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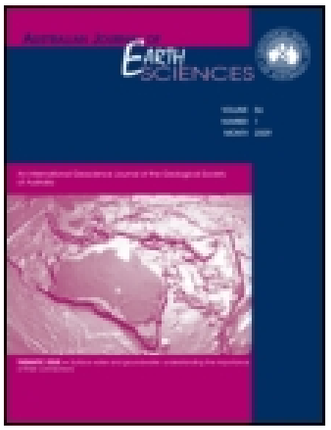

Australian Journal of Earth Sciences: An International Geoscience Journal of the Geological Society of Australia
Publication details, including instructions for authors and subscription information:
http://www.tandfonline.com/loi/taje20

Ordovician to earliest Silurian history of the Macquarie Arc, Lachlan Orogen, New South Wales

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Published online: 06 Oct 2011.

To cite this article: I. G. Percival & R. A. Glen (2007) Ordovician to earliest Silurian history of the Macquarie Arc, Lachlan Orogen, New South Wales, Australian Journal of Earth Sciences: An International Geoscience Journal of the Geological Society of Australia, 54:2-3, 143-165, DOI: 10.1080/08120090601146789

To link to this article: http://dx.doi.org/10.1080/08120090601146789

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Australian Journal of Earth Sciences (2007) **54**, (143 – 165)

Ordovician to earliest Silurian history of the Macquarie Arc, Lachlan Orogen, New South Wales

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Subduction-related volcanic and associated sedimentary successions of Ordovician age in the Junee – Narromine, Molong, Rockley – Gulgong and Kiandra Volcanic Belts are correlated using the most comprehensive biostratigraphic framework yet compiled for the Macquarie Arc. Key features of the correlation include confirmation and dating of four arc-wide magmatic phases: in the Early Ordovician (Lancefieldian – Bendigonian), late Middle to early Late Ordovician (Darriwilian to Gisbornian), Late Ordovician and Late Ordovician to earliest Silurian. Emplacement ages of intrusions that form part of the four magmatic phases complement the biostratigraphic correlations. A widespread hiatus of about 9 million years separates the last-known Bendigonian deposits associated with Phase 1 from initiation of the second magmatic phase in early Darriwilian time. The third magmatic phase comprises the geochemically distinctive Copper Hill Suite of intrusive rocks. The fourth magmatic phase, recognised by its evolved shoshonitic signature, follows an extensive volcanic hiatus marked by development of Eastonian-age shallow-marine autochthonous limestones and overlying deeper water fine-grained clastic rocks in the Junee – Narromine and western Molong Volcanic Belts. During this interval in the eastern Molong Volcanic Belt and Rockley – Gulgong Volcanic Belt, volcanism appears to have been nearly continuous with the changeover from Phase 2 to Phase 4 lavas discernible only from petrographic and geochemical criteria. Isotopic dating indicates that Phase 3 magmatism spans the Eastonian volcanic hiatus, suggesting that this intrusive episode may have been responsible for regional uplift resulting in limestone deposition. A widespread depositional and volcanic break, ranging across the Ordovician – Silurian boundary, reflects variable uplift and cessation of arc volcanism, followed by intrusion of Phase 4 porphyries in the early Llandovery that constitute the last gasp of arc activity.

KEY WORDS: biostratigraphy, Lachlan Orogen, Macquarie Arc, Ordovician, radiometric dating, Silurian, stratigraphy, volcanic belts.

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INTRODUCTION

Subduction-related Ordovician volcanic and volcaniclastic rocks in the Eastern Subprovince of the Lachlan Orogen (Figure 1) occur in four separate structural belts in central and southern New South Wales (Glen *et al.* 1998; Glen 2005). Establishing sequences of events within individual belts, correlating between them, and determining whether the belts represent one or more volcanic arcs, all rely heavily on accurate dating of stratigraphic units. Much new information has come to light in the three decades since publication of Webby's (1976) biostratigraphically based review of the Ordovician stratigraphy of southeastern Australia, which in part correlated formations in the same region (his Macquarie Volcanic Belt) as that considered here. The present study updates and refines previous correlations across the different belts (Glen *et al.* 1998; Percival *et al.* 1999; Packham *et al.* 1999, 2003), employing the most recently published Ordovician time-scale (Webby *et al.* 2004) and incorporating published and new isotopic dating of intrusive events, thus enabling clear identification of four phases of intermediate to mafic volcanism and plutonism, three of which are separated by extensive, sometimes arc-wide, hiatuses (Figure 2). Our study concentrates on the three Ordovician volcanic belts in central New South Wales (Figure 1)—from west to east, the Junee – Narromine, Molong and Rockley – Gulgong Volcanic Belts as defined by Glen *et al.* (1998)—as these are the longest ranging and best-dated successions. The Kiandra Volcanic Belt in the south of the state is also reviewed but in lesser detail, building upon earlier mapping by Owen and Wyborn (1979).

In this paper, we synthesise all available palaeontological and geochronological dating in order to establish as precisely as possible a framework for the geological

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ISSN 0812-0099 print/ISSN 1440-0952 online © 2007 Geological Society of Australia
DOI: 10.1080/08120090601146789

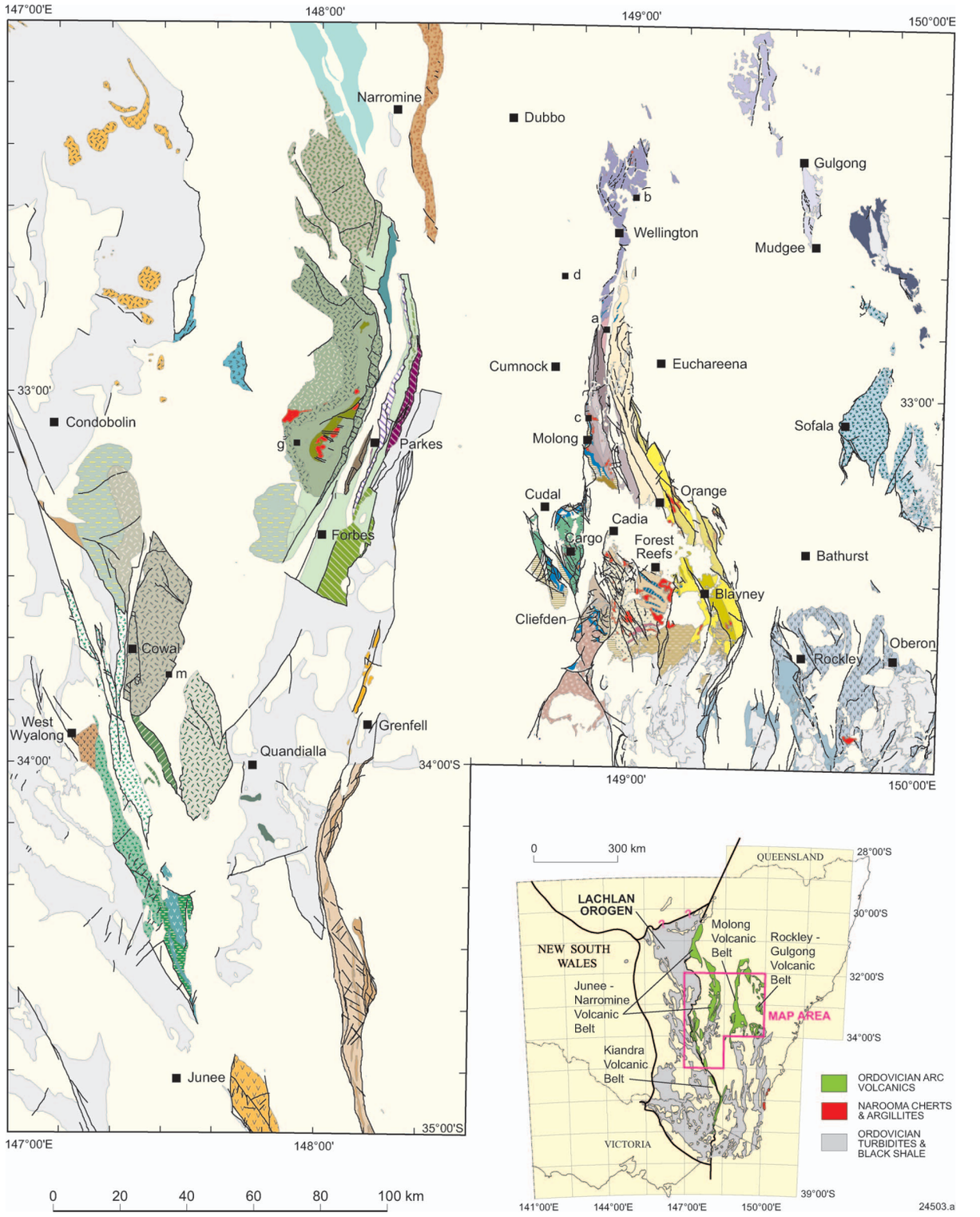

Figure 1 Map of the Macquarie Arc, eastern Lachlan Orogen showing main structural belts in central New South Wales, key localities discussed in the text and Ordovician (except where indicated otherwise) stratigraphic units. Note that some monzonites and other intrusions have ages extending into the early Llandovery; for detailed stratigraphic age ranges and correlations, see Figure 2. The Junee–Narromine Volcanic Belt is confined to the western half of the map; the Molong Volcanic Belt is distributed north–south along longitude 149°E with its eastern margin located to the east of Orange and Blayney; the Gulgong–Rockley Volcanic Belt is situated adjacent to the right-hand margin of the map largely east of Bathurst. The Kiandra Volcanic Belt is not shown on the main map but is depicted on the inset location map in southern New South Wales mainly south of latitude 36°S. a, Bakers Swamp; b, Bodangora; c, Larras Lee; d, Gunnars Dam; g, Gunningbland; m, Marsden prospect.

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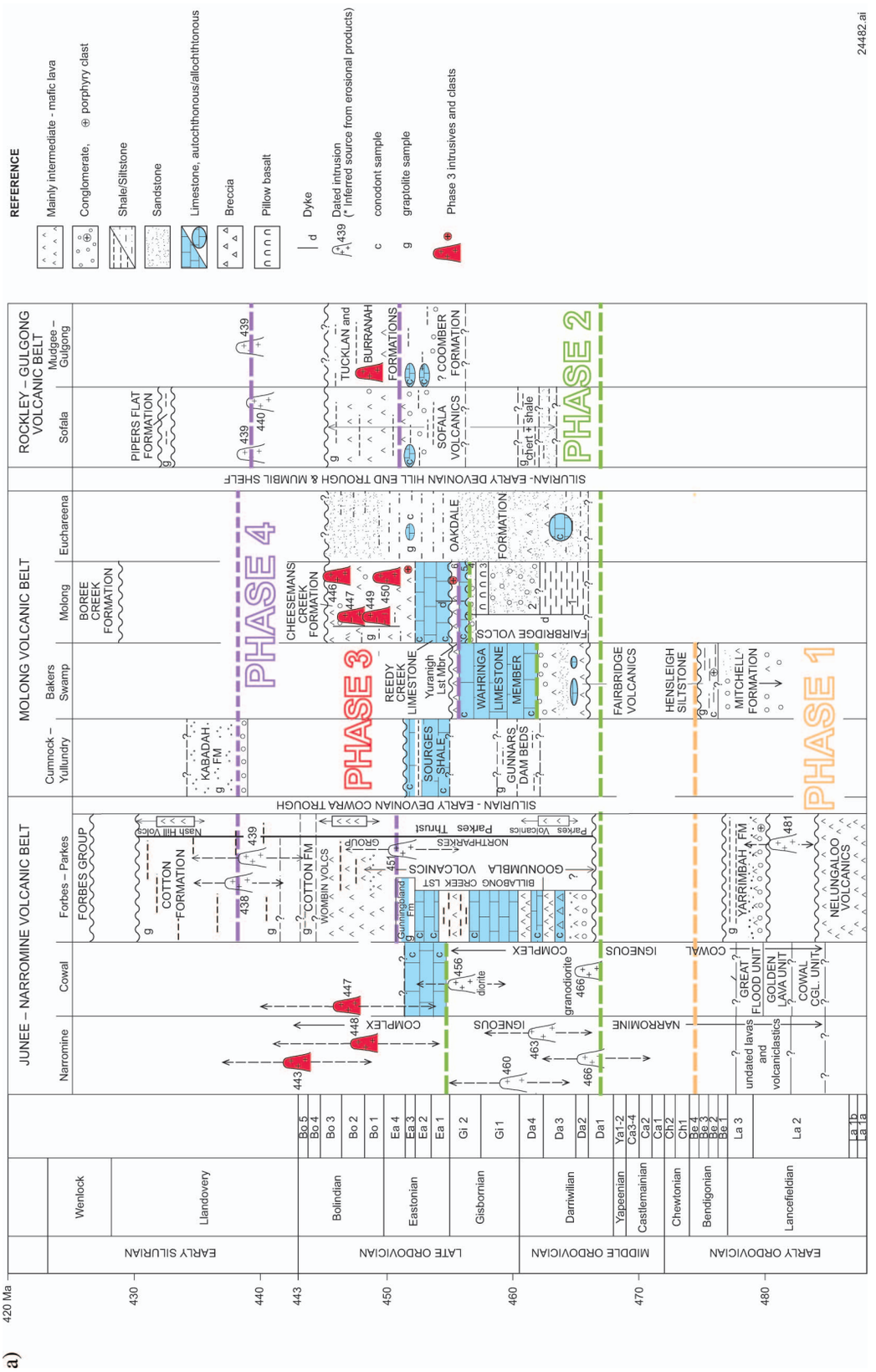
Figure 1 (*Continued*).



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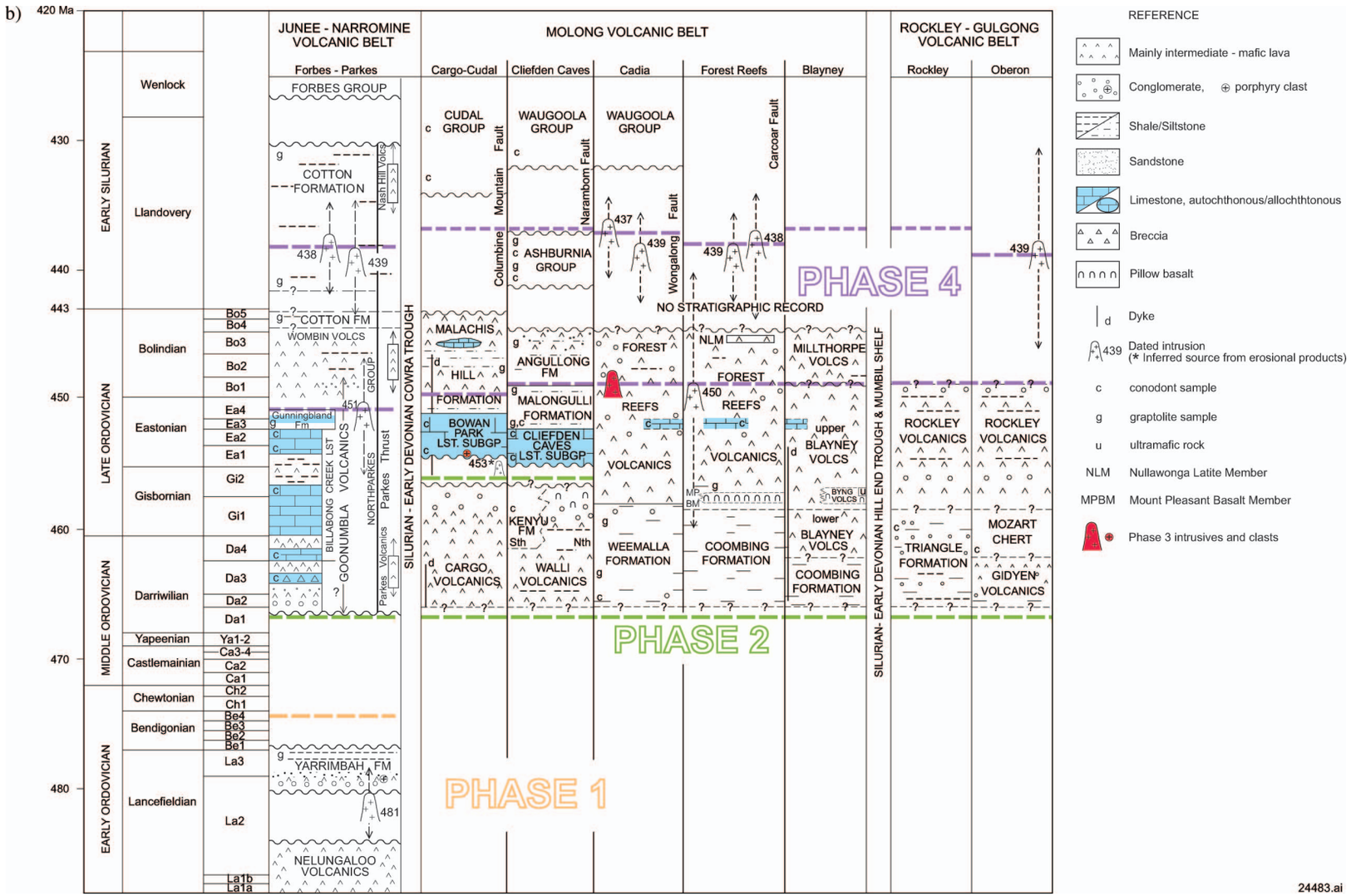

Figure 2 East–west correlation across the Macquarie Arc utilising the time-scale of Webby *et al.* (2004), depicting key elements of biostratigraphic and isotopic age data. The four phases of intermediate to mafic magmatism identified by Crawford *et al.* (2007a) are indicated (Phase 3 intrusions and eroded clasts highlighted in red; the exact positions of the undated Jenkins Road intrusion in the Mudgee–Gulgong column, and the dacite intruding the upper Forest Reefs Volcanics in the Cadia column, are uncertain). (a) Northern area; (b) central and southern areas. Note that the Forbes–Parkes column is common to both charts. Data for each column are discussed in the text, with additional information from the Cadia area (Molong Volcanic Belt) based on Holliday *et al.* (2002). Numbers in Fairbridge Volcanics (Molong column) refer to informal units of this formation as discussed in the text.

evolution of the Macquarie Arc. In particular, we have drawn on the copious geological data assembled over the past decade during second-generation mapping of the Bathurst (Raymond *et al.* 1998), Dubbo (Morgan *et al.* 1999a), Forbes (Raymond *et al.* 2000) and Goulburn (Thomas *et al.* in press) 1:250 000 sheets, primarily by the Geological Survey of New South Wales with assistance from Geoscience Australia in the first three areas.

Previous regional correlations were based on macrofossil assemblages from Upper Ordovician shallow-water carbonate-dominated successions and their lateral clastic equivalents (Webby 1969, 1974, 1975; Webby & Nicoll 1989; Percival 1992; Nicoll & Webby 1996). These have been more recently augmented by conodont faunas (Fowler & Iwata 1995; Percival *et al.* 1999; Murray & Stewart 2001; Pickett & Percival 2001; Zhen *et al.* 2001, 2004a, b; Zhen & Percival 2004a, b), enabling extension of biostratigraphic zonations back to Lower and Middle Ordovician strata. Despite their known potential for biostratigraphic correlation in deeper water sediments, graptolites have generally not been used to constrain regional syntheses of the Macquarie Arc since, with the exception of Moors (1970) and Rickards *et al.* (2001b), modern systematic studies of local faunas are almost entirely lacking.

Our correlations are necessarily constrained by the veracity of the fossil identifications, and by the quality of radiometric dating, the standards used and the extent of statistical and other errors that often encompass more than one graptolite zone or conodont zone. Black *et al.* (2003, 2004) have drawn attention to the problems associated with using standards SL13 and AS3 in SHRIMP age determinations. With these caveats, the arc-wide correlations presented in Figure 2 represent our current best interpretation and serve as the stratigraphic basis for the following papers in this thematic issue.

JUNEE–NARROMINE VOLCANIC BELT

In contrast to the other three belts of Ordovician volcanic rocks, most of the Junee–Narromine Volcanic Belt lies under cover. Accordingly, the distribution of stratigraphic units shown in Figure 1 is inferred from both outcrop and interpretation of aeromagnetic and gravity data. Palaeontological control is largely restricted to rocks exposed in the Parkes–Forbes area, with scanty fossil data available for sedimentary rocks associated with the Cowal and Currumburrama Igneous Complexes. Also briefly discussed are the Gidginbung Volcanics and the Temora Volcanics in the southern part of the belt, the Raggatt Volcanics north of Bogan Gate and the Narromine Igneous Complex between Tomingley and Narromine. Representative columns for the Parkes–Forbes, Cowal and Narromine areas, where palaeontological and isotopic age dates are most reliable, are shown in Figure 2a. The unexposed Canonbar Igneous Complex, situated at the northern extremity of the belt, and other complexes such as the Fairholme Igneous Complex are not considered here as they are described by Glen *et al.* (2007b) and in part by Crawford *et al.* (2007b).

Narromine Igneous Complex

Towards the northern end of the Junee–Narromine Volcanic Belt, near Narromine, the non-exposed Narromine Igneous Complex includes undated lavas and volcaniclastics intruded by late Middle Ordovician monzodioritic to granodioritic rocks. Geochemical affinities of these lavas with the Nelungaloo Volcanics imply a similar Early Ordovician age (Crawford *et al.* 2007b). U–Pb dating of zircon grains in the intrusions by SHRIMP I and II using standards AS3 and SL13 (Fanning 2000) provided the following dates: hornblende monzodiorite, 466 ± 3.6 Ma; monzonite, 463 ± 4 Ma; and monzodiorite, 460 ± 6 Ma. Granodiorite and dacite contained zircons with cores dated 472 ± 6 Ma and 467 ± 5 Ma (Castlemainian to early Darriwilian), respectively and rims representing a younger, Bolindian, crystallisation event at 448 ± 7 Ma and 443 ± 6 Ma, respectively. The report by Fanning (2000) did not specify which standard applied to which age determination. Chemical compositions of the granodiorite and dacite suggest affinities with the Copper Hill Suite of intrusives developed in the Molong Volcanic Belt (Crawford *et al.* 2007a), and the ages determined for the zircon rims accord reasonably well with the range of ages (456–447 Ma) known for this suite in that belt.

Cowal Igneous Complex

The stratigraphy around the Cowal mine (Figure 1) is based on drillcore intersections, there being no surface outcrop. Miles and Brooker (1998) published a threefold subdivision comprising a lower volcaniclastic succession, the Cowal Conglomerate unit, that is conformably overlain by the trachyandesite-dominated Golden Lava unit, itself conformably succeeded by the Great Flood unit (sandstone, siltstone and mudstone with a massive conglomerate) (Figure 2a). A more recent report of the mine stratigraphy (Bywater *et al.* 2004) referred to strata below the Golden Lava unit as the Lower Volcaniclastic unit and those above as the Upper Volcaniclastic unit, all intruded by a large monzodiorite intrusion in the core of a large north-northwest-plunging anticline. Bywater *et al.* (2004) also identified additional units including mudstones east of the mine in an Eastern Volcaniclastic unit, an andesite dome complex above the Upper Volcaniclastic unit, and a breccia complex at depth. There are no palaeontological data for any of these units, but at the Endeavour 42 deposit, they are intruded by the Muddy Lake Diorite (Perkins *et al.* 1995a), which has a K–Ar age of 456 ± 5 Ma (Gisbornian). Other intrusives in this area include granodiorite with an Ar–Ar age of 465.7 ± 1 Ma (early Darriwilian) (Miles & Brooker 1998). On the basis of geochemical similarities with the Nelungaloo Volcanics, Crawford *et al.* (2007b) suggest an Early Ordovician age for the Cowal Igneous Complex.

In the southeast corner of the complex, as yet unnamed limestones intersected in drillholes by Newcrest Mining Limited at their Marsden prospect near Barmedman Creek (Figure 1 locality m) contain various stromatoporoids, halysitid tabulate corals and *Tetradium tenue* together with a diverse conodont fauna

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including *Rhipidognathus* sp., *Phragmodus undatus*, *Yaoxianognathus*? *tunguskaensis* and *Belodina compressa* (Percival *et al.* 2006). These strata are of Eastonian (Ea1 – 3) age and represent the most southerly known extent of Upper Ordovician shallow-water carbonate facies in the Macquarie Arc. ICP-MS U – Pb dating of zircons from a granodiorite from the Marsden prospect gave an age of 447 ± 7 Ma, comparable with similar intrusive rocks of the Copper Hill Suite (Crawford *et al.* 2007a).

Currumburrama Igneous Complex

The Currumburrama Igneous Complex (Figure 1), though entirely buried, is inferred from remote geophysical data (Glen & Fleming 2000). Drilling by Goldminco Corporation (B. Mowatt pers. comm. 2005) has intersected andesitic lavas and volcaniclastics intruded by monzodiorites, diorites, monzonites and granites near the southeastern edge of the complex. West of Quandialla (Figure 1), a graptolite, nautiloid and sponge deep-water assemblage of probable early Bolindian age is preserved in fine-grained spiculitic rocks of the Jingerangle Formation (Percival & Lyons 2000, Percival *et al.* 2006). In lithology, depositional environment and age, these strata resemble the lower part of the Cotton Formation to the north near Forbes. Glen *et al.* (2004) placed the Jingerangle Formation in the Currumburrama Igneous Complex, but in the absence of exposed contacts, the nature of the relationship between sedimentary rocks of the Jingerangle Formation and undated volcanics of the Currumburrama Igneous Complex remains conjectural, and these units are therefore not depicted on Figure 2.

West Wyalong – Junee area

The Gidginbung and Temora Volcanics between West Wyalong and Junee were described by Warren *et al.* (1995) and are further briefly discussed by Glen *et al.* (2007b). The Gidginbung Volcanics consist of strongly altered intermediate volcanic and volcaniclastic rocks, intruded by granodiorites to monzodiorites (Warren *et al.* 1995). Extensively altered breccias and mudstones at Gidginbung may be proximal to a volcanic centre (Lawrie *et al.* 1997). An age of 435 Ma for host-rocks at Gidginbung mine (Perkins *et al.* 1995a) was determined using SHRIMP 1 and is probably too young. A suite of six samples from this area analysed by L. P. Black (pers. comm. 2005) gave ages ranging from 436.0 ± 3.1 to 439.3 ± 3.0 Ma. Hornblende from the Rain Hill Quartz Monzonite that intrudes these volcanics has an Ar – Ar date of 434.9 ± 2.3 Ma (Wormald 1993 cited by Wyborn 1996). A specimen of the index conodont *Distomodus staurognathoides*, recovered from limestone in a sedimentary section immediately overlying (presumably unconformably) the Gidginbung Volcanics, indicates an early Llandovery (Early Silurian) age, approximately 439 – 438 Ma.

Aeromagnetic data indicate that the Temora Volcanics outcrop as two belts consisting of lavas and volcaniclastic rocks (siltstone, sandstone, conglomerate) (Warren *et al.* 1995). The Dobroyde Volcanics of Wyborn (1996) are also included in this unit. As discussed by Wyborn (1996), the Junawarra Volcanics (Warren *et al.* 1995) are not included as part of the arc due to their tholeiitic geochemistry.

Parkes – Forbes area

WEST OF THE PARKES THRUST

The oldest Ordovician rocks recognised in the Junee – Narromine Volcanic Belt (Figure 1) are the Nelungaloo Volcanics (Sherwin 1979, 2000a; Krynen *et al.* 1990; Simpson *et al.* 2005) in the vicinity of Gunningbland (Figure 1 locality g), west of Parkes (Glen *et al.* 2007a). This formation consists predominantly of andesitic lavas and volcaniclastic sediments, lacking internal evidence of age. A minimum age is provided by a monzonitic intrusive originally dated at 484.3 ± 2.9 Ma using zircon U/Pb SHRIMP dating and the AS3 standard (Butera *et al.* 2001), but since revised to allow for underestimation of the correction necessary to compensate for the unusually high U content of the sample (I. Williams, quoted in Simpson *et al.* 2005 p. 866) to 481 ± 4 Ma. This revised age lies within the mid to late Lancefieldian (Early Ordovician), centred in the upper part of the La2 zone. The age of the overlying Yarrimbah Formation, as determined from graptolites in its upper part, ranges from late Lancefieldian (La3) to mid Bendigonian (Be2) (Sherwin 1979). Integration of the isotopic and palaeontologic data within the framework of the Webby *et al.* (2004) time-scale suggests a break in deposition of at least 2 million years in mid to late Lancefieldian time, during which the Nelungaloo Volcanics and associated intrusives were eroded to provide the source of igneous clasts in basal conglomerates of the Yarrimbah Formation (Glen *et al.* 2007b).

The Yarrimbah Formation (Sherwin 2000a) consists predominantly of volcaniclastic sandstone and conglomerate in its lower part, overlain by laminated siliceous mudstone (termed cherts in previous references to the unit). Notable among clasts in the conglomeratic lithologies are rare limestones, which yield microfossils (fragmentary lingulate brachiopods and indeterminate shelly debris), associated with sponges and disarticulated trilobites observed in the matrix of maroon and cream calcarenites. Although not diagnostic of age, the sponges appear identical to *Archaeoscyphia*? described from allochthonous limestones of similar lithology in the Hensleigh Siltstone, south of Wellington (Pickett *in* Percival *et al.* 2001). Graptolites recognised in the upper Yarrimbah Formation include *Didymograptus* (*Cymatograptus*) sp., *Tetragraptus approximatus* and *Pendeograptus fruticosus* (Sherwin 1979, 1990; L Sherwin pers. comm. 2000).

Simpson *et al.* (2005) interpreted the presence of abundant angular crystal fragments, a minor pumice component and predominantly volcanic clast compositions in conglomerates in the lower part of the Yarrimbah Formation as indicating derivation from contemporaneous volcanic eruptions. They suggested that thin horizons of medium-grained and coarsely porphyritic andesite (similar to those in the underlying Nelungaloo Volcanics) near the top of the Yarrimbah Formation implied ongoing volcanism in the form of

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subaqueous lavas, although they also canvassed the possibility that they were high-level sills, an interpretation we prefer. Whichever interpretation is adopted, it appears that Phase 1 magmatism was discontinuously active in this area over approximately 6 million years.

About 250 m of non-exposure intervenes between the top of the Yarrimbah Formation and the next youngest unit, which marks the base of the Northparkes Group, an association of lavas, volcaniclastic deposits, shallow-water carbonates and deeper water clastic rocks. The Northparkes Group, introduced here as a modification (by exclusion of volcanic rocks east of the Parkes Thrust) of the Northparkes Volcanic Group established by Percival (2000), passes up from the basal Goonumbla Volcanics into the Wombin Volcanics at the top. We recognise two units within the Goonumbla Volcanics: a lower part, generally poorly exposed and mainly comprising coarse-grained volcaniclastic rocks, and an upper part characterised by more coherent volcanics including lavas. In the Gunningbland area, a considerable thickness of sedimentary rocks—including the shallow-marine Billabong Creek Limestone and overlying deeper water Gunningbland Formation—separates the lower Goonumbla Volcanics and the upper part of this unit.

The base of the lower Goonumbla Volcanics in the Nelungaloo section, 5 km southeast of Gunningbland village, is interpreted as early Darriwilian in age, although there is neither palaeontological nor isotopic evidence to support this. The oldest datable horizon in the Goonumbla Volcanics occurs an additional 250 m higher up section, where a late Darriwilian (Da3) age is indicated by the presence of the conodont *Pygodus anitae* in a sedimentary breccia comprising angular volcanic clasts in a carbonate matrix (Pickett & Percival 2001).

Higher in the section, the Billabong Creek Limestone (Pickett & Percival 2001) contains late Darriwilian (Da4) conodonts, including *Pygodus anserinus* at its base. This formation, ~320 m thick, comprises mainly thick-bedded to massive limestone, with minor poorly outcropping volcanics and volcaniclastics (best indicated in aeromagnetic data) in its lower half. At or near the Middle–Upper Ordovician boundary, the clastic component disappears. Concomitant with this, a slow deepening in the environment of limestone deposition is reflected by changing facies: very shallow-water pink limestone at the base, with a distinctive stylobedded appearance, is overlain by clean coarse-grained calcarenite with occasional fossil corals. The virtual absence of volcanic detritus in such a marginal-marine depositional environment implies that any such potential sources were either non-existent or below wave-base. Younger beds of early Eastonian (Ea2) age in the Billabong Creek Limestone contain diverse shelly fossil assemblages indicative of deeper water depositional environments extending to the outer shelf edge.

This upward-deepening trend continues into the overlying Gunningbland Formation, the lower part of which (on New Durran, 3 km southeast of Gunningbland) comprises volcaniclastic rocks passing up into feldspar-rich siltstone with a deep-water fauna including trinucleid trilobites, nautiloids, lingulate brachiopods and graptolites of late Eastonian (Ea3) age (Percival 1978; Pickett & Percival 2001). On Sunnyside at Gunningbland, the upper section of the formation (including volcaniclastic sandstones exhibiting local ash beds with feldspar crystals) shows a shallowing-up trend, with re-appearance of limestone in the form of pinnacle reefs composed of corals and stromatoporoids (Pickett & Percival 2001). Offshore topographic gradients were extremely steep, with deep-water graptolitic siltstone (Ea3–4 age) adjacent only tens of metres laterally. Some 10 km northwest of Parkes around Nanardine, shallow-water corals (identical to those forming the pinnacle reefs) are preserved as clasts in sandstone, presumably derived by wave-zone erosion of the Goonumbla Volcanics, and providing evidence of the emergence of volcanic islands (Pickett 1984).

The upper Goonumbla Volcanics (most recently described by Simpson *et al.* 2005) overlie the Gunningbland Formation. They are dominated by volcaniclastic sandstones, conglomerates and ignimbrites as well as coherent basaltic andesites, trachyandesites and rarer trachytes. These rocks, together with the overlying Wombin Volcanics, represent the main phase of Late Ordovician (predominantly latest Eastonian to Bolindian) volcanism in the Parkes area that produced feldspar-phyric lavas (Crawford *et al.* 2007a). Intrusion of a monzodiorite from Goonumbla with a U/Pb SHRIMP zircon age (AS3 standard) of 450.8 ± 4.2 Ma (Butera *et al.* 2001), i.e. latest Eastonian time, was probably coincident with the resumption of volcanic activity.

Poorly exposed Ordovician rocks in the Bogan Gate–Trundle area, 40 km northwest of Gunningbland, include andesitic to trachyandesitic lavas, tuffs, volcaniclastic breccias and possibly allochthonous fossiliferous limestones. These were assigned to the Raggatt Volcanics and correlated with the Goonumbla Volcanics by Sherwin (1996) on the basis of a Late Ordovician coral, stromatoporoid and conodont fauna listed by Pickett and Ingpen (1990). The Eastonian (Ea1–3) age indicated by these fossils suggests correlation with (and possibly derivation from) the upper Billabong Creek Limestone and limestones in the Gunningbland Formation. The Raggatt Volcanics are included in the Northparkes Group.

The Cotton Formation (Sherwin 2000b) is the youngest dated Ordovician unit in the Forbes–Parkes region and consists predominantly of fine-grained siltstone, probably turbiditic in part, with isolated calcareous horizons containing a shallow-water coralline (allochthonous?) assemblage: the latter, reported by Andrews (1910), has not been able to be relocated (possibly removed during roadworks: L. Sherwin pers. comm. 1999). Graptolites are present at three levels, with possible mid to late Bolindian faunas (undescribed) in the lower part, and both Early and Late Llandovery (Early Silurian) assemblages in the upper part (Sherwin 1974). The lower Cotton Formation exhibits little or no sign of volcanic input, implying that it post-dated eruption of the Wombin Volcanics to the north. However, we know of no demonstrated contact between these two units. Continuity of deposition in the Cotton Formation across the Ordovician–Silurian boundary has been inferred by Sherwin (1970, 1973), but as yet no

evidence of latest Bolindian or earliest Llandovery graptolites has been found to support his contention. An alternative interpretation places a time break between these two somewhat different units, with only the lower section being part of the Macquarie Arc, and the upper part post-dating the first phase of accretion of the arc (Glen *et al.* 2007c).

Latest Ordovician to Early Silurian (Llandovery) intrusions represent the youngest magmatic events in this belt. U/Pb SHRIMP dating (standard unspecified) of zircons from a biotite – quartz monzonite and a late trachytic dyke at Northparkes has yielded ages of 445 ± 3.9 and 437 ± 6 Ma, respectively (Lickfold *et al.* 2007), broadly comparable with previously published U/Pb SHRIMP dates (SL13 standard) of 438.5 ± 3.6 Ma from host volcanics (Perkins *et al.* 1995a), and monzonite dates of 439.1 ± 4.5 and 438.9 ± 4.7 Ma (AS3 standard) by Butera *et al.* (2001). Dating of biotite and alteration sericite gives consistent ages around 439 Ma (Lickfold *et al.* 2007), confirming previous Ar – Ar dates for alteration sericite at Goonumbla of 439.2 ± 1.2 and 439.2 ± 5 Ma (Perkins *et al.* 1995a).

EAST OF THE PARKES THRUST

Mafic and intermediate volcanics occur in the high-strain zone situated between the Parkes Thrust and the Mugincoble Fault. The latter fault forms their eastern boundary with Ordovician turbidites and cherts of the Kirribilli Formation. Although no age data are available for these units, which include the Goobang, Nash Hill, Parkes, Bushmans, Mingelo, and Daroobalgie Volcanics, it has been previously assumed (Krynen *et al.* 1990) that they were all Ordovician, with the Daroobalgie Volcanics overlain by the Cotton Formation (Krynen *et al.* 1990). Note that in an alternative interpretation (Lyons *et al.* 2000), the Parkes Volcanics included the Daroobalgie Volcanics and the Nash Hill Volcanics was expanded to include the Back Yamma and Bushmans Volcanics. Geochemical work by Crawford *et al.* (2007a) suggests that the Daroobalgie and Mingelo Volcanics are probably correlatives of the Goonumbla Volcanics. However, high Nb, Th, U and LREE enrichment of the Nash Hill and Bushmans Volcanics, compared to known Ordovician volcanics and volcaniclastics in the area, implies these are probably Early Silurian (Llandovery) in age (Crawford *et al.* 2007a).

MOLONG VOLCANIC BELT (NORTHERN REGION)

Bakers Swamp area

The oldest unit identified in the Molong Volcanic Belt, the Mitchell Formation (Morgan & Scott 1999), is predominantly a volcaniclastic unit (Glen *et al.* 2007a) with minor lavas. Its boundary with the overlying Hensleigh Siltstone at Wahringa in the Bakers Swamp area, 26 km south of Wellington (Figure 1 locality a), is conformable and transitional. Carbonate clasts occur sporadically in the uppermost part of the Mitchell Formation; one of these (a maroon-coloured breccia) yielded the sponge *Archaeoscyphia*?, similar to that described by Pickett (*in* Percival *et al.* 2001) from an allochthonous carbonate block of comparable lithology in the lower part of the Hensleigh Siltstone. This transitional nature of the Mitchell/Hensleigh contact implies that the age of the Mitchell Formation is most likely late Lancefieldian to earliest Bendigonian, given the presence of upper *Prioniodus elegans* Conodont Zone faunas in both autochthonous and allochthonous carbonates from the lower part of the Hensleigh Siltstone (Zhen *et al.* 2001, 2004a). This age is consistent with evidence of a Bendigonian (Be2 – 3) age determined from the graptolite fauna in the upper beds of the Hensleigh Siltstone (Percival *et al.* 1999, 2001).

A hiatus (in both magmatism and deposition) from late Bendigonian to early Darriwilian time (about 9 million years: Figure 2a) is inferred between the top of the Hensleigh Siltstone and the base of the Fairbridge Volcanics. During this interval, the Hensleigh Siltstone was emergent and eroded, as evidenced by the presence of clasts of reworked Lower Ordovician limestone in the lower Fairbridge Volcanics (Percival *et al.* 1999). However, it is not clear whether the earlier Mitchell volcano was also being eroded or had already subsided.

The age of andesitic lavas in the lower Fairbridge Volcanics is constrained by rare occurrences of allochthonous limestones yielding conodonts of Da3 age (Zhen *et al.* 2004b). Higher in the section, boulder beach conglomerates (derived from the underlying lavas) are overlain by the 88 m-thick late Darriwilian – late Gisbornian Wahringa Limestone Member. Other evidence for shallow-water deposition at the base of the Wahringa Limestone Member includes hematised sandstone (interpreted as heavy-mineral beach sand) with occasional grazing trails, and laterally equivalent pink calcarenites bearing a distinctive shallow-water fossil association (macluritid gastropods and sponges) (Percival *et al.* 1999, 2001). Concentration of vein quartz and felsic volcanic detritus in the basal part of the limestone was noted by Packham *et al.* (2003). The Wahringa Limestone Member deepens upsection, to water depths of $\sim$20 m (below wave-base) estimated from strophomenid brachiopods and stromatoporoids preserved in growth position in the middle and upper beds of the unit. Conodonts from the lowermost beds of the member have been precisely dated as *Pygodus serra* Zone of late Da3 to early Da4 age (Zhen *et al.* 2004b), while those from the upper part have a late Gisbornian (Gi2) aspect. The Fairbridge Volcanics above the Wahringa Limestone Member consists of volcaniclastics, rare lavas and occasional lenses of allochthonous limestone, none of which yielded any biostratigraphically significant data.

Molong – Larras Lee area

A very different stratigraphic sequence in the Larras Lee area (Figure 1, locality c) and extending south to Molong, 35 km south of Bakers Swamp, implies lateral facies changes in the Middle Ordovician. At Larras Lee, Crawford *et al.* (2007a) have defined six informal subdivisions of the Fairbridge Volcanics. Unit 1, a basal sandstone – siltstone succession with andesitic clasts, and Unit 2 (volcaniclastic-dominated) are overlain by a 300 m-thick pillow basalt (Unit 3). Crawford *et al.* (2007a)

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recognise a major change in geochemistry from medium- and high-K to a shoshonitic signature in Unit 4 of the Fairbridge Volcanics, consisting of andesitic conglomerate and pseudobreccias, that crops out immediately south of Molong. Unit 5 is lenticular and includes the 180 m-thick Yuranigh Limestone Member, the basal part of which consists of lithic-dominated, thin-bedded carbonate rocks (with vein quartz detritus: Packham *et al.* 2003) containing abundant shelly debris typical of a turbulent shallow-water environment (a beach or strandline deposit). Increasingly deeper water deposition is indicated by marly limestones in the upper part of the Yuranigh Limestone Member, containing large strophomenid brachiopods that most probably inhabited quiet-water environments below wave-base (Percival *et al.* 2001). Late Gisbornian (Gi2) conodonts (Percival *et al.* 1999) provide the only age control in this section, indicating correlation with the uppermost Wahringa Limestone Member (Zhen *et al.* 2004b) that was being deposited at similar depths in the Bakers Swamp area to the north. The Yuranigh Limestone Member is succeeded by volcaniclastic rocks (Unit 6), ~1300 m thick, of the uppermost Fairbridge Volcanics that are equivalent to the volcaniclastic-dominated section above the Wahringa Limestone Member.

The overlying Reedy Creek Limestone marks the beginning of a major hiatus in arc volcanism (Glen *et al.* 1998; Percival *et al.* 1999; Packham *et al.* 2003), which conodont and macrofossil evidence indicates spanned virtually the entire Eastonian, throughout the western Molong Volcanic Belt. There is no discernible hiatus between cessation of volcanism (with associated volcaniclastics) and initiation of Eastonian carbonate deposition in the northern part of the belt. In common with other Eastonian carbonates on the arc, the Reedy Creek Limestone displays progressively decreasing terrigenous influx during deposition; thin-bedded mud-rich calcarenites in its lower part are replaced by more massive and pure carbonates upsection.

Quartz diorite and younger dacite that intrude the Fairbridge Volcanics and the Reedy Creek Limestone at Copper Hill, just north of Molong, represent a geochemically distinctive magmatic phase (Crawford *et al.* 2007a). Hornblende in quartz diorite at Copper Hill gave a K–Ar age of 447 ± 5 Ma (Chivas *in* Scott 1978), close to a younger dacite with a K–Ar hornblende age of 446 ± 6 Ma (A. R. Chivas pers. comm. *in* Perkins *et al.* 1995a). Hornblende tonalite at Copper Hill yielded an inferred emplacement age of 450 ± 6 Ma (U/Pb SHRIMP, standard not quoted: Geoscience Australia OZCHRON database sample 93844531). A hornblende K–Ar age of 449 ± 4 Ma (Chivas *in* Scott 1978) was obtained from the Emu Hollow Dacite, 5 km south of Copper Hill.

The main Late Ordovician phase of volcanism (late Eastonian to predominantly Bolindian in age) in the northern part of the Molong Volcanic Belt is represented by the Cheesemans Creek Formation. The lower part of this unit consists of volcaniclastic sandstone and extensive basalt–andesite sills. Minor conglomerates contain clasts of dacitic porphyry similar to those forming intrusives at nearby Copper Hill. The upper section is dominated by andesite flows, some of which are pillowed (Crawford *et al.* 2007a). Precise dating of these strata is dependent on reassessment of mostly undescribed graptolite faunas: the late Gisbornian age assigned to one assemblage by Rickards *et al.* (2001b) is contentious, whereas a late Eastonian to early Bolindian age interpreted for the same fauna by VandenBerg (2003) accords well with the stratigraphic context.

East of the Molong–Copper Hill area (Figure 1), the Molong Volcanic Belt is represented by the Oakdale Formation that consists of distal turbidites with less common debris-flow deposits. A late Darriwilian (Da3) age for the lower part of this formation is indicated by conodonts in allochthonous limestone pods (Zhen & Percival 2004a) west of Euchareena (Figure 1). About 8 km northeast of Molong on Barham Winchester, deep-water calcareous siltstones with Eastonian graptolites are assigned to the upper Oakdale Formation (Morgan *in* Pogson & Wyborn 1994).

Bodangora area

Northeast of Wellington (Figure 1, locality b), rocks mapped as the Oakdale Formation on the Dubbo 1:250 000 geological map (Morgan *et al.* 1999a) are more similar lithologically to parts of the Cheesemans Creek Formation. These strata are largely volcaniclastic in the west with more lavas and breccias in the east, and are intruded by monzonites. Thin deep-water autochthonous spiculitic limestones, interbedded with Eastonian age graptolitic siltstones, are imbricated with Lower Silurian limestones (bearing halysitid corals and pentamerid brachiopods) in the hangingwall of the Macquarie Thrust west of Bodangora. Eastonian strata at Bodangora, on Wahringa property near Bakers Swamp, and on Barham Winchester northeast of Molong are comparable in age and facies with the Malongulli Formation in the southern Molong Volcanic Belt, implying widespread development of deep-water depositional environments along the entire eastern margin of this belt at this time.

Cumnock–Yullundry area

Biostratigraphic relationships between the Kabadah Formation and Sourges Shale along the northwestern margin of the Molong Volcanic Belt need to be further resolved. The original mapping of the Kabadah Formation suggested a Devonian age (Maggs 1963), subsequently revised to Silurian (Bradley *in* Pickett 1982 p. 82). A Late Ordovician to Early Silurian age was assigned to this unit during remapping of the Bathurst and Dubbo 1:250 000 geological sheet areas (Raymond *et al.* 1998; Morgan *et al.* 1999a), partly on the basis of its strong K-rich radiometric signature, similar to that of other Ordovician volcanics and volcaniclastic units in the region. Fossil corals in calcareous sandstones in a representative section of the Kabadah Formation along Loombah Creek provide definitive evidence of an Early Silurian depositional age (J. Pickett pers. comm. 1999). We therefore believe that the distinctive red-coloured radiometric response derives from redeposited K-rich Ordovician detritus.

Cherty rocks with thin feldspar-rich laminations immediately west of Gunnars Dam (Figure 1, locality d)

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were also included in the Kabadah Formation during mapping of the Dubbo sheet area (Morgan *et al.* 1999b). These rocks yield poorly preserved graptolites of probable Late Ordovician aspect (A. H. M. VandenBerg cited in Morgan *et al.* 1999b) and conodonts no older than Darriwilian age (I. Stewart cited in Morgan *et al.* 1999b). The Gunnars Dam area is structurally complex; field observations by R. A. Glen and S. Meffre suggest a fault separates the cherty beds from the Oakdale Formation to the east, with another fault separating the cherty strata from more sandy and silty Silurian rocks to the west. Hence, these Ordovician deep-water deposits are physically separate from the redefined Kabadah Formation as well as being quite distinct in their palaeontological and lithological character. We refer to them here as the informal 'Gunnars Dam beds' of Mid? to Late Ordovician age.

The Sourges Shale near Cumnock was assigned a Late Ordovician age, based on the presence of shallow-water limestone with Eastonian (Ea1–3) corals and conodonts (Percival *et al.* 1999). The discrepancy in ages and environments between this limestone, and immediately overlying strata with Llandovery graptolites, also previously attributed to the Sourges Shale, was discussed by Morgan (1999). Mapping by I. Percival and J. Pickett (modifying the geological map published by Percival *et al.* 1999) has identified a limestone pod (possibly allochthonous) with Early Silurian corals, overlying the Upper Ordovician limestones in the immediate vicinity of latitic volcanics just northeast of Cumnock. Lack of proven Bolindian age strata between the Eastonian limestone and the Lower Silurian sedimentary rocks suggests that a disconformity separates the two, in which case the unnamed Lower Silurian succession may correlate with the Kabadah Formation.

MOLONG VOLCANIC BELT (CENTRAL REGION)

Cargo Block

The Cargo Block is separated from other Ordovician volcanics in the Molong Volcanic Belt by the Columbine Mountain Fault, interpreted as a west-dipping thrust (Glen 1998), with latest movement occurring in the Carboniferous (see also Simpson *et al.* 2007). The Columbine Mountain Fault (and the Molong Fault to the north) separates a contiguous eastern flank of the Molong Volcanic Belt from a parautochthonous part to the west comprising the Cargo Volcanics and (to the north) the Gunnars Dam beds and Sourges Shale, previously discussed, that lie in the hangingwalls of west-dipping post-Silurian thrust faults. We imply here that, prior to thrusting, these units were not adjacent to the main Molong Volcanic Belt but lay an unknown distance further west, where they formed basement to the Cowra Trough.

The term 'Cargo Andesite' (introduced by Stevens 1957) was subsumed into the Fairbridge Volcanics by Pogson and Watkins (1998). However, because of its demonstrably separate nature, both volcanically and geochemically (Simpson *et al.* 2007), the original name is here reinstated in a modified form (Cargo Volcanics) to reflect the sequence of andesites passing up into volcaniclastics and coarse conglomerates. The Cargo Volcanics contain no fossils.

Unconformably overlying the Cargo Volcanics is the 564 m-thick Bowan Park Limestone Subgroup. The basal unit of this succession, the Ranch Member of the Daylesford Limestone, is distinguished by coarse clastic to conglomeratic facies containing volcanic detritus (Semeniuk 1970, 1972), including dacite clasts. A distinctive quartz-rich sandstone at the base of this unit was equated by Webby and Packham (1982) and Packham *et al.* (2003) to the uppermost member of the Fossil Hill Limestone of the Cliefden Caves Limestone Subgroup, implying that carbonate deposition in the Cargo Block was initiated later in Ea1 time than on the east side of the Molong Volcanic Belt. The presence of an intrusive monzonite dated at 453 ± 4 Ma is inferred from detrital zircons of this age in the basal sandstone (Simpson *et al.* 2007). However, this date (derived by ICP-MS using the Temora standard) suggests an Eastonian 2 age for intrusion, which conflicts by about 1–2 million years with the mid-Eastonian 1 age postulated for the onset of carbonate-dominated deposition in the Cargo Block.

Very-low-diversity faunal assemblages, dominated by lingulid and rhynchonellid brachiopods, characterise the Ranch Member. Succeeding members of the Daylesford Limestone, and the overlying Quondong Limestone, contain shelly faunas that indicate generally progressive deepening to normal shelf water depths of ~30–50 m (Percival & Webby 1996). An abrupt deepening/drowning event in late Eastonian time is inferred from probable submarine channels that are filled with shallow-water fossiliferous calcarenite debris, represented by the Downderry Limestone Member of the Ballingoole Limestone (near the top of the Bowan Park Limestone Subgroup). Semeniuk (1972) interpreted both the lower and upper boundaries of the Downderry Limestone Member as disconformities; we concur, while noting that the lower contact encompasses the more significant time break.

Overlying laminated lithic calcilutites grading up to siliceous siltstones of the Malachis Hill Formation contain graptolites that define a faunal assemblage of late Eastonian (Ea3–4) age, together with conodonts (Zhen *et al.* 1999) comparable with those in the lower Malongulli Formation near Cliefden Caves to the south (Trotter & Webby 1995). A similar depositional environment of several hundred metres water depth can be inferred. The term Malachis Hill Formation (Stevens 1957) is preferred over the Cheesemans Creek Formation (Sherwin 1971) on the basis of publication priority in the Bowan Park district near Cudal, and because it was probably physically separate from the Cheesemans Creek type area (west of Orange) during deposition. The upper part of the Malachis Hill Formation represents the fourth phase of arc magmatism, consisting of a mixture of volcaniclastic detritus and subordinate andesitic primary lavas (Stevens 1957; Semeniuk 1970; Simpson *et al.* 2007). Revised identifications of graptolites collected from siltstones immediately above the andesites (locality g3 of Semeniuk 1970) support an early Bolindian age, most probably Bo2 (G. H. Packham pers.

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comm. 2005). Shallow-water limestone lenses of probable allochthonous origin towards the top of the Malachis Hill Formation contain a Bolindian coral fauna younger than any other known in the region (Webby 1969; McLean & Webby 1976). Although the exact age of this fauna is unclear—several attempts at extracting conodonts from the limestone have proved fruitless—it must be at least mid-Bolindian (Bo3) and could be as young as Bo4, given the stratigraphic separation of nearly 300 m measured by Semeniuk (1970) between the graptolitic horizon and the limestone lenses.

Not shown on the correlation chart for this area (Figure 2b) is the Upper Ordovician Millambri Formation, a largely turbiditic unit composed of volcaniclastic sandstone and conglomerate containing clasts of volcanics. This formation and the underlying silt-rich Rockdale Formation (Ryall 1966), are exposed in the Cranky Rock Anticline located on the Belubula River due south of Cargo. The Canomodine Limestone, beneath the Rockdale Formation, contains corals and stromatoporoids equivalent in age to those in the Quondong and Ballingoole Limestones (Bowan Park Limestone Subgroup). The Rockdale Formation has been merged with the time-equivalent Malongulli Formation by Krynen and Pogson (1998) but as the two units are in different blocks, it would be prudent to reinstate the former. Although lacking definitive Ordovician fossils, the Millambri Formation is probably correlative in age with the Angullong Formation and the Malachis Hill Formation. Both Ryall (1966) and Percival (1976) erred in attributing dark fine-grained graptolitic siltstones in the vicinity of Liscombe Pools (5 km west of Cliefden) to the Millambri Formation; these middle Llandovery strata more likely belong to the Ashburnia Group.

MOLONG VOLCANIC BELT (SOUTHERN REGION)

Volcanic activity dominates the Middle Ordovician record in this region (Figure 2b), especially around, and south of, Cliefden Caves, where primary volcanism is represented by the Walli Volcanics. Cessation of volcanic activity prior to the Eastonian allowed deposition of fringing carbonate-dominated sediments. These carbonates are not present east of the Narambon Fault, where Darriwilian (upper Middle Ordovician) deep-water sedimentation (represented by the turbiditic Weemalla and Coombing Formations) was succeeded, most probably in early Gisbornian times, by predominantly volcaniclastic rocks of the Forest Reefs Volcanics (Packham *et al.* 1999).

Cliefden Caves area

On superpositional evidence, the Walli Volcanics are the oldest unit in the Cliefden Caves area but are undated in the absence of isotopic or palaeontological data. The overlying Cliefden Caves Limestone Subgroup is of early Eastonian (Ea1–2) age, based on conodont faunas (Zhen & Webby 1995). Packham *et al.* (2003) implied a hiatus, approximating the late Gisbornian to earliest Eastonian interval, between the top of the Walli Volcanics (which they correlated with the top of the Cargo Volcanics at Bowan Park) and the base of the limestones. Given the presence of pillow lavas (indicative of submarine volcanism) and structural discordance (Smith 1967), and evidence for active erosion in very shallow water, we argue for a significantly earlier cessation of volcanism in the Cliefden Caves area to allow time for uplift (Figure 2b). Further, we correlate the upper part of the Walli Volcanics with the Kenyu Formation in the Boorowa area to the south, which contains limestones with early Gisbornian conodonts.

Smith (1967) interpreted the relationship between the Walli Volcanics and the Cliefden Caves Limestone Subgroup as an angular unconformity, based on differences in structural trends in the Limestone Creek area where the upper part of the volcanics consists largely of primary andesites (including some with prominent pillows). Elsewhere, the top of the Walli Volcanics immediately beneath the Fossil Hill Limestone in the type section on Fossil Hill comprises spectacular volcaniclastic conglomerates. The prevalence of subangular and poorly sorted clasts of a diverse range of lithologies (including andesitic volcanics and jasper and the presence of detrital vein and igneous quartz) indicates that volcanism ceased, with these rocks subsequently uplifted and eroded, prior to the onset of carbonate deposition.

High clastic content in the lower part of the Fossil Hill Limestone (particularly mudstones of the basal Gleesons Member and calcarenites in the Wyoming Limestone Member) indicates continuing contemporaneous erosion from an emergent volcanic island (Webby & Packham 1982) or reworking of volcaniclastic detritus. The terrigenous component dramatically ceases above a 3 m-thick lithic sandstone at the top of the Fossil Hill Limestone (Webby & Packham 1982; Packham *et al.* 2003), implying that the volcanic source had eroded to sea-level by this time. Cessation of erosion provides a time-line enabling correlation between the Cliefden and Bowan Park limestone successions. Similarity between these two areas is further indicated by identical successions of benthic faunal assemblages indicative of overall progressively increasing water depth (Percival & Webby 1996), although a series of transgressions and regressions (possibly related to local isostatic adjustments and erosional cycles) are recognised in the Fossil Hill Limestone (Webby & Packham 1982; Webby & Percival 1983). Subsequent carbonate deposition on a slowly subsiding island platform or atoll persisted for 3–4 million years, culminating in development of a widespread shallow-marine shelf facies that is recognisable across and along the Molong Volcanic Belt.

Instability and drowning of the carbonate platform began after Eastonian 2 time and are reflected in: (i) cutting of submarine valleys into the Vandon Limestone at the top of the Cliefden Caves Limestone Subgroup; and (ii) deposition of the deep-water Malongulli Formation, disconformably above the Cliefden Caves Limestone Subgroup (Webby & Packham 1982). The basal late Eastonian (Ea3) part of the Malongulli Formation

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contains limestone breccia derived in part from the upper part of the Bowan Park Limestone Subgroup (or equivalents such as the Canomodine Limestone), now lying 35 km to the north-northwest. These latter areas thus remained at shallow depths or became emergent after drowning of the eastern margin of the Molong Volcanic Belt (Webby & Packham 1982).

The fourth phase of arc volcanism in the Cliefden area is represented by the Angullong Formation, which consists of primary andesite and volcaniclastic rocks such as conglomerate, sandstone and siltstone. Near Malongulli Trig, south of the Belubula River 5 km west of Cliefden, the Angullong Formation conformably overlies the Malongulli Formation that contains a prolific early Bolindian (Bo1) graptolite assemblage (including *Climacograptus uncinatus*) in its uppermost beds (Percival 1976; Jenkins 1978). Age constraints in the Angullong Formation are sparse, but the upper beds of this unit north of the Belubula River on Koonoona include deep-water laminated siltstone bearing graptolites (Jenkins 1978), now recognised as mid-Bolindian (Bo3) in age. At least two, possibly three, graptolite zones appear to be missing across the Ordovician–Silurian boundary in an area known as The Basin, just west of the boundary with Angullong (Jenkins 1978), although non-fossiliferous conglomerates and volcaniclastic sandstones at the exposed top of the Angullong Formation may have been deposited in late Bolindian time (Glen *et al.* 2007c). Overlying shallow-water limestones of the Wire Gully Limestone Member, at the base of the Ashburnia Group (Pogson 1998), yield conodonts correlated with the *vesiculosus* graptolite Zone of early Llandovery age (Bischoff 1986), indicating a hiatus of up to 3 million years.

Cadia and Forest Reefs areas

Late Ordovician volcaniclastics and lavas of the Forest Reefs Volcanics in the vicinity of Cadia mine conformably overlie the Weemalla Formation (Holliday *et al.* 2002; Wilson *et al.* 2003). Conodonts from calcareous strata towards the exposed base of the Weemalla Formation on the Belubula River suggest an early Darriwilian (Da2) age (Zhen & Percival 2004b), and at nearby Cadiangullong Creek, south of Cadia, siltstone slightly higher in the formation contains a diverse graptolite assemblage of Da3 age, listed (but not illustrated or described) by Smith (1966). A potential constraint on the age of the upper part of the unit is provided by an unconfirmed record (Taylor 1988) of the indicator graptolite for the early Gisbornian, *Nemagraptus gracilis*.

The Forest Reefs Volcanics were subdivided by Squire (2001) and Squire and McPhie (2007) into a lower feldspar-phyric part dominated by feldspathic sandstone, breccia, basalt and mudstone, and an upper pyroxene-phyric part characterised by the presence of trachyandesite and sandstone that are less feldspathic and richer in mafic volcanic lithic grains and ferromagnesian crystal fragments. Around the Cadia and Ridgeway deposits, Wilson *et al.* (2003) recognised five lithofacies in the Forest Reefs Volcanics. Three of these are sedimentary associations with a significant volcaniclastic component. Primary igneous activity comprises basalt and andesite flows (lithofacies 4) and comagmatic porphyritic dykes (lithofacies 5). Lithofacies 3 of Wilson *et al.* (2003) includes calcareous horizons containing late Eastonian (Ea3) conodonts; these beds were previously included in the upper Weemalla Formation by Packham *et al.* (1999) but have subsequently been reassigned to the upper Forest Reefs Volcanics (Wilson *et al.* 2003; Zhen & Percival 2004b). Accumulation of the Forest Reefs Volcanics spanned much of the Late Ordovician (Figure 2b), although it is probable that volcanism and associated deposition of volcaniclastic sediments were not continuous throughout this interval. Hiatuses are suspected but are difficult to quantify due to lack of a continuous fossil record. In the Forest Reefs area, discordant strike trends in the formation are interpreted by Squire and McPhie (2007) as evidence for an internal unconformity.

Age control for the base of the Forest Reefs Volcanics in the vicinity of Junction Reefs ($\sim$18 km south of Cadia) is provided by occurrence of a trilobite, *Shumardia* sp. described by Webby (1973), associated with poorly preserved late Darriwilian to Gisbornian graptolites. Occurrence of similar graptolites in siltstones of the lower Forest Reefs Volcanics, about 150 m stratigraphically above the Mt Pleasant Basalt Member in Mandurama Ponds Creek (Stevens 1957), provides the only biostratigraphic constraint on initiation of Forest Reefs volcanism. In Figure 2b, we assign a late Gisbornian age to the graptolite occurrences, while recognising the imprecision of this interpretation. Limestone in the upper Forest Reefs Volcanics around the Sheahan-Grants mine in the Junction Reefs area has a late Eastonian (Ea3) age comparable to the calcareous horizons sampled at Cadia (Packham *et al.* 1999).

Near Forest Reefs, the Forest Reefs Volcanics overlie the Coombing Formation, with the boundary placed at the changeover from feldspar-phyric to pyroxene-phyric lavas below the base of the Mt Pleasant Basalt Member (Squire 2001; Squire & McPhie 2007). This implies that the Weemalla Formation is not present east of the Wongalong Fault (*contra* Packham *et al.* 1999 figure 2: see discussion in Zhen & Percival 2004b). Laser ablation ICP-MS U/Pb dating of detrital zircons at around 468 ± 4 Ma indicates that the Coombing Formation was mainly derived from an early Darriwilian volcanic source (Crawford *et al.* 2007a). Based on this evidence, together with the inferred age from graptolites in the lower Forest Reefs Volcanics that overlie both the Weemalla and Coombing successions, we place the top of the Coombing Formation in the Forest Reefs region near to the top of the Weemalla Formation. This view differs from that of Packham *et al.* (2003) who tentatively accepted an Early Ordovician age for the Coombing Formation, based on Ar–Ar plateau ages between 480 and 467 Ma from monzodiorites at Sheahan-Grants mine (Perkins *et al.* 1993) that appear to be older than the enclosing strata.

In the Cadia area, a hiatus separates the Upper Ordovician Forest Reefs Volcanics from unconformably overlying Lower Silurian sedimentary rocks including limestones. Late Llandovery age conodonts were recovered from limestones encountered in exploratory drilling by Pacific Copper at Cadia (Packham *et al.* 1999), and limestone intersected in drillcore 2 m above the

Forest Reefs Volcanics south of Cadia contains a diverse earliest Wenlock conodont assemblage (Rickards *et al.* 2001a). Both conodont faunas indicate correlation with levels high in the Waugoola Group. Intrusive activity during this Early Silurian hiatus is represented by monzonite to monzodioritic bodies of the Cadia Intrusive Complex (Holliday *et al.* 2002), with U/Pb SHRIMP ages of 439 ± 6 Ma on the Cadia Hill Monzonite (Geoscience Australia OZCHRON database sample 93844535: standard unspecified), and 437.1 ± 3.5 Ma calibrated against both SL13 and AS3 (Squire & McPhie 2007). One undated quartz dacite intrusion, located 5 km west of Cadia, is petrographically distinct from the Cadia Intrusive Complex and shows affinities to dacitic intrusions at Copper Hill near Molong (Holliday *et al.* 2002).

Blayney area

The southeastern part of the Molong Volcanic Belt is dominated by basalt and basaltic andesite of the Byng Volcanics and Blayney Volcanics. The clinopyroxene-phyric and olivine-phyric lower part of the Blayney Volcanics is readily distinguished from the upper part of this formation, which is plagioclase-phyric (Crawford *et al.* 2007a). The age of the boundary of the lower Blayney Volcanics with the underlying Coombing Formation is palaeontologically and isotopically unconstrained but is probably late(?) Darriwilian, i.e. older than the inferred age of the Mt Pleasant Basalt Member in the nearby Forest Reefs – Junction Reefs area. Limestone bodies (although mainly recrystallised and barren) near the top of the upper Blayney Volcanics include the Cowriga Limestone Member that contains conodonts with a generalised Late Ordovician age (R. S. Nicoll cited in Wyborn & Krynen 1998). While cognisant of the imprecision of this dating, we have chosen to correlate this limestone with others in adjacent areas that have accurate late Eastonian ages (Figure 2b). Crawford *et al.* (2007a) correlate the Byng Volcanics with the geochemically similar upper part of the Blayney Volcanics and distinguish a new unit, the Millthorpe Volcanics (with strong petrographic and geochemical affinities to the Forest Reefs Volcanics) for rocks between Millthorpe and Blayney that were previously mapped as Byng Volcanics by Meakin *et al.* (1997).

ROCKLEY – GULGONG VOLCANIC BELT (SOUTHERN REGION)

Rockley area

At Triangle Creek, 6 km south-southwest of Rockley in the southwestern part of the Rockley – Gulgong Volcanic Belt, co-occurrence of the conodonts *Periodon aculeatus* and *P. grandis* indicates an early Gisbornian (Gi1) age for siliceous siltstones several hundred metres stratigraphically below the faulted top of the volcaniclastic Triangle Formation (Fowler & Iwata 1995; Crawford *et al.* 2007a). We concur with Fowler and Iwata (1995), who suggested correlation of the overlying Rockley Volcanics with the Sofala Volcanics of the northern Rockley – Gulgong Volcanic Belt.

Oberon area

The oldest unit recognised from this area is the Budhang Chert, previously referred to as a member of the Triangle Formation by Murray and Stewart (2001), but here accorded formation status. This unit contains a deep-water Early Ordovician conodont fauna, including *Paracordylodus gracilis*, indicating an age range of late Lancefieldian to early Bendigonian (La3 – Be2). Stratigraphic relationships between the isolated fault-bounded outcrop of the Budhang Chert and younger rocks are not clearly understood at present. We suspect that affinities of this chert lie with the turbidite association of the Adaminaby Superterrane, rather than with the Macquarie Arc.

Definitely part of the arc is a succession of massive pyroxene basalt lavas and breccias, with clasts of lava and chert, named the Gidyen Volcanic Member (of the Triangle Formation) by Murray and Stewart (2001). This is overlain by the Mozart Chert, which these authors also referred to as a member of the Triangle Formation. Their nomenclature is at variance with that of Stuart-Smith *et al.* (1998), who had earlier recognised an unnamed chert at the base of the Rockley Volcanics, passing up into a mafic to ultramafic volcanic unit and then a talc schist – sandstone – siltstone unit. The Mozart Chert, of late Darriwilian – early Gisbornian age (based on the presence of the conodonts *Periodon* and *Pygodus*), is in turn succeeded by the Rockley Volcanics (Rockley Volcanic Member of Murray & Stewart 2001), which comprise a package of volcaniclastic material together with metabasalt, amphibolite and ultramafic rocks (Murray & Stewart 2001).

Murray and Stewart's (2001) concept of a Triangle Formation (subdivided into four members) to include all Lower to Upper Ordovician units in the southern Rockley – Gulgong Volcanic Belt differs from relationships in volcanic belts further west that are characterised by major hiatuses in volcanic activity, and is not adopted here. Our revised view of the stratigraphy and terminology (Gidyen Volcanics, Mozart Chert, Rockley Volcanics and Triangle Formation) is shown in Figure 2b.

The Rockley Volcanics are intruded by the Swatchfield Monzonite, originally recognised by Stuart-Smith *et al.* (1998). This body has been remapped by Meffre (2003) who determined a U/Pb zircon ICP-MS age of 437 ± 8 Ma using the Temora standard (Crawford *et al.* 2007a).

ROCKLEY – GULGONG VOLCANIC BELT (NORTHERN REGION)

Sofala – Mudgee – Gulgong area

In the northern Rockley – Gulgong Volcanic Belt, the Sofala Volcanics consist mainly of deep-water turbiditic volcaniclastic rocks, cherts, minor lavas and intrusive rocks (Watkins 1998). The oldest age constraint comes from Packham (1968 p. 115) who reported poorly preserved graptolites, apparently *Glyptograptus teretiusculus*, which could indicate a late Darriwilian to

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Gisbornian age, from a horizon a little below the middle of the formation. He also mentioned the presence of a very thick chert unit at about this level. This suggests a potential correlation with the Mozart Chert in the southern Rockley – Gulgong Volcanic Belt. No further graptolites were located during recent remapping of the Bathurst sheet by the Geological Survey of New South Wales, and sampling of various cherts in the Sofala area failed to find diagnostic conodonts.

Conodonts date the maximum age of allochthonous blocks of limestone in the upper part of the Sofala Volcanics in the Wattle Flat area as late Eastonian (Ea3) (Percival 1999). The youngest age recorded from the Sofala Volcanics is mid-Bolindian (Bo2 – 3), based on very poorly preserved graptolites (A. H. M. VandenBerg cited in Rickards *et al.* 1998). The Sofala Volcanics are intruded by Early Silurian monzonite and diorite with Ar – Ar plateaux ages of 438.5 ± 0.8 and 439.9 ± 0.8 Ma (Perkins *et al.* 1995b).

The oldest overlying unit, the Pipers Flat Formation of Pickett *et al.* (1996), is here redefined (fieldwork by I. Percival, R. Glen & L. Barron) as comprising only slates and interbedded siltstones with mid- to late Llandovery graptolites. The overlying Tanwarra Shale passes upwards from basal conglomerates into cross-bedded and then planar-bedded, traction-current sandstones, a calcareous horizon with corals of latest Wenlock to earliest Ludlow aspect, and then into shales with graptolites (Pickett *et al.* 1996). This interpretation of an upwards-deepening succession in the Tanwarra Shale essentially reverts to the original definition of Packham (1968). Inclusion of these uppermost Wenlock limestones within the uppermost part of the Sofala Volcanics by Carr *et al.* (2003 p. 321) disregards remapping and redefinition by Pickett *et al.* (1996), Watkins (1998), and Glen and Watkins (1999), as well as ignoring the impact of Early Silurian hiatuses (Figure 2a).

Ordovician units in the Mudgee – Gulgong area include the Coomber, Burranah and Tucklan Formations (Meakin & Morgan 1999). The Coomber Formation, interpreted as a deeper water volcaniclastic-dominated apron to the arc by Pemberton *et al.* (1994) and Fergusson and Colquhoun (1996), is undated. Ages of the Burranah and Tucklan Formations, determined from conodonts in discrete limestone lenses, are no older than Eastonian (Ea1 – 3) (Percival 1999). The Burranah Formation is intruded by an undated dacite along Jenkins Road, near Mudgee (A. J. Crawford pers. comm. 2001) that is petrographically similar to the Copper Hill Suite of intrusives in the Molong Volcanic Belt. An Early Silurian monzodiorite with a U/Pb SHRIMP age of 435.4 ± 4.6 Ma (Meakin & Morgan 1999 appendix 1) also intrudes the Burranah Formation, although this age may be too young by 3 – 4 million years as it was derived using the SL13 standard.

KIANDRA VOLCANIC BELT

The Kiandra Volcanic Belt is a north-northeast-trending, fault-bounded package of Ordovician volcanics and associated rocks that crosses the New South Wales – Victoria border, from the Benambra area in east Gippsland north to the Brindabella Range southwest of Canberra (Figure 1). Age data, whether palaeontological or radiometrically determined, are sporadic and appear to confirm only the presence of Darriwilian to Gisbornian rocks. Ordovician volcanics in the New South Wales part of the Kiandra Volcanic Belt, comprising the Kiandra Group and the Gooandra Volcanics, have been described by Wyborn (1977), Owen and Wyborn (1979), Wyborn (1992) and Meffre and Scott (1999), and from the Victorian sector by Allen (1987, 1988, 1991) and Stolz *et al.* (1997).

The only fossil ages presently known come from the Kiandra Group. In the Victorian part of the belt, the Kiandra Group is represented by the Blueys Creek Formation, which consists of the Banksia Chert Member (thin-bedded chert) overlain by the Brumby Mafic Arenite Member (chert and volcaniclastic turbidites). Conodonts from the Banksia Chert Member include *Periodon aculeatus*, *P.* cf. *grandis* and *Belodina* sp. (I. Stewart *in* Allen 1987, cited in Allen 1988), indicative of a probable Gisbornian age.

In New South Wales, the Kiandra Group consists of the Temperance Formation (interbedded volcanic sandstone, siltstone and cherts), and the overlying Nine Mile Volcanics (volcanic breccias and flows) (Owen & Wyborn 1979; Meffre & Scott 1999). At least part of the Temperance Formation is mid-Darriwilian to early Gisbornian in age, based on the presence of the conodont *Pygodus* sp.

The Nine Mile Volcanics consist largely of volcaniclastic sandstone, siltstone and conglomerate. Also present are shoshonitic andesitic volcanics and less common strongly pyroxene-phyric basaltic lavas (Meffre & Scott 1999). Age control in the Nine Mile Volcanics is provided by graptolites in shale above an undated limestone, possibly from near the top of the formation. Sherrard's (1954) graptolite identifications from this level, reinterpreted by A. H. M. VandenBerg (pers. comm. 1996), include *Dicellograptus divaricatus salopiensis*, *Pseudoclimacograptus scharenberg*, *Mesograptus multidens*, *Orthograptus calcaratus*, *Amplexograptus arctus* and *Hallograptus mucronatus*. This assemblage has a Gisbornian aspect.

High-K to shoshonitic pyroxenite – gabbro – monzonite sills intrude the Nine Mile Volcanics. One such intrusion, the Doubtful River Gabbro, also possibly intrudes the Temperance Formation and gave a K – Ar hornblende date of 455 ± 10 Ma (approximately earliest Eastonian), interpreted as the age of intrusion by Owen and Wyborn (1979). Late stage, subduction-related cross-cutting hornblende-phyric calc-alkaline dykes are also present (Meffre & Scott 1999).

Excluded from the Kiandra Group are tholeiitic coherent volcanics. The MORB-like Gooandra Volcanics of Owen and Wyborn (1979) have a pronounced Nb anomaly, indicating a subduction influence in their mantle source (arc- or backarc-like tholeiites) (Meffre & Scott 1999). The Jagungal Basalt, localised in a fault zone and consisting of pillow basalts associated with folded cherts, also has MORB-like characteristics (Meffre & Scott 1999). The associated cherts contain small organic filaments of possible cyanobacterial affinity, but these have no precise age connotation.

CORRELATIONS BETWEEN VOLCANIC BELTS

The Ordovician history of the three volcanic belts in central New South Wales involves four magmatic phases (Figure 2), the initial two separated by a major hiatus during which there was arc-wide uplift and erosion. The second and fourth magmatic phases are separated (especially in the two western belts) by an interval of widespread carbonate deposition in Eastonian time. The third magmatic phase, with a distinctive geochemical signature, involves intrusions largely contemporaneous with this Late Ordovician volcanic hiatus.

First magmatic phase (Early Ordovician *ca* 489–474 Ma)

The first episode of volcanism in the Junee–Narromine Volcanic Belt is represented by Lower Ordovician lavas of the Nelungaloo Volcanics, intruded by monzonites (481 ± 4 Ma). Lavas in the Narromine and Cowal Igneous Complexes are correlated on geochemical similarities (Crawford *et al.* 2007b).

In the northern Molong Volcanic Belt, the conformable transition between the Mitchell Formation and the Hensleigh Siltstone is no older than Bendigonian 2. This suggests that the Mitchell Formation (conglomerate with clasts of lavas) largely correlates with the conglomeratic lower part of the Yarrimbah Formation (given that the upper part of this unit is no younger than Be2), rather than the Nelungaloo Volcanics, in the Junee–Narromine Volcanic Belt. Such a correlation implies that the first phase of Ordovician volcanism ceased slightly earlier in the Junee–Narromine Volcanic Belt than in the Molong Volcanic Belt. The similarity of the sedimentary succession in both the Hensleigh Siltstone and Yarrimbah Formation suggests almost identical settings adjacent to eroding volcanic edifices.

Early to Mid-Ordovician magmatic hiatus

No sedimentary, volcanic or intrusive rocks with ages between latest Bendigonian and early Darriwilian have yet been found in any of the four volcanic belts in New South Wales.

In the Junee–Narromine Volcanic Belt, an interval of ~9 million years separates the top of the Yarrimbah Formation and the base of the Goonumbla Volcanics. A similar gap, probably of slightly shorter duration, occurs in the Molong Volcanic Belt, between the top of the exposed Hensleigh Siltstone and the base of the Fairbridge Volcanics. Although the critical areas are in part covered by regolith, there appears to be a true hiatus in volcanism, although we cannot completely discount the possibility that any volcanics erupted or sediments deposited in this interval were eroded before Phase 2 magmatism began.

The nature of the hiatus is constrained by: (i) a possible low-angle discordance between the trends of the Hensleigh Siltstone and the Fairbridge Volcanics; (ii) mild tectonism responsible for uplift and erosion of limestone clasts from Hensleigh Siltstone into the lower Fairbridge Volcanics; the low Colour Alteration Index (CAI = 2) of conodonts, identical in both the Hensleigh Siltstone carbonates and those clasts redeposited into the Fairbridge Volcanics, implies that the Hensleigh Siltstone was probably not buried deeply nor strongly deformed during the hiatus; and (iii) similarity in geochemistry of volcanics in the lower part of the Goonumbla Volcanics (below or interbedded with the lowermost Billabong Creek Limestone) with that of the underlying Nelungaloo Volcanics (Crawford *et al.* 2007b), which may reflect uplift of the Nelungaloo Volcanics and reworking of eroded material into the younger volcanics (Simpson *et al.* 2005 p. 873).

In summary, the extensive latest Early to Mid Ordovician hiatus appears to be characterised by considerable uplift (though there is evidence of only minor structural disruption) following the cessation of the Early Ordovician magmatic phase and associated clastic sedimentation.

Second magmatic phase (late Middle to early Late Ordovician *ca* 466–457/454 Ma)

In the Junee–Narromine Volcanic Belt, Phase 2 volcanism is initially represented by andesitic lavas (termed the basal mafic unit by Crawford *et al.* 2007a) and volcaniclastic sedimentary rocks of the lower Goonumbla Volcanics that underlie and occasionally are interbedded with the Billabong Creek Limestone in the Gunningbland area. Phase 2 also includes intrusive rocks emplaced into the Narromine Igneous Complex (466–460 Ma) and Cowal Igneous Complex (465–456 Ma).

Phase 2 magmatism is widespread in the Molong Volcanic Belt. It is represented by the Fairbridge, Cargo and Walli Volcanics in the west, and the Blayney and Byng Volcanics together with the Mt Pleasant Basalt Member at the base of the Forest Reefs Volcanics in the east, as well as by porphyry intrusions. Volcanism was apparently initiated around Da2 time (466 Ma), based on conodonts from the lowermost Weemalla Formation. Definitive age control for the base of the other units is lacking. This 466 Ma age is identical with that inferred for initiation of Phase 2 in the Junee–Narromine Volcanic Belt. The second magmatic episode concludes in the western Molong Volcanic Belt in late Gisbornian time, prior to widespread deposition of the Eastonian carbonate-dominated successions. The end of Phase 2 is not as clearly defined stratigraphically in the eastern Molong Volcanic Belt, where the Eastonian magmatic hiatus is not developed.

In the southern Rockley–Gulgong Volcanic Belt, Phase 2 magmatism is represented by the Triangle Formation and Gidyen Volcanics, and continues into the overlying Rockley Volcanics. In the absence of palaeontological or isotopic age control, we somewhat arbitrarily equate the beginning of Phase 2 in the Rockley–Gulgong Volcanic Belt with our best estimate for onset of this episode in the volcanic belts to the west (Figure 2b). Rudimentary but acceptable palaeontological data higher in the succession support broad correlation of the volcaniclastic-dominated upper part of the

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Triangle Formation which contains Gisbornian conodonts (Fowler & Iwata 1995), with the upper Darriwilian – lower Gisbornian Mozart Chert. This same level is potentially equivalent to the lower chert-rich part of the Sofala Volcanics in the northern Rockley – Gulgong Volcanic Belt, although this correlation is reliant on an old identification of a graptolite that remains unverified.

Late Ordovician volcanic hiatus

This hiatus is best defined in the western Molong Volcanic Belt (Glen *et al.* 1998; Percival *et al.* 1999; Packham *et al.* 2003), where thick carbonate accumulations in the Cliefden, Cargo, Cumnock and Molong areas during the Eastonian stage (455 – 450 Ma) mark a significant hiatus in arc volcanism. Thicknesses of 360 – 460 m for the Cliefden Caves Limestone Subgroup, 564 m for the Bowan Park succession and about 850 m (Pogson & Watkins 1998: probably an overestimate) for the Reedy Creek Limestone indicate the stability of the carbonate platform during this hiatus in arc volcanism. Precise dating based on conodont identifications from the Cliefden Caves Limestone Subgroup (Zhen & Webby 1995), the equivalent Reedy Creek Limestone (Percival *et al.* 1999), and the Daylesford, Quondong and Ballingoole Limestones of the Bowan Park Limestone Subgroup (Zhen *et al.* 1999) indicates that carbonate deposition lasted from Ea1 to Ea2 on the eastern side of the Molong Volcanic Belt, and slightly longer (mid-Ea1 to Ea3) on the western part of the belt. The volcanic hiatus continued after cessation of carbonate deposition, with deeper water clastic sedimentation including minor limestones extending into early Bolindian (Bo2) time in the Bowan Park area (lower Malachis Hill Formation), and concluding slightly earlier (Bo1) at the top of the Malongulli Formation near Cliefden Caves.

No equivalent hiatus occurs in the eastern Molong Volcanic Belt where volcanism and associated volcaniclastic deposition were largely continuous throughout the Eastonian (Packham *et al.* 1999). Instead, thin limestone beds of Ea3 age are present in the upper part of the Forest Reef Volcanics and near the top of the upper Blayney Volcanics (Packham *et al.* 1999). Coeval limestones lenses scattered throughout volcaniclastics of the northern Rockley – Gulgong Volcanic Belt (Percival 1999) have previously been regarded as allochthonous, but the only potential source area known of the appropriate age is on the western Molong Volcanic Belt (Ballingoole, Cargo and Canomodine Limestones), far from the Forest Reefs Volcanics and units further east. Alternatively, these limestones may represent cappings on seamounts that briefly reached depths suitable for limestone accumulation. Implications are discussed below.

West of Parkes in the Junee – Narromine Volcanic Belt, interbedded volcanics, volcaniclastics and limestones that characterise the pre-lower Gisbornian section of the Billabong Creek Limestone, were replaced during the Eastonian hiatus in arc volcanism by near-continuous carbonate deposition overlain by mixed clastics and carbonates extending nearly to the top of the Eastonian. Unnamed limestone known only from subsurface drillcore in the Marsden area is also of Eastonian age.

Third magmatic phase (Late Ordovician *ca* 456 – *ca* 443 Ma)

This phase comprises intrusive rocks of dacitic composition, forming the geochemically distinctive Copper Hill Suite defined by Crawford *et al.* (2007a). Most known Phase 3 intrusive rocks have isotopic ages that post-date the early Eastonian limestones. However, there is evidence that the earliest intrusives with a Copper Hill Suite affinity pre-date these carbonate rocks, and were eroded during the uplift associated with the Eastonian volcanic hiatus. The presence of felsic hornblende-bearing clasts, petrographically similar to the Copper Hill intrusives, in volcaniclastic rocks at the top of the Fairbridge Volcanics (Crawford *et al.* 2007a) suggests that a Phase 3 intrusion was unroofed in latest Gisbornian time (*ca* 456 Ma) during or immediately after deposition of the Yuranigh Limestone Member. Most Copper Hill Suite rocks, which cluster around 451 – 448 Ma, occur as dacitic and granodioritic intrusions in the Narromine and Cowal – Marsden areas (Junee – Narromine Volcanic Belt), and in the Cheesemans Creek Formation in the vicinity of Copper Hill (Molong Volcanic Belt) (Figure 2). One undated intrusion of this type is known from the Cadia area, while another is reported from the northern Rockley – Gulgong Volcanic Belt. The youngest known intrusion assigned to the Copper Hill Suite is a dacite from the Narromine Igneous Complex, from which the rims of zircon crystals have been dated at 443 ± 6 Ma, centred on the Ordovician – Silurian boundary (Crawford *et al.* 2007b). Thus, available dating suggests that Phase 3 magmatism extends over $\sim$13 million years, spanning the Late Ordovician volcanic hiatus between Phases 2 and 4. Whether this spread of ages is real or just reflects different isotopic dating techniques is unclear.

Fourth magmatic phase (Late Ordovician – earliest Silurian *ca* 457/454 – *ca* 438 Ma)

In the Junee – Narromine Volcanic Belt, the fourth magmatic phase is represented by the shoshonitic upper part of the Goonumbla Volcanics and the overlying Wombin Volcanics, by inferred Late Ordovician volcanics east of the Parkes Thrust, and by porphyries. Other Upper Ordovician volcanic and volcaniclastic units are inferred to occur further south towards Quandialla (Currumburrama Igneous Complex) though the geochemical signature of these rocks is unknown. In the absence of conclusive biostratigraphic evidence to the contrary, we consider stratigraphic continuity across the Ordovician – Silurian boundary to be highly unlikely, and suggest that the widespread hiatus at this level is indicative of significant tectonism with associated intrusive activity (Glen *et al.* 2007c).

Phase 4 magmatism is most clearly defined in the western Molong Volcanic Belt (Crawford *et al.* 2007a), where it is represented by shoshonitic volcanic and volcaniclastic rocks (including the upper Cheesemans Creek, Angullong, Millambri and upper Malachis Hill

Formations) that overlie Eastonian limestones and deep-water clastic rocks. Here Phase 4 began in latest Eastonian to early Bolindian time (no earlier than 450 Ma) and had apparently ceased by Bo4 time (around 444 Ma).

In the eastern and northern Molong Volcanic Belt, Phase 2 and Phase 4 rocks are distinguished by an upward geochemical change from volcanics with medium- to high-K calc-alkaline to shoshonitic affinity, to volcanics entirely of shoshonitic affinity (Crawford *et al.* 2007a). This changeover is recognisable in the Fairbridge Volcanics (Figure 2a) in volcanics overlying the Yuranigh Limestone Member in the Molong area and the Wahringa Limestone Member in the Bakers Swamp area (Crawford *et al.* 2007a), constraining the initiation of Phase 4 to latest Gisbornian or earliest Eastonian time.

In the eastern Molong Volcanic Belt, definite Phase 4 volcanics and volcaniclastics are represented by the upper Forest Reefs Volcanics at Cadia and Forest Reefs – Junction Reefs areas, and the Millthorpe Volcanics near Blayney. The Forest Reefs Volcanics are intruded by Llandovery porphyries dated around 439–435 Ma that are also assigned to Phase 4. Published ages of *ca* 435 Ma are based on U–Pb SHRIMP dating using standard SL13, and may be 1% too young compared to the Temora standard (Black *et al.* 2003). Hence, these ages probably also cluster around 439–438 Ma.

Farther north, the inception of shoshonitic magmatism is especially difficult to determine in the deep-water finer grained volcaniclastic rocks of the Oakdale Formation to the east.

Phase 4 in the southern Rockley – Gulgong Volcanic Belt is represented with certainty only by the 437 Ma Swatchfield Monzonite. In the northern Rockley – Gulgong Volcanic Belt, the fourth magmatic phase includes volcaniclastics of the upper Sofala Volcanics and Burranah and Tucklan Formations (above isolated Eastonian limestones). Phase 4 magmatism in the Rockley – Gulgong Volcanic Belt concluded (as elsewhere) with sporadic Early Silurian intrusive activity (440–437 Ma) along the entire belt.

IMPLICATIONS FOR MODELS OF ARC EVOLUTION

The integrated biostratigraphic and radiometric age framework for the Macquarie Arc as described above has several key implications.

Four chronologically distinct phases of igneous activity are recognised, that are most clearly delineated in the Junee – Narromine Volcanic Belt and western Molong Volcanic Belt, but generally extend arc-wide. Our correlations support the view (Glen *et al.* 1998) that all three northern belts (and probably also the Kiandra Volcanic Belt) formed in the one arc that was disrupted, first by accretion in the Llandovery and subsequently by rifting in the Silurian and Devonian. At the time of its formation, the Macquarie Arc was significantly wider than indicated by the current geographic configuration of these belts, which reflects shortening along major faults such as the Parkes Thrust, Narambon Fault and Columbine Mountain Fault. Macquarie Arc rocks also underlie the Hill End Trough and parts of the Cowra Trough, and seismic data indicate that similar rocks are present west of the Junee – Narromine Volcanic Belt where they have been overthrust by the Gilmore Fault Zone (Glen *et al.* 2007c).

Hiatuses involving cessation of volcanism, sometimes associated with mild tectonism, separate magmatic Phases 1 and 2, and Phases 2 and 4, in the western half of the arc. Previously published models of arc evolution have sought to explain the causes of these hiatuses, especially that separating Phases 2 and 4 (Glen *et al.* 1998; Glen 2005), and this is discussed in Crawford *et al.* (2007c).

The first hiatus, spanning the late Early Ordovician and half the ensuing Middle Ordovician, is characterised by uplift and erosion, for which there is no clear tectonic cause, although Glen (2005) has speculated on an association with an increase in, or onset of, new backarc spreading in the Wagga Basin.

During the second hiatus of Late Ordovician age, an extensive carbonate platform (early to mid-Eastonian in age) developed in the Junee – Narromine Volcanic Belt and western Molong Volcanic Belt, to be replaced in late Eastonian – early Bolindian time by deeper water fine-grained clastic sedimentation. The eastern Molong Volcanic Belt and Rockley – Gulgong Volcanic Belt lack evidence of this hiatus, as in both areas near-continuous volcanism and volcaniclastic deposition prevailed, with only very minor development of scattered limestones. Our correlations confirm the concept of two separate lithostratigraphic provinces coexisting across the arc in Eastonian time, first proposed by Packham *et al.* (1999) who defined an Eastern Province (characterised by extensive volcanism and volcaniclastic deposition) and a Western Province (dominated by contemporaneous Eastonian carbonate deposition), separated south of Orange by the Narambon Fault that marks a sharp boundary between two regions now only a few tens of kilometres apart with very different Upper Ordovician stratigraphies. We agree that these two provinces were more widely separated (also inferred from restoration of west block reverse movement on the Narambon Fault), although the observed lithological differences might be emphasised by deeper water conditions to the east, thus precluding widespread carbonate formation there. The remarkable similarity between the Billabong Creek Limestone and the Wahringa Limestone Member – Yuranigh Limestone Member – upper Fairbridge Volcanics – Reedy Creek Limestone succession stands in marked contrast to contemporaneous deep-water volcaniclastic-dominated deposits characterising the eastern province. Absence of proven Phase 1 magmatism in the eastern Molong Volcanic Belt and the Rockley – Gulgong Volcanic Belts precludes conjecture as to when the two provinces were initiated. Although Phase 4 shoshonitic magmatism was of arc-wide extent, the provinces were probably not brought into close proximity until major movement on the Narambon Fault and other faults in post-early Llandovery time during accretion of the arc (Glen *et al.* 2007c).

The seamount subduction model of Glen *et al.* (1998) attributed the Eastonian magmatic hiatus to a temporary cessation of subduction induced by collision of a

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buoyant seamount with a west-dipping subduction zone. While the seamount subduction hypothesis cannot be ruled out (Meffre *et al.* 2007), the simple model of a westward propagating wave of uplift and subsidence in the dormant arc requires modification in the light of two lines of subsequent data. First, absence of extensive autochthonous limestones in the eastern Molong Volcanic Belt and the Rockley – Gulgong Volcanic Belt (that should be present in the event of an arc-wide uplift) could alternatively be explained by water depths in the eastern Macquarie Arc being beyond the limits of carbonate deposition. Tectonic separation between western and eastern provinces, discussed previously, may also contribute to the observed distribution of limestones. Second, the Phase 3 magmatism provides support for an alternative scenario, relating the tectonic event that generated the geochemically distinctive Copper Hill Suite to the earlier (Gisbornian) cessation of Phase 2 magmatism and subsequent regional uplift, as proposed by Crawford *et al.* (2007a). Available isotopic dating suggests that Phase 3 magmatism spans the Late Ordovician hiatus in volcanism, with evidence of late Gisbornian unroofing of Phase 3 intrusions in the Molong and Bowan Park areas implying that such rocks were intruded in early Gisbornian time. The observed concentration of Copper Hill Suite intrusives in the Junee – Narromine and western Molong Volcanic Belts may have been initiated by limited east-dipping subduction (Glen *et al.* 2007d; Meffre *et al.* 2007) and corresponds to the presence there (rather than in the eastern part of the arc) of late Darriwilian to Gisbornian limestones.

Packham *et al.* (2003 p. 267) suggested that the Late Ordovician volcanic hiatus in the Molong Volcanic Belt was consistent with erosion of the extinct (Middle Ordovician) Cargo 'volcano' followed by a relative rise in sea-level, with little or no uplift involved. While it is true that in the section at Molong, there appears to be no discernible erosional break at the base of the Reedy Creek Limestone, there is good evidence (both sedimentological and palaeoecological) for deposition of the older Yuranigh Limestone Member on eroded (and presumably uplifted) Fairbridge Volcanics. Exposures of the basal Fossil Hill Limestone in the Cliefden Caves area, and the Ranch Member of the Daylesford Limestone at Bowan Park, also clearly display evidence (well-rounded conglomerates and shoreline deposits) of uplift and erosion of the unconformably underlying volcanics which, in the case of the Walli Volcanics, were erupted as pillow basalts. We argue that concentrations of detrital vein and volcanic quartz and felsic igneous detritus such as noted in the basal Eastonian limestones of the Molong Volcanic Belt by Packham *et al.* (2003) can only be reasonably explained by regional changes in igneous activity (Glen *et al.* 2003) and magmatic fractionation patterns, with significant uplift and erosion prior to carbonate deposition.

Finally, our correlations, in combination with the geochemical work of Crawford *et al.* (2007a), enable identification of major igneous centres throughout the arc. Phase 1 centres are recognised in the Junee – Narromine Volcanic Belt (the Nelungaloo 'volcano', Narromine and Cowal Igneous Complexes) and in the Molong Volcanic Belt (the Mitchell 'volcano'). Phase 2 centres in the Molong Volcanic Belt include the Cargo 'volcano', geochemically similar to the Parkes Volcanics in the Junee – Narromine Volcanic Belt (Crawford *et al.* 2007a) and the separate Walli 'volcano.' The Fairbridge Volcanics and the lower Blayney Volcanics appear to have more diffuse sources: although both formations include pillow lava units and minor coherent volcanics, they are mainly composed of volcaniclastic rocks. Sources of the Triangle and Gidyen Volcanics in the Rockley – Gulgong Volcanic Belt are poorly known. Phase 4 centres occur at Goonumbla in the Junee – Narromine Volcanic Belt (Simpson *et al.* 2005), in the Forest Reefs – Blayney – Byng area (Molong Volcanic Belt) and in the Sofala – Mudgee – Gulgong region (Rockley-Gulgong Volcanic Belt).

The main centres of igneous activity in the Macquarie Arc do not seem to have undergone any major migration with time, although the Phase 4 centres in the southern Molong Volcanic Belt lie further east than centres of Phase 1 and 2 activity. Our correlations show that the magmatic phases extend arc-wide, consistent with geochemical data that point to a major change in magma source and thus to a tectonic event associated with the onset of this second hiatus in volcanism (Crawford *et al.* 2007a).

CONCLUSIONS

Four magmatic phases, generally of arc-wide extent, are defined using the best available biostratigraphic and isotopic framework assembled for the Macquarie Arc throughout the Ordovician and earliest Silurian. The first magmatic phase, Early Ordovician in age, is marked initially in the Molong and Junee – Narromine Volcanic Belts by the development of emergent volcanic islands fringed by shallow-water carbonates, and concludes with subsidence into deep water. A hiatus in volcanism of ~9 million years, from late Bendigonian to early Darriwilian time, is recognised across these belts. Towards the end of this hiatus, some of the Lower Ordovician sequence was uplifted and eroded, and redeposited into sediments associated with the second phase of arc-wide magmatism.

Volcanism during this second magmatic phase was largely submarine, especially in the Rockley – Gulgong Volcanic Belt and southern Molong Volcanic Belt. In the northern Molong Volcanic Belt, locally emergent volcanic centres were flanked by shallow-water carbonate deposition rapidly passing offshore into deep-water environments marked by the accumulation of mass-flow and turbidite deposits.

The geological histories of the Molong and Junee – Narromine Volcanic Belts were closely aligned during the second (Late Ordovician) volcanic hiatus, with this similarity most apparent in the northern and western parts of the Molong Volcanic Belt. Although depositional and volcanic episodes in each area differ in detail, the broad succession of carbonate and subsequent clastic sedimentation is readily correlated across the western province of the Macquarie Arc. A third magmatic phase, the Copper Hill Suite of intrusives, appears

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to be associated with the regional uplift event that led to deposition of the widespread Eastonian carbonate platform in the western part of the arc.

The fourth magmatic phase, characterised by lavas and intrusive porphyries with evolved shoshonitic geochemistry, started immediately after the second hiatus in the Junee–Narromine and western Molong Volcanic Belts in latest Eastonian to early Bolindian time. In the eastern Molong Volcanic Belt and the northern Rockley–Gulgong Volcanic Belt, this fourth phase is represented by a similar significant petrologic and geochemical change, which, in the absence of the Late Ordovician hiatus, is not as precisely constrained in age. Throughout the Macquarie Arc, this volcanic component of the fourth magmatic phase concluded near the end of the Ordovician, although comagmatic monzonites and monzodiorites were not emplaced until the early Llandovery (earliest Silurian).

ACKNOWLEDGEMENTS

This study forms part of a project involving CODES (the Centre for Ore Deposit Studies at the University of Tasmania), the Geological Survey of New South Wales, and selected mineral exploration companies previously and presently active in the Lachlan Orogen. We appreciate company sponsors providing access to newly discovered fossils, and benefited from extensive discussions with our colleagues during the course of the project. Illustrations were drafted by Margaret McLaren, Robyn Sharpe, Phillip Carter, Cheryl Hormann and Trisha Moriarty. We thank Tony Crawford and Dennis Pogson for peer reviews, and are particularly grateful to Gordon Packham and Barry Webby for detailed journal reviews that helped considerably to improve the final version. The paper is published with permission of the Deputy Director-General, New South Wales Department of Primary Industries–Mineral Resources, and is a contribution to IGCP Project 503: Ordovician Palaeogeography and Palaeoclimate.

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Received 22 April 2003; accepted 27 October 2006

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