower intercept with concordia, it signifies the U-Pb radiogenic component in the rutiles and titanites, which in this case is interpreted as the timing of metamorphic growth. All the data fall along a single regression line, with an MSWD=0.71. Given that most of the coexisting rutiles and titanites have <1 ppm, each determination has a large analytical error and is dominated by the non-radiogenic ‘common’ Pb. Hence, there is a long extrapolation of a regression line with a broad error envelope to a lower concordia intercept. The ‘closure’ temperature of both titanite and rutile are ca. 600°C (Scott and St-Onge,
1995; Cherniak, 2000), which is above the peak metamorphic temperature of these rocks. Therefore the lower concordia intercept of $332 \pm 140$ Ma (95% confidence) is interpreted as giving the age of metamorphic rutile + titanite growth, but with poor precision.

6.1. Garnet rare earth element chemistry

6.1. Sampling and analytical method

Fragments of garnet grains were cast in an epoxy resin disc and polished to reveal cross sections for analysis. They were analysed by the LA-ICP-MS method at GEMOC at Macquarie University, Sydney using a New Wave UP213 Nd:YAG 213 nm laser ablation system coupled to an Agilent 7700 quadrupole ICP-MS. The setup for LA-ICP-MS analysis and analytical procedure were described by Jackson et al. (2004) and NIST610 glass was used as a calibration standard. REE analyses of 21-06 and RB-2 garnets are summarised in Table 3.

6.2. Results

Overall the garnet REE abundances are low, but relative to chondrite, the heavy REE are enriched and the light REE are depleted (Fig. 5b). There are no Eu anomalies, indicating that the garnet grew in a plagioclase-free environment, during high pressure metamorphism.

7. Discussion

7.1. Interpretation of the garnet and zircon rare earth element signatures and the zircon ages
With two exceptions (see below) all zircons in samples 21-06, RB-2 and RB-3 are strongly enriched in the heavy REE and depleted in the light REE relative to chondrite, and show negative Eu anomalies (Fig. 5b). These patterns and abundances are coincident with magmatic zircons grown from common granitic melts, in the presence of plagioclase (Belousova et al., 2002). The only zircons that equilibrated with garnet (based on lesser enrichment of the heavy REE) are 21-06 grain 15 (a 1100 Ma low Th/U oval grain), and RB-2 analysis 6.1 (a 560 Ma abraded overgrowth; Fig. 5b). However, both these show negative Eu anomalies, indicating equilibration with plagioclase. Therefore they did not grow *in situ* in a plagioclase-free high-pressure metamorphic environment, but instead are detrital grains reflecting lower pressure metamorphism in the sediment source regions. In contrast, the 21-06 and RB-2 garnets do not show Eu anomalies and thus did not equilibrate with plagioclase. Consequently, none of zircons grew in equilibrium with the high pressure metamorphic garnet. Instead, they are interpreted as detrital grains, thereby they only give a maximum age for the high-pressure metamorphism.

### 7.2. Timing of high-pressure metamorphism

At Green Point (31°25’49”S 152°55’26”E) the Port Macquarie serpentinite mélange carrying the high-pressure rocks is in tectonic contact with low metamorphic grade Carboniferous volcanic lithic arenites previously ascribed to the upper portion of the Watonga Formation (Buckman et al., in preparation). In the south at Tacking Point the serpentinites are in tectonic contact with an Ordovician-Devonian metachert and Devonian igneous complex (Buckman et al., in preparation). On the basis that these tectonic contacts are not entirely late faults much younger than upper crustal emplacement of the serpentinite with its high-pressure inclusions, these field
relationships would argue that the Port Macquarie serpentinite mélange is Late Carboniferous or younger. In the southern part of the Port Macquarie area, Early Triassic, >220 Ma alluvial sedimentary rocks of the Lorne Basin unconformably overlie serpentinites, giving the minimum age for the high pressure metamorphism (Leitch, 1980; Och et al., 2007b; Pratt, 2010).

Zircons from sample 21-06 were concentrated in 2010, whereas zircon concentrates for the other two samples RB-2 and RB-3 were prepared in 2011. All samples gave a small yield of zircons (<50 grains per kilogram), with similar morphology and age distribution, including young, ca. 250 Ma zircons. Given that the zircon concentrates were prepared at two different times, and the zircons hand-picked and cast into epoxy mounts at different times, we consider it unlikely that the ca. 250 Ma zircons are laboratory contaminants, but are the youngest detrital grains from these samples. Even if these ca. 250 Ma grains are discounted (such as being sourced from the thin albite pegmatite veins), it should be noted that the majority of the grains are Middle Carboniferous or Devonian in age, and are thus still appreciably younger than the Ordovician age for the high-pressure metamorphism proposed by Och et al. (2003, 2007b, 2010). Thus from geological relationships and from our zircon studies we conclude that the high-pressure metamorphism occurred before deposition of the Early Triassic Lorne basin rocks and probably after 251 Ma, but certainly after ca. 340 Ma.

7.2. Contradiction between zircon U-Pb and mica K-Ar and Ar-Ar ages?

The interpretation of the zircon data as indicating late Palaeozoic to earliest Triassic high-pressure metamorphism contradicts the interpretation of older, >400 Ma, Ar-Ar and K-Ar mica ages as indicating growth during Ordovician high-pressure metamorphism (Fukui et al., 1995; Hyodo, 2008; Och et al., 2003, 2007a, 2010).
However, high pressure metamorphic terranes commonly give anomalously old Ar-Ar mica ages, relative to ages of high pressure metamorphism established by U-Pb zircon and Sm-Nd mineral isochrons (Kelley et al., 1994). The apparently too-old ages are most likely the result of excess argon. This is despite the fact that the mica samples might yield apparently acceptable stepped-heating plateau. Three examples of this follow: (a) In the ultra high-pressure terrane of Dabie Shan, China, Ar-Ar mica ages are Neoproterozoic, whereas U-Pb and Sm/Nd ages are ca. 220 Ma (Li et al., 1999). The macroscopic reason for this was investigated by Li et al. (1999). They noted that high-pressure rocks from the same outcrop failed to equilibrate Ar-isotopes between each other, on a centimetre to metre scale. They proposed that excess $^{40}\text{Ar}$ was derived internally from the high-pressure rocks, through the degassing of early-generation (detrital) minerals, but was not completely expelled from the rock body prior to its closure for the isotopic system. This is accord with the ultra-high-pressure metasedimentary rocks containing early Palaeozoic and Precambrian detrital zircons (Maruyama et al., 1998), demonstrating the presence of old crustal material in these rocks. (b) The high pressure rocks of the Dora Maira Massif in the western Alps give ca. 100 Ma Ar-Ar ages, conflicting with ca. 40 Ma zircon U-Pb ages (Arnaud and Kelley, 1995). (c) In a study of blueschist and low-temperature eclogite facies rocks from the Tavsanli Zone of NW Turkey (Sherlock and Kelley, 2002), samples were analysed by the UV Laser Ablation method and yielded apparent ages ranging from 72 to 155 Ma. Detailed in situ intra-grain profiling of phengites, and analyses of sodic-amphibole, lawsonite, quartz and garnet, provided the detailed spatial distribution of excess argon within different minerals in the same samples. This revealed that they acted as near closed systems where excess argon was preferentially partitioned into phengite. They concluded that excess argon was derived in situ, rather than having been introduced by fluids. Therefore, Sherlock and Kelley (2002)
proposed that a combination of stepped-heating and in situ laser spot analyses is required to provide the most reliable understanding of Ar-Ar ages from high pressure metamorphic rocks. They also advised that any Ar-Ar mica ages that are older than the Rb-Sr ages on the same micas should be regarded as suspect, and likely to contain excess argon. In the case of the Port Macquarie Rocky Beach eclogites and blueschists, to our knowledge, Rb/Sr mica ages have not been published. Such Rb/Sr mica ages would help to clarify the meaning of the Ar-Ar ages. Given the general problems of excess argon in low-temperature, high-pressure, Ar-Ar mica geochronology (e.g., Li et al., 1999; Arnauld and Kelley, 1995; Sherlock and Kelley, 2002) and the present lack of Rb/Sr dating of the Rocky Beach phengites, we contend that our zircon ages give a more robust constraint on the timing of metamorphism than the available mica ages.

7.3. Appraisal of conditions of the high-pressure metamorphism at Port Macquarie

Appraisal of the glaucophane schists suggests pressures of ca. 0.7-0.6 GPa, with temperatures of 450-350°C (Och et al., 2003). These conditions coincide with the stability field of glaucophane-bearing assemblages of Maresch (1977) and are in accord with albite being present with the glaucophane (Och et al., 2003). The low temperature for the blueschists is in accord with the widespread coexistence of lawsonite + glaucophane (Barron et al., 1976; Och et al., 2003), because lawsonite is generally regarded as breaking-down above ca. 400°C at these pressures (Newton and Kennedy, 1963; Carson et al., 2000). Thus overall, the Port Macquarie glaucophane schist assemblages record low grade, only moderate high-pressure regimes. This regime is typical for blueschists from the upper parts of subduction zones, such as in the Central Franciscan of California (Oh et al., 1991).
Associated with the Rocky Beach blueschists are eclogites, which can occur together as thin interfolded layers (Fig. 2a). This has been recorded in other low-grade high-pressure rocks, such as in the Franciscan of California (Oh et al., 1991; their Fig. 2). They demonstrated that at low grade, the development of glaucophane-dominated blueschists versus omphacite-bearing eclogites is controlled by variations in Mn/(Mn+Fe) in bulk compositions.

Presently, the only constraint on the pressure-temperature conditions for the Rocky Beach eclogite assemblages are the coexistence of lawsonite with garnet + omphacite, together with a single estimate of ca. 560°C from garnet-omphacite Fe-Mg cation exchange thermometry (Och et al., 2003). Och et al. (2003) based this temperature on a single pair of garnet-omphacite analyses (Tables 1 and 2 of Och et al., 2003). These Tables indicate that there is considerable variation in Fe/Mg between different garnet and omphacite analyses. By using different permutations of the published analyses this would lead to possible calculated temperatures between 900-300°C at 0.7-0.6 GPa (calculated by the Excel™ spreadsheet PX-NORM of Sturm, 2003). Och et al. (2003) did not explain why one particular mineral pair giving 560°C was chosen. Furthermore, we note the rather modest Al₂O₃ content (mostly 10-7 wt%) of the Port Macquarie omphacites, which is similar to those from other low grade (<1 GPa, <450°C) eclogites (e.g., from the Franciscan - Oh et al., 1991), rather than those encountered >1.0 GPa medium to high temperature eclogites. Therefore, the suggestion of a much higher temperature (560°C) and pressure (>1.5 GPa) for the Rocky Beach eclogites compared with the blueschists needs to be substantiated.

Instead, we consider it most likely that the eclogites formed within the same low temperature, moderately high-pressure conditions as the blueschists, with the eclogite-blueschist transition governed by bulk composition. Supporting this interpretation are; the intimate interlayering of eclogites with blueschists (samples
21-06 and RB-2; Fig. 2a), the three times higher Mn/(Mn+Fe) of the eclogite (Table 1),
and as presented in Och et al. (2003) the stability of lawsonite with omphacite and
garnet.

If all the blueschist and eclogite facies rocks formed at low grade conditions at
only <450°C, it is unlikely that new in situ metamorphic zircon overgrowths would
have developed during the high-pressure metamorphism. Thus the grains from
eclogites show the same pitted surfaces as those from the blueschists, without the
development of euhedral, faceted metamorphic overgrowths, as is seen on zircons
definitely grown during higher temperature eclogite facies metamorphism (Fig. 3).

7.3. Late Palaeozoic blueschist-eclogite facies metamorphism at Port Macquarie and
the evolution of the New England Orogen and eastern Gondwana

The Port Macquarie serpentinite mélange is an isolated occurrence on the
coast, and there is no proven connection with the Great Serpentine Belt of the
Peel-Manning fault system inland (Fig. 1a). Thus given the geological complexity of
the New England Orogen, with several Devonian-Permian arc-accretionary
assemblages present, we do not presume here a Late Palaeozoic to earliest Triassic age
for the crustal emplacement of the Great Serpentine Belt, and our following discussion
on tectonic significance will be based solely on our data from the Port Macquarie
serpentinite.

In the Permo-Triassic, eastwards migration of arc activity meant that the
Carboniferous Tablelands-Wandilla subduction complex was the basement of the
250-225 Ma north-south components of the New England batholith and related
volcanic rocks (Fig 1a; Cawood et al., 2011; Rosenbaum et al., 2012). We note that the
youngest detrital zircons in our eclogite and blueschist facies samples coincide with
the inception of this arc activity, which is likely to be their ultimate source. At the
extreme eastern fringe of Australia in south-eastern Queensland is the enigmatic
Gympie terrane, which seems to be a Permian island arc assemblage (Sivell and
McCulloch, 2001). Prior to opening of the Tasman Sea, perhaps this was contiguous
with the Brook Street Terrane in New Zealand and the Teremba Terrane of New
Caledonia (Harrington, 1983; Spandler et al., 2005). We propose that the emplacement
of Port Macquarie serpentinite mélangé carrying the eclogites and blueschists was
related to subduction at the margin of Gondwana and/or to the juxtaposition a
Palaeopacific Ocean Permian island arc (now represented by the dispersed Gympie,
Brook Street and Teremba Terranes) against an Andean eastern Australian margin
(represented by the 260-230 Ma New England Batholith and volcanic components,
Fig. 1a; Cawood et al., 2011; Rosenbaum et al., 2012). This event might be the driving
force behind the Permo-Triassic Hunter Bowen orogeny (Carey and Browne, 1938),
during which basement Carboniferous Tamworth Belt forearc rocks were thrust
westwards over Permian back-arc basin rocks of the eastern fringes of the
Sydney-Gunnedah-Bowen basins (Fig. 1a).

8. Conclusions

(1) Three samples of metasedimentary eclogite and blueschist facies rocks from the Port
Macquarie serpentinite mélangé at Rocky beach yielded detrital zircons with ages from >3000
to 251 Ma.

(2) Metamorphic overgrowths developed in situ during high-pressure metamorphism were not
detected. This is in accord with reappraisal of the conditions of the blueschist and eclogite
facies metamorphism, for which all the available information indicate low-grade for both, with
temperatures of $<450^\circ$C at 0.7-0.6 GPa. If so, this would be in accord with the lack of
syn-eclogite facies metamorphic zircon growth.

(3) Hence, from the U-Pb zircon studies, the youngest detrital grains (251 Ma) give the
maximum age for the high-pressure metamorphism. Lorne Basin alluvial sedimentary rocks
(≥220 Ma) deposited unconformably on Port Macquarie serpentine mélange give the
minimum age for the high-pressure metamorphism.

(4) The zircon U-Pb age constraint for the high-pressure metamorphism and the previously
reported Ordovician Ar-Ar mica ages for this event are contradictory. Our explanation for this
is that the micas contain excess $^{40}$Ar, as is commonly found in micas from low-grade
high-pressure metamorphic terranes. This can be tested by Rb/Sr dating of the micas with
apparent Ordovician Ar-Ar ages.

(5) The new age constraints suggest that the high-pressure metamorphic event at Port
Macquarie occurred during late Palaeozoic to earliest Triassic subduction at the Gondwanan
margin. Cessation of the high pressure metamorphism and emplacement into the crust of the
serpentine mélange carrying blocks of the high pressure rocks might have been related to
juxtaposition of an exotic arc terrane (now represented by the Gympie terrane of southern
Queensland) rafted into a 260-230 Ma Andean arc at the margin of Gondwana. This coincides
with the timing of the 260-230 Ma Hunter-Bowen orogeny.

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zircons from the Franklin metamorphics of Tasmania. Shane Paxton of the Australian National
University is thanked for zircon separations. The paper benefited from the constructive reviews by Peter Cawood and Chris Clark.

References


Geochronological Center Special Publication 4, 70 pp., Berkeley Geochronological Center, Berkeley, California.


Figure captions

Figure 1. (a) Geological sketch map of the southern end of northeastern New South Wales. (b) Synthesis of New England Orogen terranes in northeastern New South Wales and south-eastern Queensland.

Figure 2. Field photographs of eclogite and blueschist blocks at the northern end of Rocky Beach, Port Macquarie. Location of analysed samples is indicated.

Figure 3. Transmitted light photomicrographs of zircons. (a-c) Grains from Rocky Beach eclogite and blueschist facies rocks. (d) Example of an igneous zircon from the Devonian Tacking Point Gabbro, Port Macquarie. (e) Examples of syn-metamorphic (low HREE, no Eu negative anomaly) eclogite facies zircons from the Franklin metamorphics, Tasmania.

Figure 4. Cathodoluminescence images of representative zircons from Rock Beach eclogite and blueschist facies rocks.

Figure 5. Analytical data for minerals from Rocky Beach eclogite and blueschist facies rocks. (a) Summary Tera-Wasserburg $^{238}U/^{206}Pb$ versus $^{207}Pb/^{206}Pb$ plot for zircons. Analytical errors are depicted at the 2 σ level. (b) Chondrite-normalised plot of REE analyses for zircons and garnets. Noted analyses are as follows: (i) 21-06 oval low Th/U ca. 1100 Ma zircon 15.1 and (ii) RB-2 abraded ca. 560 Ma metamorphic rim have negative Eu anomalies and low HREE abundance, showing they equilibrated with plagioclase and garnet; (iii) example of a 500 Ma syn-eclogite facies zircon from the Franklin metamorphics of Tasmania, equilibrated with garnet but not plagioclase and thus shows no Eu anomaly but low HREE abundance (from Fergusson et al., submitted).

Figure 6. Back Scatter Electron image of part of the titanite + rutile porphyroblast RB-4. LA-ICP-MS U-Pb dating sites are indicated.

Figure 7. Tera-Wasserburg $^{238}U/^{206}Pb$ versus $^{207}Pb/^{206}Pb$ plot of rutile and titanite analyses. Analytical errors are depicted at the 2 σ level.

Table Captions

Table 1. Whole rock XRF geochemical analyses.
Table 2. Summary SHRIMP zircon U-Th-Pb data
Table 3. SHRIMP and LA-ICP-MS zircon and garnet rare earth element analyses.
Table 4. Summary LA-ICP-MS rutile and titanite U-Pb data
RB-3
grain 1
250 Ma

RB-3
grain 2
350 Ma

RB-2
grain 2
595 Ma
rim

Tacking Point
Gabbro
magmatic
zircon

syn-eclogite
facies (500 Ma)
metamorphic
zircons,
Tasmania

50 µm
La Ce Pr Nd (Pm) Sm Eu Gd Tb Dy Ho Er Tm Yb Lu

12-06-15.1 is 1100 Ma oval low Th/U grain. -ve Eu anomaly and low HREE shows it equilibrated with plagioclase and garnet.

RB-2-6.1 is an abraded 560 Ma metamorphic rim. Equilibrated with garnet (low HREE) and plagioclase (-ve Eu anomaly).

All other metamorphic (then abraded) rims equilibrated with plagioclase, but not garnet. ~250 Ma and 380-340 Ma grains show the same pattern.

10,000
0.04
0.08
0.12
0
10
20
30
400
600
800
1000
1400
1800
2200
207Pb/206Pb

post-Archean zircon data (samples 21-06, RB-2, RB-3)
common Pb composition

RB4 rutiles and titanites

lower intercept with discordia
332±140 Ma
(MSWD=0.71)
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<td>101.22</td>
<td>99.70</td>
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Mn/(Mn+Fe) | 0.030 | 0.011 | 0.011 |

V        | 416   | 400   | 329   |
Cr       | 260   | 161   | 138   |
Ni       | 64    | 68    | 67    |
Cu       | 111   | 27    | 23    |
Zn       | 120   | 165   | 73    |
Rb       | 61    | 18    | 48    |
Sr       | 182   | 96    | 185   |
Ba       | 1102  | 401   | 1072  |
Y        | 44    | 28    | 33    |
Zr       | 106   | 103   | 275   |
Nb       | 16    | 11    | 36    |
La       | <2    | <2    | 24    |
Ce       | 20    | 2     | 93    |
Hf       | 4     | 3     | 5     |
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Site grain type and analysis location: p=prism, ov=oval, fr=fragment, e=end, m=middle, c=core, r=rim/overgrowth
Site CL imagery: osc=oscillatory zoned, h=homogeneous, hd=homogeneous dark, low luminescence, sz=sector zoned, rex=recrystallised
All analytical errors are given a 1 σ
Concordance and $^{207}\text{Pb}/^{206}\text{Pb}$ ages only given for >1000 Ma old sites
f$^{206}\%$ is the amount of $^{206}\text{Pb}$ modelled as non-radiogenic, based on measured $^{204}\text{Pb}$
### Zircons from 21-06 Eclogite Layer Within Blueschists: LA-ICP-MS in ANU

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### Zircons from RB-2 Blueschist Interlayered with Eclogite: SHRIMP in Hiroshima University

<table>
<thead>
<tr>
<th>Analysis</th>
<th>Site</th>
<th>Age La</th>
<th>Ce</th>
<th>Pr</th>
<th>Nd</th>
<th>Sm</th>
<th>Eu</th>
<th>Gd</th>
<th>Tb</th>
<th>Dy</th>
<th>Ho</th>
<th>Er</th>
<th>Tm</th>
<th>Yb</th>
<th>Lu</th>
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</thead>
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### Zircons from RB-3 Retrogressed High-Pressure Block: SHIMP in Hiroshima University

<table>
<thead>
<tr>
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<th>Site</th>
<th>Age La</th>
<th>Ce</th>
<th>Pr</th>
<th>Nd</th>
<th>Sm</th>
<th>Eu</th>
<th>Gd</th>
<th>Tb</th>
<th>Dy</th>
<th>Ho</th>
<th>Er</th>
<th>Tm</th>
<th>Yb</th>
<th>Lu</th>
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<tbody>
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### Example of Eclogite Metamorphic Zircon the Franklin Metamorphics, Tasmania: SHRIMP in Hiroshima University

<table>
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<th>Analysis</th>
<th>Site</th>
<th>Age La</th>
<th>Ce</th>
<th>Pr</th>
<th>Nd</th>
<th>Sm</th>
<th>Eu</th>
<th>Gd</th>
<th>Tb</th>
<th>Dy</th>
<th>Ho</th>
<th>Er</th>
<th>Tm</th>
<th>Yb</th>
<th>Lu</th>
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<tbody>
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### Garnet from Eclogite 21-06: LA-ICP-MS in Macquarie University
<table>
<thead>
<tr>
<th>Sample</th>
<th>Type</th>
<th>Fracture</th>
<th>Equant</th>
<th>Prismatic</th>
<th>Oval</th>
<th>Grain Fragment</th>
<th>Grain End</th>
<th>Grain Middle</th>
<th>Overgrowth</th>
<th>Core</th>
<th>SHRIMP U-Pb</th>
<th>LA-ICP-MS Approximate U-Pb</th>
</tr>
</thead>
<tbody>
<tr>
<td>2106-1A</td>
<td>fr</td>
<td>0.004</td>
<td>0.003</td>
<td>0.004</td>
<td>0.026</td>
<td>0.047</td>
<td>0.043</td>
<td>0.290</td>
<td>0.063</td>
<td>0.370</td>
<td>0.049</td>
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</tr>
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<td>0.001</td>
<td>0.003</td>
<td>0.023</td>
<td>0.078</td>
<td>0.059</td>
<td>0.410</td>
<td>0.130</td>
<td>1.270</td>
<td>0.250</td>
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</tr>
<tr>
<td>2106-2</td>
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<td>0.004</td>
<td>0.004</td>
<td>0.035</td>
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<td>0.080</td>
<td>0.550</td>
<td>0.130</td>
<td>1.090</td>
<td>0.200</td>
<td>0.600</td>
</tr>
<tr>
<td>2106-3</td>
<td>fr</td>
<td>0.012</td>
<td>0.017</td>
<td>0.008</td>
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<td>0.100</td>
<td>0.069</td>
<td>0.510</td>
<td>0.097</td>
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<td>0.010</td>
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<td>0.480</td>
<td>0.110</td>
<td>1.060</td>
<td>0.320</td>
<td>1.460</td>
</tr>
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<td>2106-5</td>
<td>fr</td>
<td>0.003</td>
<td>0.002</td>
<td>0.003</td>
<td>0.036</td>
<td>0.063</td>
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<td>0.120</td>
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</table>

**garnet from blueschist RB-2: LA-ICP-MS in Macquarie University**

<table>
<thead>
<tr>
<th>Sample</th>
<th>Type</th>
<th>Fracture</th>
<th>Equant</th>
<th>Prismatic</th>
<th>Oval</th>
<th>Grain Fragment</th>
<th>Grain End</th>
<th>Grain Middle</th>
<th>Overgrowth</th>
<th>Core</th>
<th>SHRIMP U-Pb</th>
<th>LA-ICP-MS Approximate U-Pb</th>
</tr>
</thead>
<tbody>
<tr>
<td>RB2-1A</td>
<td>fr</td>
<td>0.002</td>
<td>0.003</td>
<td>0.003</td>
<td>0.027</td>
<td>0.059</td>
<td>0.047</td>
<td>0.360</td>
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<td>0.017</td>
<td>0.008</td>
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<td>0.079</td>
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<td>0.004</td>
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<td>0.004</td>
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</table>

Grain morphology: eq=equant, p=prismatic, ov=oval, fr=grain fragment, e=grain end, m=grain middle, r=overgrowth, c=core
CL imagery: osc=oscillatory zoned, sz=sector zoned, h=homogeneous, hd=homogeneous dark, rex=recrystallised
Age is SHRIMP U-Pb age, apart from *=LA-ICP-MS approximate U-Pb age
<table>
<thead>
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<th>Analysis</th>
<th>U (ppm)</th>
<th>$^{238}\text{U}/^{206}\text{Pb}$</th>
<th>$^{207}\text{Pb}/^{206}\text{Pb}$</th>
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<tbody>
<tr>
<td>R-4</td>
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<td>0.7322 ± 0.0717</td>
</tr>
<tr>
<td>R-5</td>
<td>0.26</td>
<td>2.368 ± 0.208</td>
<td>0.7532 ± 0.0724</td>
</tr>
<tr>
<td>R-6</td>
<td>0.07</td>
<td>1.176 ± 0.053</td>
<td>0.7760 ± 0.0333</td>
</tr>
<tr>
<td>R-7</td>
<td>0.02</td>
<td>1.203 ± 0.126</td>
<td>0.7709 ± 0.0792</td>
</tr>
<tr>
<td>R-8</td>
<td>0.13</td>
<td>1.491 ± 0.109</td>
<td>0.7878 ± 0.0583</td>
</tr>
<tr>
<td>R-9</td>
<td>0.51</td>
<td>1.496 ± 0.047</td>
<td>0.8118 ± 0.0254</td>
</tr>
<tr>
<td>R-9</td>
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</tr>
<tr>
<td>R-10</td>
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<td>0.7566 ± 0.0480</td>
</tr>
<tr>
<td>R-11</td>
<td>0.47</td>
<td>1.532 ± 0.066</td>
<td>0.7223 ± 0.0602</td>
</tr>
<tr>
<td>R-12</td>
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<td>0.536 ± 0.027</td>
<td>0.8912 ± 0.0647</td>
</tr>
<tr>
<td>R-13</td>
<td>0.07</td>
<td>1.173 ± 0.049</td>
<td>0.7166 ± 0.0534</td>
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<td>R-14</td>
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<td>0.7224 ± 0.1608</td>
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<tr>
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<td>0.7889 ± 0.0100</td>
</tr>
<tr>
<td>T-2</td>
<td>2.05</td>
<td>1.029 ± 0.016</td>
<td>0.7893 ± 0.0124</td>
</tr>
<tr>
<td>T-3</td>
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<td>0.8393 ± 0.0134</td>
</tr>
<tr>
<td>T-4</td>
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<tr>
<td>T-5</td>
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<td>0.7947 ± 0.0118</td>
</tr>
<tr>
<td>T-6</td>
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<td>0.8143 ± 0.0119</td>
</tr>
<tr>
<td>T-7</td>
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<td>0.8126 ± 0.0149</td>
</tr>
<tr>
<td>T-8</td>
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<td>0.8151 ± 0.0129</td>
</tr>
<tr>
<td>T-09</td>
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<td>0.801 ± 0.024</td>
<td>0.7768 ± 0.0650</td>
</tr>
<tr>
<td>T-10</td>
<td>1.54</td>
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<td>0.7636 ± 0.0811</td>
</tr>
<tr>
<td>T-11</td>
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<td>0.435 ± 0.013</td>
<td>0.7994 ± 0.0724</td>
</tr>
<tr>
<td>T-12</td>
<td>0.71</td>
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</table>

R=rutile, T=titanite. Analytical errors given at 1σ