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Inception and early evolution of the Ordovician Macquarie Arc of Eastern Gondwana margin: Zircon U-Pb-Hf evidence from the Molong Volcanic Belt, Lachlan Orogen

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Inception and early evolution of the Ordovician Macquarie Arc of Eastern Gondwana margin: Zircon U-Pb-Hf evidence from the Molong Volcanic Belt, Lachlan Orogen

Abstract

The Ordovician Macquarie Arc is faulted against and surrounded by coeval Gondwana-derived quartz turbidites, in the Lachlan Orogen of southeastern Australia. How these juvenile, island arc rocks were emplaced and structurally interleaved with continental margin sequences of eastern Gondwana is an ongoing debate. Understanding the inception of this arc is critical in building an accurate Early Paleozoic reconstruction for eastern Gondwana. However, the arc's inception and early evolution are poorly-constrained due to the lack of reliable geochronological data. Integrated field observations, zircon U-Pb-Hf, mineral and whole rock geochemistry are presented here for the oldest units of the Molong Volcanic belt, which is a central strand of the dissected Macquarie arc. Geochemistry indicates that these are calc-alkaline rocks with high-K (locally shoshonitic) to medium-K affinities. The stratigraphically-lowest mafic volcanic and volcano-sedimentary rocks of the Mitchell Formation have a unimodal earliest Ordovician zircon U-Pb age of 479.8 ± 3.8 Ma, with uniform depleted mantle like initial zircon ϵ_{Hf} values of +12 to +13. The lack of pre-Ordovician zircons and the uniform positive initial ϵ_{Hf} value indicate that the Macquarie arc was initiated in an intra-oceanic setting, far from the influence of eastern Gondwana. The stratigraphically overlying Fairbridge Volcanics includes some more evolved volcano-sedimentary components and has a polymodal, but mostly Ordovician, zircon U-Pb age population, with a youngest component at 444.3 ± 2.4 Ma. Overall, its Ordovician grains show a greater spread in initial ϵ_{Hf} values from +14 down to +8 than the Mitchell Formation, which suggests some continental influence, at least 35 million years after the arc was initiated. The lack of any significant Gondwanan inheritance in Early Ordovician volcanoclastic rocks of the Macquarie Arc along with geochemical comparisons with modern island arcs, suggest that the arc evolved outboard of Gondwana probably as a result of steep, easterly-directed subduction. As the Macquarie arc approached eastern Gondwana there would have been increasing quantities of continent derived sediment subducted beneath the arc which may account for the decreasing and more variable zircon initial ϵ_{Hf} values and the appearance of sparse Precambrian continental detritus at the arc's moribund stage. The collision of the Macquarie island arc with Gondwana initiated the Benambran Orogeny and reflects an important mechanism of episodic continental growth involving the addition of juvenile oceanic crust to a continental margin, which contrasts and alternates with periods of purely accretionary growth.

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Inception and early evolution of the Ordovician Macquarie Arc of Eastern Gondwana margin: Zircon U-Pb-Hf evidence from the Molong Volcanic Belt, Lachlan Orogen

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Key words

Macquarie Arc; Molong Volcanic Belt; eastern Gondwana; arc initiation; zircon U-Pb-Hf; juvenile affinity

1. Introduction

Island arc volcanism contributes significantly to continental growth through Earth history (Draut et al., 2009). Modern active island arcs can be accessed with robust information available on the subduction polarity that can be supported by geophysical techniques, such as seismic topography of Taiwan (Lin et al., 2009), or with the detailed geochemical constraints on the temporal and spatial variation of magma types, such as Fiji and Izu island arcs. (Brandl et al., 2017; Gill and Whelan, 1989). For ancient arcs, the superimposed collisional tectonics required to preserve them in continental domains results in fragmentation, decreasing the potential for preservation of the evidence necessary to establish their initiation, evolution and subduction polarity (Draut and Clift, 2013). Despite that, the geochronology, geochemistry and isotopic analyses on fragments of preserved island arcs can still be accessed, to investigate the crustal growth and the closure of Paleo-oceans, such as within the New England Orogen in eastern Australia (e.g., Manton et al., 2017), Grampian Orogen in Ireland (e.g.,

Draut et al., 2009), and Himalaya Orogen (e.g., Buckman et al., 2018). In this paper we focus on the Ordovician Macquarie Arc within the Lachlan Orogen at the eastern margin of Gondwana (Fig. 1).

The Ordovician Macquarie Arc within the Lachlan Orogen of eastern Australia developed in an intra-oceanic setting but coevally with passive margin of quartz-rich turbidites, cherts and black shales (Adaminaby Group). The Macquarie Arc is juxtaposed against the Adaminaby Group both to the east and west and this relationship has created ongoing debate as to the origin and tectonic setting of the arc (see discussions by Aitchison and Buckman, 2012 and Glen et al., 2009a). At present, the Macquarie Arc is generally acknowledged as an intra-oceanic island arc based on the comparison with adjacent terranes (Glen et al., 2009a), Nd isotope and geochemistry characteristics (Crawford et al., 2007a) and petrology, contact structures and stratigraphy of its main units (Fergusson and Colquhoun, 2018; Meffre et al., 2007). Controversy continues on the arc-continent collision mechanisms (Aitchison and Buckman, 2012; Fergusson, 2003; Fergusson, 2009; Fergusson and Colquhoun, 2018; Glen et al., 2007a; Glen, 2013; Gray and Foster, 2004; Meffre et al., 2007). Key questions addressed here to understand the arc's early history are: (i) when did the Macquarie Arc initiate? (ii) was it initiated with a completely juvenile affinity or did it start with some continental influences but then became more juvenile with time and (iii) how and when did a continental influence appear in the arc, shortly prior to it being extinguished and incorporated into the Gondwana margin?

No earliest Ordovician volcanic ages relating to the arc's inception have been acquired previously via zircon U-Pb dating (Glen et al., 2016), and minimal spatial variations are displayed in the chemistry of its igneous rocks (Crawford et al., 2007a). However, in this study, we report zircon U-Pb ages and Hf isotopic compositions from rare zircons extracted from the Early Ordovician volcanic rocks integrated with mineral and whole rock geochemistry, to better constrain the inception and early evolution of the Macquarie Arc. Understanding the initiation of the Ordovician Macquarie Arc is a missing component to understand crustal growth of eastern Gondwana. The resolution of issues around the Macquarie Arc's early history is relevant to the understanding of accretionary systems and continental growth worldwide.

2. Geological Setting

The Tasmanides of eastern Australia is a collection of orogenic belts, formed along the Palaeo-Pacific margin of Gondwana (Glen, 2005, 2013; Scheibner and Basden, 1998). In south-east Australia, it is divided into three orogenic belts, which from west to east are: the Neoproterozoic-Early Ordovician Delamerian Orogen (Foden et al., 2006), the Cambrian-Devonian Lachlan Orogen (Gray and Foster, 2004) and the Early Paleozoic-Mesozoic New England Orogen (Manton et al., 2017; Fig.1A). The Lachlan Orogen mainly contains rocks ranging in age from lower Ordovician to Carboniferous, with some Cambrian greenstones in central Victoria and on the south-east coast of New South Wales (Crawford and Keays, 1987). Based on major faults, the Lachlan Orogen is divided into three subprovinces (Gray, 1997). The Macquarie Arc in the Eastern Subprovince of the Lachlan orogen is dominated by Ordovician volcanic rocks, which experienced four main deformation events: the Benambran Orogeny (Ordovician-Silurian), the Bindian Orogeny (Silurian-Devonian), the Tabberabberan Orogeny (Early to Middle Devonian) and the Kanimblan Orogeny (Early Carboniferous) (Glen, 2005).

The Macquarie Arc is preserved as three main structural units, with the Junee-Narromine Volcanic Belt in the west passing through Parkes, the Molong Volcanic Belt in the middle near Orange, and the Rockley-Gulgong Volcanic Belt in the east through Gulgong and Oberon (Fig.1B). The Kiandra Volcanic Belt in the south around the Snowy Mountains might be the southern extension of the western Junee-Narromine Belt (Crawford et al., 2007a; Glen, 2005).

Magmatism within the Macquarie Arc is proposed to have commenced before 484 Ma and ceased around 438 Ma, and to have occurred in four phases (Crawford et al., 2007b; Glen et al., 2007b; Percival and Glen, 2007). Phase 1 (~484-474 Ma) volcanism is followed by ~9 million years hiatus, phase 2 (~463-455 Ma) volcanism is terminated by a regional uplift and erosion with a short hiatus, phase 3 (~456-445 Ma) is represented by the Copper Hill Suite (Fig. 2C) in the southern Molong Volcanic Belt and phase 4 (~445-438 Ma) is characterised by the Late Ordovician extrusive and intrusion of shoshonitic porphyries (Crawford et al., 2007a; Crawford et al., 2007c; Glen et al., 2009b). Crawford et al. (2007a) reported initial ϵNd whole rock values of mostly +7.5 to +2.5, which were

calculated using best estimates of the ages of the rocks from the central and western belts. These indicate an increasing continental influence as the arc evolved. This is contrary to detrital zircon age data presented by Glen et al., (2011) who obtained a strong Precambrian Gondwana signature for a sample they assigned to the Mitchell Formation, which is one of the oldest units of the Macquarie Arc. This Gondwanan continental signature is a characteristic feature of the Adaminaby Group but atypical of the Macquarie Arc and at odds with juvenile crustal Nd isotopic signatures of Crawford et al. (2007a). This raises concerns about Glen et al.'s sample, which may not belong to the Mitchell Formation of the Macquarie Arc. Therefore, we targeted the Mitchell Formation and the Fairbridge Volcanics in this study, to further understand the inception and paleo-tectonic setting of the Macquarie Arc.

3. Stratigraphy

The Mitchell Formation and the Fairbridge Volcanics both crop out in the Molong Volcanic Belt, which is approximately 190 km long with mainly volcanoclastic rocks in the eastern part, and a mixture of volcanic and volcanoclastic rocks, limestone with igneous intrusions in the western part (Fig. 2C; Glen et al., 2009b). The Early Ordovician Mitchell Formation is generally regarded as the oldest unit within the Macquarie Arc and it is conformably overlain by the Hensleigh Formation (Fig. 3; Morgan and Scott in Meakin et al., 1999). The Middle to Late Ordovician Cabonne Group, which includes the Fairbridge Volcanics, unconformably overlies the Hensleigh Formation (Fig. 3), which is recognised as a different volcanic phase because of a hiatus marked by limestone occurring between them (Percival et al., 1999; Percival and Glen, 2007). The following unit descriptions are mainly based on our geologic transect (Fig. 2D & Fig. 4) and geological maps and stratigraphy taken from the latest 1:250, 000 scale Dubbo and Bathurst maps (Morgan et al., 1999; Raymond et al., 1998).

3.1. Mitchell Formation

The Mitchell Formation has a minimum thickness of 1,100 m (Glen et al., 2007b), and crops out at 25-30 km south of Wellington (Fig. 1B & Fig. 2D). According to the mapping by Scott et al. (2000), its outcrop is restricted to a ~12.5m section due to generally poor exposure. Our transect through the

Mitchell Formation (Fig. 4) encountered coarse-grained volcanoclastic rocks (including breccias and conglomerates) and fine-grained volcanoclastic sandstone (with thinly bedded layers). Figure 3 shows the stratigraphic relations of the units and relative position of our samples.

3.1.1 Boundary and structure

In the north, the Mitchell Formation is overlain by the Hensleigh Formation (the *Hensleigh Siltstone* in some publications; Morgan and Scott in Meakin et al., 1999, Glen et al., 2007b). In our transect, the contact with the Mitchell Formation is not observed, because it lies in the path of a river and could have been excised by local faulting (Fig. 4 C1, C2), but elsewhere this contact has been described as transitional or conformable (Glen et al., 2007b). In the south, the Mitchell Formation is faulted against the Late Ordovician Oakdale Formation (Morgan and Scott in Meakin et al., 1999, Percival et al., 1999). To the east, the Mitchell Formation is faulted (Neurera Fault) against Silurian-Devonian volcanic and sedimentary rocks. In the Bakers Swamp area, the Mitchell Formation is poorly-exposed and occupies a north-plunging anticline (Glen et al., 2007b; Percival et al., 1999; Scott et al., 2000). Our NNE-SSW transect encompasses the stratigraphically-lowest rocks in the core of this fold through to the upper parts of the Formation. Because of the overall poor exposure, most observations and samples are from road cuttings.

3.1.2 Biostratigraphy and lithology

Sponge *Archaeoscyphia* and graptolites from carbonate breccias in the uppermost Mitchell Formation and the overlying Hensleigh Formation imply that the Mitchell Formation is most likely to be Late Lancelfordian (≤ 482 Ma) to earliest Bendigonian (~ 476 Ma) (Percival et al., 1999; Percival and Glen, 2007). The basal contact of the Mitchell Formation is tectonic, marked by the Neurea Fault (Morgan and Scott in Meakin et al., 1999). The lowest-preserved stratigraphic levels are dominated by poorly-sorted volcanic-derived conglomerate (e.g., 16MF01, Fig. 7A, see also supplementary 2). Most clasts are basaltic to andesitic, ranging in size 0.2-8 cm. Samples of this unit display dislocations in feldspars, undulose extinction in rare quartz and extensive sericitic alteration. Some samples (represented by sample 16MF02) have smaller clasts that are predominantly sub-angular to sub-

rounded basalt and andesite (60-70%), feldspar fragments (10-15%), minor sandstone and chert fragments, with abundant calcite and sericite alteration (Fig. 4B and supplementary 2). Overlying the basal volcanic conglomerates are fine-grained volcanoclastic sandstones (sample 17MF31, Fig. 4A) which are succeeded by volcanic breccia, dominated by a purple andesitic matrix with some gabbro and diorite clasts (sample 17MF30, supplementary 2). The upper portion of the succession is dominated by fine-grained volcanic sandstones (samples 16MF03S, supplementary 2), followed by medium to fine-grained volcanoclastic rocks (sample 16MF03V, supplementary 2) rich in plagioclase feldspar and poor in potassic phases. The overall Mitchell Formation sequence is consistent with descriptions by Glen et al. (2007b), who remarked that this package showed a transition from andesite-basaltic andesite near the base into more trachytic-felsic andesite compositions near the top. Clasts throughout the Mitchell Formation are predominantly basaltic, and their lack of significant rounding or weathering indicates proximal sources in a submarine environment (Morgan and Scott in Meakin et al., 1999).

3.2. Hensleigh Formation (siltstone)

The Hensleigh Formation was separated from the conformably underlying Mitchell Formation, owing to the presence of a richly fossiliferous band with its earliest Bendigonian fauna (Morgan and Scott in Meakin et al., 1999). This formation's thickness varies from ~310 m along Bakers Swamp Creek to 2 km southwest of Warringa homestead (Morgan and Scott in Meakin et al., 1999, Percival et al., 1999, Glen et al., 2007). In our transect, any stratigraphic contacts were obscured by soil or alluvium and we obtained one sample from the Bakers Swamp area (Fig. 4).

3.2.1 Boundaries and structure

The Hensleigh Formation conformably overlies on the Mitchell Formation, and is unconformably overlain by the Fairbridge Volcanics (Percival and Glen, 2007). Glen et al. (2007b) suggest the Hensleigh Formation represents the upper part of a fining-upwards succession of the Mitchell Formation. The Hensleigh Formation is part of the north plunging anticline (Glen and Watkins, 1999).

3.2.2 Biostratigraphy and lithology

Generally, this formation is described as laminated graptolitic siltstone, with minor volcanoclastic sandstone and allochthonous limestone (Glen et al., 2007b). Conodonts and graptolites in this unit indicate an early to middle-late Bendigonian (476-472 Ma) age (Glen et al., 2007b; Zhen et al., 2000).

The basal unit is dominated by thinly bedded siltstone interlayered with volcanoclastic sandstones and contains basalt and andesite clasts (Morgan and Scott in Meakin et al., 1999, Glen et al., 2007b). In the upper part siltstones are locally interbedded with massive to thinly bedded volcanoclastic sandstones, which are poorly sorted and contains a small number of porphyritic volcanic clasts (Morgan and Scott in Meakin et al., 1999). The interbedded fine grained feldspar-rich volcanoclastic sandstones throughout the sequence (16HS01, supplementary 2) indicate that any cessation was probably localised, and during the early Bendigonian, magmatism continued in the Macquarie Arc.

3.3. Fairbridge Volcanics

The true thickness of the Fairbridge Volcanics in the Bakers Swamp area is ~2 km (Crawford et al., 2007a; Percival et al., 1999). Percival et al. (1999) introduced the Fairbridge Volcanics as three sub-belts according to the dispersed occurrences. Further, Crawford et al. (2007a) sub-divided the western sub-belt into six units, according to the lithological changes on the west dipping volcanic sequences. In this study, we focus on the Bakers Swamp area, but a few samples were also taken from the western sub-belt for comparison.

3.3.1 Boundaries and structure

In the Bakers Swamp area, the Fairbridge Volcanics unconformably overlie the Hensleigh Formation with a slight angular discordance, which is confirmed from fossil assemblages indicating a ~9 million years hiatus (Percival et al., 1999; Percival and Glen, 2007). In the western sub-belt, the Fairbridge Volcanics is faulted at its base and conformably overlain by the early Eastonian (~453-450 Ma) Reedy Creek limestone or the Cheesemans Creek Formation, where the former is absent at the south part of the sub-belt (Fig. 2C, Percival et al., 1999). To the north, it is intruded by the Copper Hill

porphyries (~456-442 Ma) around Molong (Fig. 3, and the references therein), and faulted against the structurally overlying the Oakadale Formation, and unconformable with the Late Devonian Cotambal Group to the west (Percival et al., 1999).

3.3.2 Biostratigraphy and lithology

The fossil assemblages indicate this unit is from Early Darriwillian to Late Bolindian (~468-444 Ma) (Morgan et al., in Meakin et al., 1999). The Fairbridge Volcanics are dominated by volcanic breccia, conglomerate and volcanic rocks, which are intercalated with finer volcanoclastic sandstones, volcanic sandstones and minor cherty siltstones (Morgan et al., in Pogson and Watkins, 1998). Autochthonous limestones of the Warringa limestone member and Yuranigh limestone member occur in the Bakers Swamp area and south of Molong in the western sub-belt, respectively (Crawford et al., 2007a; Percival et al., 1999).

In the Bakers Swamp area, the Fairbridge Volcanics was not differentiated but may belong to the upper units based on the north plunging fold structure. It is dominated by feldspar-rich sandstones and siltstones (e.g. 16FV01, supplementary 2), and medium to fine grained gabbro with mafic and felsic xenoliths (e.g. 16FV02, supplementary 2). These juvenile volcanoclastic sandstones indicate a proximal deposition setting, where the interlayering of siltstone indicates breaks of volcanism occurring in a quiet and deep marine environment. The presence of limestone clasts in the upper part of the sequences indicates a source from a shallow marine environment accompanying the volcanic activity (Morgan et al., 1998 in Pogson and Watkins, 1998).

4. Analytical techniques

4.1. Zircon U-Pb analyses

Zircon separations were conducted at the Institute of Hebei Regional Geological Survey, China (see details in supplementary1). Sample 16FV01-1 with very low grain yields were repeated for separation at Australian National University (ANU) (supplementary1). Sample zircons were then handpicked and mounted with standard TEMORA zircon grains, on a glass plate with double-sided tape, which

were then cast with epoxy resin in a mould. The zircon mounts were then polished prior to Cathodoluminescence (CL) imaging. The CL images were obtained via the Scanning Electron Microscope JSM-6490 MonoCL4 at the University of Wollongong (UOW).

Zircon U-Pb dating was on the SHRIMP RG instrument at ANU (except sample 16FV01-1), following the analytical protocols of Williams (1998). The intensity of the primary O^{2-} ion beam was 4–6 nA and spot sizes were ~15–20 μm , with each site rastered for 120 seconds prior to analysis. Six scans through the mass stations were made for each age determination. The raw data were reduced using the application *POXI-SC* developed by ANU, which combines the functionality of the previous two ANU applications *PRAWN* and *LEAD*, which require the now-redundant Mac OS9 operating system and access to the spreadsheet software *Kaleidagraph*TM. Quoted errors on isotopic ratios also take into account non-linear fluctuations in ion count rates above that expected from counting statistics alone (Williams, 1998). This is particularly important for old, damaged, high U+Th zircons, where damage has resulted in post-crystallisation heterogeneity of species on the sub-micron scale, as now revealed by atom probe microscopy (e.g., Peterman et al., 2016). Reliance on counting statistics only in measurements of such targets would result in under-estimation of analytical error. Correction for common Pb was made according to the measured ^{204}Pb and common lead composition for the likely age of the sample from Cumming and Richards (1975). U-Th abundance was calibrated based on the reference zircon SL13 (U = 238 ppm), which was in a set-up mount, and $^{206}Pb/^{238}U$ were corrected using the TEMORA standard with a concordant age at 417 Ma (Black et al., 2003). The reduced and calibrated data were then assessed and plotted using Isoplot 4.1 (Ludwig, 2003). Calculated mean ages are presented at 95% confidence.

To increase confidence in the calibration of U-Pb, Temora was distributed as several clusters of grains in different parts of the epoxy mounts. These were visited in a random fashion during the analytical session. This allows the best estimate across the mount of the U-Pb calibration and uncertainty with it. If the standard is located in a single cluster some distance on the mount from an unknown, then subtle differences in conductivity of the mount's gold coat and any need to slightly refocus the secondary ion optics for different parts of the mount can lead to systematic error in calibration of U-Pb age for

the unknowns. In no cases were there any such systematic variations detected. This adds greatly to confidence that the U-Pb ages on the unknowns are accurate and can be compared between samples. In order to allow comparison with other data, the weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages in this paper are quoted with the U-Pb calibration error added in quadrature.

Zircon U-Pb dating of sample 16FV01-1 was conducted on laser ablation-inductively coupled plasma mass spectrometry (LA-ICP-MS) with New Wave ESI 193nm laser ablation system and Thermo ELEMET XR high resolution mass spectrometer, at National Research Center of Geoanalysis, China. Single grain analyses included 20s on background gas, 40s on collecting signals from laser ablation on zircons and 20s on purging after ablation. Every ten unknown grains, we had two GJ-1 (609 Ma, Jackson et al., 2004) and one Plesovice (337 Ma, Slama et al. 2008) standard zircons being analysed (supplementary 1). The raw data were processed with Glitter (Version 4.0) (Van Aetherbergh, 2001).

4.2. Zircon Hf isotope analyses

Zircon Hf isotopic ratios were carried out on the RSES ThermoFinnigan Neptune multi-collector ICP-MS with 193 nm excimer laser system at ANU using methods following those of Hiess et al. (2009). The laser pulsed at 5 Hz with energy density of around 10 J/cm², and the beam 42 X 42µm. A gas blank and a suite of reference zircons with varying REE contents (Monastery, Mud Tank, FC1, Plesovice and QGNG) were analysed after every 10-15 unknown sample spots throughout the session as quality control monitors.

As the zircons were obtained from volcanoclastic rocks, some grains used for U-Pb SHRIMP analyses were too small (less than 40 µm) to be analysed by LA-ICP-MS. Therefore, some Hf analyses are on different, larger, zircon grains from the same sample, but without SHRIMP U-Pb analyses. More detailed method description, complete sample Lu-Hf isotopic data and reference zircon analyses are presented in supplementary data 1.

4.3. Whole rock geochemistry

Fifteen representative samples (supplementary 1) were crushed using a chromium steel TEMA ring mill for geochemical analyses. The major elements were determined by X-ray fluorescence (Spectro-

Analytical Instrument XEPOS energy dispersive polarisation XRF spectrometer) at UOW, with uncertainties less than 3%. 12% tetraborate to 22% metaborate flux mixture was used for samples dominated by mafic compositions, whereas 57% to 43% flux mixture was used for the felsic samples, as following the instructions of Norrish and Chappell (1977).

The trace elements (including the Rare earth elements (REEs)) were analysed by Inductively Coupled Plasma-Mass Spectrometry (ICP-MS) in ALS Mineral Division, Brisbane, having a precision expectation of $\pm 10\%$. Samples 16FV01Sl, 16FV01, 16FV01Sa, 16MF01, 16MF01c and 16MF01c2 were analysed with the ME-MS81 procedure, which is a 30-element package by lithium borate fusion and ICP-MS. Samples 16MF03S, 16MF03V, 16MF02, 16FV02ii, 16FV03, 16FV02, 16FV02i, 16HS01 and 16CC01 were analysed with the ME-MS61r procedure, which is a 48-element package by four acid, ICP-AES and ICP-MS. In both methods, the prepared samples were added to lithium flux for melting and then cooled and dissolved in an acid mixture containing nitric, hydrochloric and hydrofluoric acids. But in the ME-MS61r method, the REEs may not be totally soluble. Further details of standards for monitoring drift can be obtained from the ALS laboratories.

4.4. Mineral chemistry

Polished thin sections for petrographic analyses were prepared at UOW. Electron Microprobe analyses of pyroxenes and plagioclases were carried out by using JXA-8230 at the Institute of Mineral Resources, Chinese Academy of Geological Sciences. The analyses were performed under 20 kV accelerating voltage and 20 nA beam current with a 5 μm diameter analysis site (Su et al., 2017). The raw data were corrected with ZAF standard processing protocols after (Armstrong, 1995), with precision ranging from 1% to 5% (Su et al., 2017).

5. Zircon geochronology and Lu-Hf isotopic compositions

5.1. Zircon morphology and U-Pb geochronology

Representative samples from the Mitchell Formation and Fairbridge Volcanics were chosen for geochronology. Cathodoluminescence (CL) images of zircon grains are given in Fig. 5, the U-Pb data are given in supplementary 1, and the data are plotted in diagrams in Fig. 5.

5.1.1. Mitchell Formation

The Mitchell Formation is characterised by low yields of zircons from the volcanoclastic rock samples which reflects the predominantly basaltic nature of the source rocks. Ten zircons obtained from a volcanoclastic rock sample (16MF02) were used for SHRIMP U-Pb dating. Both prismatic (smaller) and abraded zircons (larger) are present, with crystal size ranging from 50-70µm by 60-100µm. Some grains have opaque inclusions, while others are colourless. Th, U and Th/U values are 35-241ppm, 126-481ppm and 0.24-0.50, respectively (supplementary 1). All grains have low levels of common Pb and yield concordant ages indistinguishable from each other. The weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of all the ten grains is 479.8 ± 3.8 Ma ($n = 10$, MSWD = 0.95), which is interpreted as the age of the volcanic component in the Mitchell Formation.

Another mafic volcanoclastic rock (16MF01) yielded only three zircons with ages 394 Ma, 1003 Ma and 3094 Ma. Given the limited zircons from this sample and their disparate ages, with an apparent, Early mid-Devonian age of one grain and no Ordovician age, we consider these grains probably are laboratory contamination.

Felsic volcanoclastic rock (16MF03S) yielded 20 zircons, but only a few were fully analysed with a $^{206}\text{Pb}/^{238}\text{U}$ age at 427.9 Ma ($n = 3$, MSWD = 0.64). Other analyses were aborted, when accumulating data indicated Permian or younger ages. Considering this sample is of an Ordovician rock of the Macquarie Arc, grains as young as Permian must be laboratory contamination. Other grains examined in reconnaissance mode yielded ages of 61-67 Ma. Therefore, we have no confidence that any of these zircons 'belong' to the sample. The zircon separation of another felsic volcanoclastic rock

sample (16MF03V) yielded only two grains with ages of 683 and 1821 Ma and reconnaissance analyses on other grains yielded ages of < 350 Ma. Again, there is no confidence that any of the separated grains belong to the sample. This highlights a potential difficulty of dating basaltic-andesitic rocks, where the yield of zircons is always going to be low. Therefore, we highlight the importance of geological context and biostratigraphic controls in assessing and calculating the accurate ages where zircon yields are very low.

5.1.2. Fairbridge Volcanics

Two volcanoclastic rock samples (16FV01Sa and 16FV01) from the Fairbridge Volcanics were used for zircon dating. In 16FV01Sa, 12 zircons were analysed, and most are small (50-70 μm) and clean, with a few big dark rounded grains of ~ 100 μm . In CL images (Fig. 5), the majority of the zircons displayed long prismatic to equant euhedral shapes, and have intense oscillatory or band zoning; others (6.1, 7.1, 9.1, 12.1) are smaller in size and having no clear internal structures. The U and Th contents and Th/U ratio are 81-3558 ppm, 23-3394 ppm and 0.22-1.18, respectively. The youngest zircon age at 435.1 ± 10.9 (1 σ) Ma was obtained from spot 11.1. A second analysis of this grain (11.2) yielded an age of 399.1 ± 10.1 (1 σ) Ma. We consider this grain has suffered a small degree of ancient Pb loss and therefore 435.1 ± 10.9 Ma more closely reflects the crystallization age of this grain. Given the high U content (2796-3558 ppm) in 6.1, 7.1, 9.1 and 12.1, and the possibility of a matrix contrast with the low-U Temora standard causing systematic error in U-Pb calibration (e.g., Magee Jr et al., 2017), $^{207}\text{Pb}/^{206}\text{Pb}$ ages were used for weighted mean age calculation of these grains, together with $^{206}\text{Pb}/^{238}\text{U}$ ages from the other grains. Based on this approach, a weighted mean age is 454.2 ± 6.4 Ma (95% confidence, MSWD = 1.00, Fig. 5B1). This is marginally younger than, but within error of the previously presented Darriwilian age (~ 467 -459 Ma, Fig. 3) for part of the Fairbridge Volcanics (Glen et al., 2011).

In sample 16FV01, most of the extracted zircons are round, clear and small to medium in size (70-30 μm by 100-50 μm), and a few of them are yellow in colour with larger sizes. In CL images, well preserved entire zircons show concentric oscillatory zoning and a few of them show recrystallisation

patches. The larger zircons are grain fragments. Four small grains show intense oscillatory zoning (3.1, 7.1, 11.1 and 13.1), some larger sizes of fragments show wider bands (2.1, 4.1, 5.1, 5.1A and 9.1), and the remaining have no zoning features. The Th, U content and Th/U ratio (except grain 8.1) are 27-703 ppm, 109-969 ppm and 0.24-0.73, respectively. Zircon-8.1 has high Th =2006 ppm, U =4145 ppm and Th/U ratio 0.48. Owing to the high U+Th content, the $^{207}\text{Pb}/^{206}\text{Pb}$ age of this grain was used for weighted mean age assessment, while other grains were calculated with the $^{206}\text{Pb}/^{238}\text{U}$ ages. Grain 7.1 shows a Mesoproterozoic age (1401 Ma), and grain 1.1A with the apparently youngest age (430 Ma) appears recrystallised in images, and therefore might have suffered some loss of radiogenic Pb, while the other 10 grains form a non-Gaussian distribution in which two age components of 440.1 ± 5.3 Ma and 469.4 ± 4.7 Ma (2σ) were resolved by mixture modelling in Isoplot (Fig. 5C1).

Sample 16FV01-1 is from the same locality as 16FV01, with the same rock type and similar zircon morphology (Fig. 5D), but it was analysed with different method of LA-ICPMS. Twenty-one grains were analysed, and it displayed a weighted mean age of 442.7 ± 3.2 Ma (Fig. 5 D1, see also supplementary 1).

5.2. Hf isotopic compositions

Most of the Lu-Hf isotope analyses were performed on the same zircon grains where U-Pb data were obtained. As some of the SHRIMP dated zircons are too small to be analysed with LA-ICP-MS, some additional analyses were undertaken on other bigger grains without U-Pb ages (Fig. 6, supplementary 1).

5.2.1 Mitchell Formation

Zircon grains from the juvenile volcanoclastic rock sample (16MF02) are very small, thus only three analyses were obtained from two grains (16MF02-9.1 (one analysis) and 16MF02-10.1 (two analyses)) with $^{206}\text{Pb}/^{238}\text{U}$ ages at 482.4 Ma and 489.0 Ma. The initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios are 0.28282, 0.28287 and 0.28286, initial ϵHf (t) values of +12.0, +13.8 and +13.6 based on the corresponding $^{206}\text{Pb}/^{238}\text{U}$ ages. The ϵHf single stage model ages calculated with respect to a depleted mantle model (supplementary 1)

T(DM) are 604 Ma, 538 Ma and 546 Ma. The highly positive initial ϵ_{Hf} values indicate that the Mitchell Formation volcanoclastic rocks were sourced from entirely juvenile components.

A negative ϵ_{Hf} (t) value = -17.8 was obtained on a zircon with an age of 425 Ma from felsic volcanoclastic rock of 16MF03S. The significance of this younger age coupled with the negative the ϵ_{Hf} (t) value is explored in the discussion section of the paper.

5.2.2 Fairbridge Volcanics

Fifteen zircons from a volcanoclastic rock sample (16FV01Sa) were chosen for Lu-Hf analysis. Because of the small size of some zircons dated by SHRIMP, only 6 Lu-Hf analyses were performed on grains with U-Pb ages, while the other analysed 9 zircons are from the same sample but no direct ages. As the SHRIMP U-Pb zircon data from this sample display a concordant age at 454.2 Ma (Fig. 5B1), the weighted mean age can be used for ϵ_{Hf} (t) calculation (except zircons of 16FV01Sa-E and L). The results show initial $^{176}\text{Hf}/^{177}\text{Hf}$ values between 0.28235 and 0.28291, and initial ϵ_{Hf} (t) values between +8.1 and +14.7. The corresponding ϵ_{Hf} depleted mantle model ages T(DM) range from 462 Ma to 747 Ma (except for 16FV01Sa-E and L). These two grains of 16FV01Sa-E and L have no SHRIMP U-Pb zircon ages, and T(DM) ages are 1087 Ma and 1238 Ma. These older model ages suggest Late Mesoproterozoic juvenile crust formation and sourcing of these grains from Gondwana. However, we cannot preclude the possibility of laboratory contamination on these two grains.

6. Mineral chemistry

Electron microprobe data (supplementary 1) were obtained on the plagioclase and pyroxene from a volcanoclastic rock of the Mitchell Formation (16MF01; Fig. 7A). Most of the analysed plagioclases are from the lithic clasts (except mf01-18, which is hard to identify as lithic clast or matrix; Fig. 7D). Plagioclases from clasts mf01-8 and mf01-18 are moderately altered to chlorite and sericite, and those from mf01-12 and -17 are partly pseudomorphed by undulose quartz. The pyroxenes are less altered with a few of them having restricted chlorite alteration (Fig. 7B, C). Figure 7B shows the clinopyroxene and plagioclase are in one volcanic clast, and Figure 7C displays the clinopyroxene with alteration zone in the matrix.

6.1. Plagioclase

Eight plagioclase fragments and crystals were analysed. In Figure 7D, the feldspars with chlorite and sericite alteration from clast mf01-8 show less sodium content compared with others, falling in the anorthoclase and oligoclase field; the feldspars with sericite alteration from mf01-18 are albitic, and those with silica replacement are albitic except one oligoclase. Generally, the feldspars are in the range of albite-oligoclase to approaching anorthoclase in composition. Coupled with the chlorite and sericite development, this shows widespread alteration of protolith igneous plagioclase.

6.2. Pyroxenes

Twenty pyroxenes were analysed. In Figure 7E, it shows 14 pyroxenes plot in diopside field, 3 are marginal between augite and diopside, and 2 of them are apparently anomalously Ca rich. Relative to the clinopyroxene-orthopyroxene thermometer, they indicate equilibration between 800-700°C. In Figure 7F, the Al(iv)-Al(vi) diagram shows more than half of the pyroxenes were formed at mid-pressure, and small mounts (6 out of 20) apparently at high pressure. The tectonic setting discrimination diagrams (Fig. 7G-H) indicate the pyroxenes crystallised in an island arc setting with high Al and calc-alkaline affinities (Fig. 7I).

7. Whole rock geochemistry

7.1. Alteration

Even when the most homogeneous samples free of modern weathering have been selected for analysis, it is essential to assess the geochemical data to check the degree of ancient weathering. The M (mafic)-F (felsic)-W (weathering) algorithm of Ohta and Arai (2007) was applied to assess the degree of ancient alteration. As shown in Fig. 8A, three felsic volcanic rock samples (16MF03S, 16MF03V and 16MF01c2) from the Mitchell Formation and 1 sample (16HS01) from the Hensleigh Formation are displaced from the trend for fresh calc-alkaline rocks, indicating they have been slightly to moderately modified by ancient weathering. On the other hand, the mafic volcanoclastic rocks from the Mitchell Formation and the Fairbridge Volcanics are relatively unaltered.

To avoid the alteration and metamorphism factors, “immobile” elements, such as the HFSE and REE which are least sensitive to hydrothermal alteration and weathering are used to evaluate petrological features and tectonic settings (MacLean, 1990). The “mobile” elements such as LILE are only applied for samples regarded as fresh from field and petrographic evidence. Additionally, because the Cr-steel mill was used for crushing samples, V, Cr, Co and Ni abundances may be compromised, and are not used in interpretations.

7.2. Major and trace elements

7.2.1. Mitchell Formation and Hensleigh Formation

The volcanic and volcanoclastic rocks from the Mitchell Formation mainly fall in the andesite or basalt field (Fig. 8B), and display calc-alkaline features with medium to high potassium or even shoshonite affinity (Fig. 8C, D). The three felsic volcanic rocks (16MF01c2, 16MF03S and 16MF03V) fall in the rhyolite or dacite section. Generally, the volcanic rocks from the Mitchell Formation have moderate geochemical composition variation. Sample 16HS01 from the Hensleigh Formation shows relatively higher content in SiO₂ (74.88 wt.%), but lower Mg# (28.55), compared with the Mitchell Formation (supplementary 1).

The volcanic and volcanoclastic rocks from the Mitchell and Hensleigh formations show similar chondrite-normalised REE patterns with LREE enrichment and (La/Sm)_N ranging from 1.99 to 3.11 (supplementary 1). Samples from the Mitchell Formation have higher ΣREE content and slightly higher fractionation compared with the Hensleigh Formation (Fig. 8E).

In the primitive mantle normalized spider diagram (Fig. 8F), the volcanic and volcanoclastic rocks from the Mitchell and Hensleigh Formations all show LILE enrichment such as Ba, K, Pb and Sr. They have negative Th, Ti and P anomalies, and strongly negative anomalies for the HFS elements Nb, Ta, Zr, and Hf, especially for samples of 16MF02, 16MF01c, 16MF01 and 16HS01. Felsic samples 16MF03S, 16MF03V and 16MF01c do not show obvious Zr and Hf anomalies.

7.2.2. Fairbridge Volcanics

Seven samples from the Fairbridge Volcanics, include two volcanoclastic rocks (16FV01, 16FV01Sa), 1 gabbro (16FV02) and two diorite xenoliths (16FV02i, 16FV02ii), 1 andesite clast (16FV03) and 1 siltstone (16FV01sl). These samples display the same rock type of andesite/basalt (Fig. 8B) and high-K calc-alkaline features (Fig. 8C, D), as the Mitchell Formation. Generally, samples from the Fairbridge Volcanics have limited geochemical composition variation, except that the siltstone (16FV01sl) has higher SiO₂ (92.13 wt.%), and lower Mg# (3.0) than the volcanoclastic rock samples (e.g. 16FV01; see supplementary1).

The Fairbridge Volcanics display a slight enrichment of LREE with (La/Sm)_N varying from 1.87 to 2.29 (supplementary 1), except one of the xenoliths at 1.40. These volcanic rocks exhibit lower ΣREE content and lower LREE/HREE fractionation compared with the Mitchell Formation (Fig. 8E). The primitive mantle normalized spider diagrams of the Fairbridge Volcanics show LILE enrichments such as Ba, K, Pb and Sr, and have U positive anomaly (Fig. 8F). They also have negative anomalies for the HFS such as Th, Nb, Ta and Ti. Two samples of diorite xenoliths have similar trace element patterns, but stronger Nb, Ta, Zr and Hf negative anomalies (Fig. 8F).

8. Discussion

8.1. Assessing data from samples with small yields of small zircons

The basaltic-andesitic protoliths that are the focus of this paper are silica-undersaturated to low degree of silica saturation rocks, where the growth of magmatic zircon is inhibited, with most of the typically ~ 50 ppm zirconium budget of such rocks mostly incorporated in other phases, such as clinopyroxene. However, there is still the possibility of rare zircon crystallisation, such as in residual melt patches, where localised fractionation leads to silica saturation and zircon crystallisation. However, these zircons will be scarce, particularly in rapidly-cooled volcanic rocks.

For some samples there was an apparent problem in the mineral separation resulting in some laboratory contaminants to the zircon population for the first batch of samples process. In some low

yield samples, we encountered zircons of Carboniferous or even much younger age. These are clearly contaminants. In cases where zircon data was dominated by young and particularly old zircon ages and few or no Ordovician ages were present, then all data were rejected, even if one or two grains yielded ‘sensible’ ages. In this way, the data we present on the few zircons in only some samples can be considered robust.

8.2. Contradictory U-Pb age datasets

8.2.1. Mitchell Formation

The Mitchell Formation is dominated by basic to intermediate volcanic and volcanoclastic rocks with little or no detrital quartz. With such provenance, samples of the Mitchell Formation should have very low yield of (small) volcanogenic zircons. This was found to be the case in our sample suite. Samples (16MF01, 16MF03S and 16MF03V) yielded ≤ 20 zircons each, with many grains too small or altered to be dated with confidence. Of the few grains dated in these samples, some of them were too young (< 400 Ma), to belong to the Mitchell Formation, that is, younger than the conformably overlying Hensleigh Formation, and some clearly do not belong to the collected sample (< 100 Ma) but are laboratory contaminants. Given the presence of such contaminant grains, a few grains that give ‘believable’ Silurian (428 Ma, $n = 3$) and Precambrian (500-700 Ma & >1000 Ma) ages cannot be relied upon, and therefore we have decided not to use any of these data.

On the other hand, we regard our sample 16MF02 that contains a small, high Th/U zircon population with a unimodal age of 479.8 ± 3.8 Ma ($n = 10$), as a robust accurate and precise age for this part of the Mitchell Formation. The unimodal age and the small to modest yield of zircons are consistent with this unit being regarded as the stratigraphically lowest part of the Macquarie arc (Fig. 3). This is a marked contrast with the Mitchell Formation sample reported by Glen et al. (2011). Their sample had a large yield of zircons with a complex age spectrum, dominated by Cambrian to Precambrian zircons, i.e. largely Gondwanan-like continental provenance. Besides, the youngest age population out of 73 grains is at ~ 494 Ma (Glen et al., 2011), which is older than the presumed arc initiation time at Early Ordovician from fossil evidence (Zhen et al., 2003). Glen et al.’s Hf isotopic data confirmed this

Gondwanan influence, with a ~494 Ma population having initial ϵHf values from ~zero to -15. Our sample (16MF02) yielded very few zircons with a narrow range of highly positive ϵHf values (+12 to +14), which indicates an entirely juvenile signature, free of Gondwanan continental influence (Fig. 6). Our proposition is that Glen et al. (2011) sample's affinity is with the peri-Gondwanan Adaminaby Group, rather than with the Macquarie arc, as their sample has the same age spectrum as the coeval continental sedimentary rocks of the Adaminaby Group summarized by (Meffre et al., 2007). Thus, this 'Mitchell Formation' sample from Glen et al. (2011) might not belong to the Mitchell Formation of the Macquarie Arc.

8.2.2. *Fairbridge Volcanics*

Given our three samples (16FV01Sa, 16FV01 and 16FV01-1) and one sample from Glen et al. (2011) of the Fairbridge Volcanics are all from the Bakers Swamp area (Fig. 2D), we put all the data together for assessment, which revealed three main age components at around 444 Ma, 459 Ma and 470 Ma and a few Cambrian and Precambrian ages (Fig. 9A, A1). As these samples are volcanisedimentary rocks, the youngest component 444.3 ± 2.4 Ma (2σ) indicates the maximum deposition age, which is stratigraphically younger than 'early Darriwilian to late Gisbonian' (Percival et al., 1999). This age is also slightly younger than (but still within the error) of ages for the Copper Hill intrusions in the southern part of the Molong volcanic belt, intruded into the Fairbridge Volcanics (Fig. 2C). The minimum age of these intrusions is 446 ± 6 Ma (Chivas reported in Perkins et al., 1995a, K-Ar method on hornblende). A possibility to explain this contradiction is the sedimentation process of the Fairbridge Volcanics did not cease with the intrusions, but instead erosional materials from both the intrusions and host volcanics were redeposited.

The dominant age component at 458 Ma may represent the main volcanism of the Fairbridge Volcanics, whereas the ~470 Ma component is recycled material from the Mitchell Formation. A few Cambrian zircons (~507 Ma) with positive ϵHf values (+13.01) are interpreted by Glen et al. (2011) as derived from a Cambrian arc basement. This interpretation is supported by recent studies by Fergusson and Colquhoun (2018). The appearance of one Precambrian zircon indicate a small, older

Gondwana continental component. However, we cannot entirely preclude the possibility of laboratory contamination for this single Precambrian grain.

8.3. Early evolution of the Macquarie Arc-from the most robust zircon U-Pb data integrated with Hf isotopes and geochemistry affinities

8.3.1. U-Pb-Hf

We suggest that the distinct lack of Precambrian zircons in the Early Ordovician volcanic rocks of the Macquarie Arc is telling evidence for its allochthoneity relative to Gondwana. In-situ zircons from the Mitchell Formation to the Fairbridge Volcanics as the arc matured reveal the zircon age spectrum changes from unimodal to multi-modal (Fig. 9). The Mitchell Formation $\epsilon\text{Hf}_{(\text{t-zirc})}$ is positive and indistinguishable from model depleted mantle in the Early Ordovician (Fig. 6), indicating the oldest components of the arc are devoid of any older Gondwanan continental material. The Fairbridge Volcanics also contain juvenile components similar with the depleted mantle, but there is a slight depression in initial $\epsilon\text{Hf}_{(\text{t-zirc})}$ values (Fig. 6) indicating the first involvement of the Gondwana continental component. The evolution away from depleted mantle estimates is also consistent with initial ϵNd values from +7.3 of the Mitchell Formation to +6.8 and +5.6 of the Fairbridge Volcanics (Crawford et al., 2007a). The highly positive ϵHf values (+13.01) of Cambrian zircons (~507 Ma) in the Fairbridge Volcanics obtained by Glen et al. (2011), indicate that juvenile Cambrian arc basement was incorporated into Late Ordovician volcanoclastic rocks of the Macquarie Arc.

The Mitchell Formation is the oldest known unit within the Macquarie Arc, thus the U-Pb-Hf features can best reflect the timing and setting of the arc initiation, which then is placed at shortly prior to 479.8 ± 3.8 Ma, in an entirely intra-oceanic setting, well outboard of the Gondwana continental margin.

8.3.2. Geochemistry

Early Ordovician basal-intermediate volcanic and volcanoclastic rocks of the Molong Volcanic Belt display calc-alkaline to alkaline (high Al and K content) affinities (Fig. 7H, I; Fig. 8 C, D). From the

Mitchell Formation to the Fairbridge Volcanics, the rocks change slightly from the high-K (or shoshonite) to medium-K calc-alkaline (Fig. 8D). The trace element tectonic setting diagrams (Fig. 10) shows a slight trend from the calc-alkaline to tholeiite, and the Σ REE gently decreased (Fig. 8E) with less fractionation.

When we compare the geochemical patterns with arcs globally such as the Andean alkaline, Tonga tholeiite, Fijian shoshonite and Tonga Trench boninite systems (Fig. 11), the Macquarie Arc rocks have almost the same signatures as the Fijian shoshonites (Fig. 8E, F), which might be derived from low-degree partial melting of metasomatized sub-arc lithospheric mantle (Leslie et al., 2009). The slight change from high-K calc-alkaline to tholeiite field for the Macquarie Arc (Fig. 10) fits the most recent proposed temporal variation for modern island arcs (Brandl et al., 2017), such as Fiji and Izu-Bonin-Mariana (Brandl et al., 2017; Gill and Whelan, 1989). This is interpreted to be due to reduced input of unmetasomatized depleted mantle (Gill, 1987). Since the modern analogies of Fiji and Mariana island arcs are both related with Mariana-type high dip angle subduction (Chen and Brudzinski, 2001; Gvirtzman and Stern, 2004), the Macquarie Arc might have been initiated in the same setting. Besides, during the Macquarie Arc's inception (Early Ordovician), mixture components (K-rich calc-alkaline) with both calc-alkaline and alkaline/shoshonitic affinities, also indicate a signature of initially high dip slab subduction as Fiji (Morrison., 1980; Fig. 12). Additionally, with high dip angle subduction, it is easier for subduction direction polarity to flip (Aitchison and Buckman, 2012), following arc-continent collision.

8.3.3. Geodynamic evolution of the Macquarie Arc

The Macquarie Arc is often interpreted as a 'back-arc basin' in order to accommodate assertions that Palaeozoic subduction has always been west-dipping beneath eastern Gondwana (e.g., Gray and Foster, 2004). However, as mentioned by Meffre et al. (2007), in the Japanese back-arc area, the drill-cores show the samples have both eastly-sourced arc and westly-sourced Chinese continent derived sediments (Marsaglia and Ingersoll, 1992). In the west of Macquarie Arc, almost no Ordovician mixture of the arc and continental sediment is reported (Fergusson and Colquhoun, 2018). Therefore,

we suggest instead that the Macquarie Arc's back-arc lay to the *east* of the arc as the passive margin of Gondwana was being subducted eastward (outboard) beneath the Macquarie arc (Fig. 12). In addition, the U-Pb-Hf and geochemistry evolution features from the Early to Middle-Late Ordovician, we propose that the Macquarie Arc was initiated entirely within the Panthalassa Ocean via easterly-directed high dip angle subduction shortly before 479.8 ± 3.8 Ma. The arc was positioned on the overriding plate whilst the eastern Gondwana margin evolved as a passive continental margin (Fig. 12). Subduction of the intervening ocean crust to the east, beneath the Macquarie Arc, eventually drew the arc closer to Gondwana before arc-continent collision results in the Benambran Orogeny at the Ordovician-Silurian boundary, as proposed by Aitchison and Buckman (2012). Steep subduction with a back-arc laying to the east is also consistent with the global statistics (Lallemand et al., 2005), where the steep dips (larger than 58°) are often associated with extensional oceanic retreating upper plates (81%-90% island arcs fit this).

The accretion of the Ordovician Macquarie Arc to eastern Gondwana represents an important mechanism of continental growth characterised by the addition of juvenile oceanic crust that is almost devoid of zircons. Most island arcs are predominantly mafic and zircon-poor meaning they are not always visible in detrital zircon studies. Thus, the juvenile arcs incorporated into continents will be underestimated in U-Pb zircon provenance studies and Hf isotopic modelling of continent evolution that arise from analysis of zircons. Our results show that detailed, targeted studies of Macquarie Arc volcanoclastic rocks yield rare, but important, zircon age populations that help determine the inception and evolution of the arc before accretion with Gondwana. This study on the Macquarie Arc can be applied on other ancient arcs where detailed arc initiation geophysical information is not possible because of later tectonic events, for example, the exotic island-arc in Kamchatka, eastern Russia (Geist et al., 1994) and New England Orogen, eastern Australia (Manton et al., 2017), which would provide important information for understanding the mechanisms of construction of continents.

9. Conclusion

(1) The new unimodal zircon U-Pb age at 479.8 ± 3.8 Ma, from a Mitchell Formation volcanoclastic rock, represents the latest possible initiation age of the Macquarie Arc. The highly positive $\epsilon\text{Hf}_{(t-\text{zirc})}$

(+12 to +13) and the lack of Precambrian “Gondwana-derived” inheritance indicate the Macquarie Arc initiated remote from the influence of any continental crust in the Early Ordovician.

(2) The Late Ordovician Fairbridge Volcanics display a multi-modal age spectrum with the youngest component at 444.3 ± 2.4 Ma, which is younger than the age proposed in previous studies, and with Early Ordovician to Precambrian inherited components. The corresponding $\epsilon\text{Hf}_{(\text{t-zirc})}$ values show an evolution to less positive (+14 to +8), which is consistent with the evolution of many arc systems that gradually cannibalise the older basement material or become influenced by the introduction of continental-derived sediments prior to collision or accretion with a continental margin.

(3) The Macquarie Arc was probably initiated by steep, east-dipping subduction in an intra-oceanic environment within the Panthalassa Ocean, well outboard of the eastern margin of Gondwana.

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Figure Captions

Figure 1. A) Orogenic system of the Tasmanides, eastern Australia (After Glen, 2005, 2013); B) Simplified Geological map of the Macquarie Arc (After Colquhoun et al. 2017). VB = Volcanic Belt, ACT = Australia Capital Territory.

- + Anticline — Unconformity, Top of U towards younger rocks ★ Samples
 + Syncline - - Geological boundary, position approximate — Road
 ▲ Thrust-fault — Geological boundary, position accurate — Rivers
 — Fault ▽ Angular unconformity ● Younging direction

SILURIAN TO DEVONIAN BASINS

- Late Devonian } Sedimentary and volcanic rocks
 Devonian }
 Silurian }

ORDOVICIAN MACQUARIE ARC

- Copper Hill intrusions
 Cheesemans Creek Formation (Ocac)
 Reedy Creek Limestone (Ocad)

Oakadale Formation

- Mafic volcanic sandstone (Ocao)
 Limestone (Ocao_l)

Fairbridge Volcanics

- Volcanic rocks, volcanic sandstone-Undifferentiated (Ocap)
 Unit5-Volcaniclastic sandstone & conglomerate (Ocap_e)
 Unit4-Andesitic lava breccia & conglomerate with pl phenocrysts (Ocap_d)
 Unit3-pillow basalt & breccia with large cpx phenocrysts (Ocap_c)
 Unit2-Course volcaniclastic sandstone with basaltic clasts (Ocap_b)
 Unit1-Volcaniclastic sandstones with antistatic clasts (Ocap_a)

Fairbridge Volcanics-Limestones

- Allochthonous limestone (Ocap_l)
 Warringa Autochthonous limestone lens (Ocap_w)

Hensleigh Formation

- Laminated volcaniclastic rocks and siltstone (Ouhh)

Mitchell Formation

- Volcaniclastic rocks (Ouum)

Figure 2. C) Geological map of the Molong Volcanic Belt. The units are revised after Crawford et al. (2007a); (D) North Molong Volcanic Belt in Bakers Swamp area, after Colquhoun et al. (2017).

Figure 3. Stratigraphic column in the study area. Stratigraphy sequence and fossil ages are from Percival (1999) and Percival et al. (1999). The Copper Hill Intrusion ages: 447 ± 5 Ma (K-Ar method)

is after Chivas in Perkins et al. (1995a), 449.1±1.5 Ma (Ar-Ar method) is after (Perkins et al., 1995b), 450±6 Ma (SHRIMP zircon U-Pb method) is after Geoscience Australia OZCHRON-1 database, sample no. 93844531. The Fairbridge Volcanics zircon ages are from this paper and the Australia stages are after Off et al. (2010) and the references therein. FM = Formation.

Figure 4. Cross section (A'-B' in Figure 3) across the Mitchell Formation and the Fairbridge Volcanics. A) Fine grained volcaniclastic sandstone of the Mitchell Formation, B) Volcaniclastic rock of the Mitchell Formation, C1-2) Contact zones of these three Formations. The dip directions and dip angles were collected from the field and also based on Colquhoun et al. (2017).

Figure 5. CL images of zircons from the Mitchell Formation and the Fairbridge Volcanics with ages (in Ma), and the U-Pb concordia and weighted mean or probability density diagrams. FM = Formation, Vol = Volcanics.

Figure 6. Zircon U-Pb ages vs. initial ϵ_{Hf} values for the volcaniclastic rock samples (16MF02 & 16FV01Sa) from the Mitchell Formation and the Fairbridge Volcanics.

Figure 7. Electron microprobe analyses on sample 16MF01 (A) from the Mitchell Formation; B-C) Thin sections from 16MF01; B) shows clinopyroxenes and plagioclases within a volcanic clast, and C) shows the clinopyroxene in the matrix; D) Albite-Orthoclase-Anorthite diagram for feldspar discrimination; E) Clinopyroxene classification diagram (Morimoto, 1988) and the pyroxene thermometry projection (Lindsley, 1983); F) Al(vi)-Al(iv) diagram for clinopyroxenes (Aoki and Shiba, 1973); G-H) Clinopyroxene for tectonic setting discrimination diagrams. G) is after Beccaluva et al. (1989). In H), outlines of island arc tholeiites and back-arc basin basalt are after Hawkins and Allan (1994) and N-MORB is after Stakes and Franklin (1994); I) is after Ulrych (1986).

Figure 8. A) MFW (Mafic-Felsic-Weathering) diagram (Ohta and Arai, 2007); B) Nb/Y vs $Zr/TiO_2 \cdot 0.0001$ is after Winchester and Floyd (1977); C) FeOT-Na₂O+K₂O-MgO is after Irvine and Baragar (1971); D) MgO vs P₂O₅ is after Crawford et al. (2007a). Only mafic volcaniclastic rock samples were plotted in D); E) Chondrite normalized REE diagram (Sun and McDonough, 1989); F) Primitive Mantle normalized trace element diagram (Sun and McDonough, 1989).

Figure 9. Zircon U-Pb age probability density plots. A) the data (in total n= 106) is from samples 16FV01Sa (n = 12), 16FV01 (n = 12), 16FV01-1 (n = 21) and one sample (n = 61, LA-ICP-MS) from Glen et al. (2011). A1) the ages in this plot are those younger than 485 Ma from dataset A), except the youngest grain at 430 Ma (n=100), using Gaussian deconvolution method for calculation. B) is the probability density plot from the Mitchell Formation.

Figure 10. Tectonic setting classification diagrams for mafic rocks (Wood, 1980).

Figure 11. The comparisons on REE patterns and spider diagrams of the volcanic rocks between the early Macquarie Arc and other subduction zones from different tectonic settings, with the average values. The Andean alkaline (average of sample number: $n=14$) is from Holm et al. (2016), the Tonga tholeiite ($n=9$) is from eastern Lau spreading centre Ewart et al. (1994), the Fijian shoshonite ($n=2$) is from Leslie et al. (2009), and the Tonga Trench bononite ($n=14$) is from Falloon et al. (2008). The Mitchell FM ($n=3$), the Hensleigh FM ($n=1$) and the Fairbridge Vol ($n=6$) are from this paper.

859

860 Figure 12. Cartoon tectonic model for the initiation of the Macquarie Arc. A) The Macquarie Arc is
 861 initiated with an east-dipping subduction at high dip angle, and the magma sources are from different
 862 depths; B) As the arc approached the Gondwana margin, continental sediments involved in the
 863 subducted slab.