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Geochemical and Geochronological Constraints on the Origin and Evolution of Rocks in the Active Woodlark Rift of Papua New Guinea

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Abstract

Tectonically active regions provide important natural laboratories to glean information that is applicable to developing a better understanding of the geologic record. One such area of the World is Papua New Guinea, much of which is situated in an active and transient plate boundary zone. The focus of this PhD research is to develop a better understanding of rocks in the active Woodlark Rift, situated in Papua New Guinea’s southernmost reaches. In this region, rifting and lithospheric rupture is occurring within a former subduction complex where there is a history of continental subduction and (U)HP metamorphism. The lithostratigraphic units exposed in the Woodlark Rift provide an opportunity to better understand the records of plate boundary processes at many scales from micron-sized domains within individual minerals to regional geological relationships.

This thesis is composed of three chapters that are independent of one another but are all related to the overall goal of developing a better understanding of the record of plate boundary processes in the rocks currently exposed in the Woodlark Rift. The first chapter, published in its entirety in Earth and Planetary Science Letters (2011 v. 309, p. 56 – 66), is entitled ‘Lu-Hf garnet geochronology applied to plate boundary zones: Insights from the (U)HP terrane exhumed within the Woodlark Rift’. This chapter focuses on the use of the Lu-Hf isotopic system to date garnets in the Woodlark Rift. Major findings of this study are that some of the rocks in the Woodlark Rift preserve a Lu-Hf garnet isotopic record of initial metamorphism and continental subduction occurring in the Late Mesozoic, whereas
others only preserve a record of tectonic processes related to lithospheric rupture during the initiation of rifting in the Late Cenozoic.

The second chapter is entitled ‘Geochemical and geochronological constraints on the origin of rocks in the active Woodlark Rift of Papua New Guinea: Recognizing the dispersed fragments of an active margin’. This chapter uses a panoply of geochronological (U-Pb zircon) and geochemical (Lu-Hf and Sm-Nd isotopes, trace/REEs, and major elements) tools to investigate the origin the major lithostratigraphic units in the Woodlark Rift. Important findings in this chapter include the establishment of a tectonic link between sialic metamorphic rocks in the Woodlark Rift and the remnants of a Late Cretaceous aged bi-modal volcanic province along Australia’s northern Queensland coast. This link is important because it identifies another rifted fragment of the former Australian continental margin in Gondwana, and demonstrates the complexity of recognizing the dispersed fragments of active margins.

Another important finding of this chapter is that Quaternary aged high silica rhyolites erupted in the western Woodlark Rift have mantle isotopic and geochemical signatures, and are therefore not the extrusive equivalents of partially melted metamorphic rocks found in the lower plates of large metamorphic core complexes. This is important because it signifies that lithospheric rupture has already occurred, despite the fact that mid-ocean ridge basalts are not yet being erupted and there are still topographically prominent metamorphic core complexes in the region. This chapter is not yet published, but is being prepared for submission to Gondwana Research.
The third chapter is entitled ‘Zircon growth in rapidly evolving plate boundary zones: Evidence from the active Woodlark Rift of Papua New Guinea’. The original purpose of this chapter was simply to use U-Pb dating of zircons from felsic and intermediate gneisses in the Woodlark Rift to understand the history of rocks from (U)HP terranes that don’t preserve the (U)HP metamorphic paragenesis. It was soon realized that the types of U-Pb zircon analyses typically performed on a SIMS instrument were going to be insufficient to fully understand the geochemical and geochronological records within zircons from these rocks. Because of this, traditional SIMS analyses for zircons from these rocks are augmented by U-Pb age and elemental depth profiles that elucidate the isotopic and geochemical nature of the sharp boundaries between different aged domains in these polygenetic zircons. The results presented in this chapter demonstrate that zircon U-Pb ages record specific plate boundary events that can be related to the development of the Woodlark Rift, and that traditional assumptions regarding geochemical equilibrium might not hold true in all situations.
GEOCHEMICAL AND GEOCHRONOLOGICAL CONSTRAINTS ON THE ORIGIN AND EVOLUTION OF ROCKS IN THE ACTIVE WOODLARK RIFT OF PAPUA NEW GUINEA

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THESIS

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Lu-Hf garnet geochronology applied to plate boundary zones: Insights from the (U)HP terrane exhumed within the Woodlark Rift

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Abstract

High-pressure and ultra high-pressure (U)HP metamorphic rocks occur in many of the world’s major orogenic belts, suggesting that subduction of continental lithosphere is a geologically important process. Despite the widespread occurrence of these rocks, relatively little is known about the timescales associated with (U)HP metamorphism. This is because most (U)HP terranes are tectonically overprinted and juxtaposed against rocks with a different history. An exception to this are the Late Miocene (U)HP metamorphic rocks found in active metamorphic core complexes (MCC) in the Woodlark Rift of southeastern Papua New Guinea. This region provides a rare opportunity to study the garnet Lu-Hf isotopic record of (U)HP metamorphism in a terrane that is not tectonically overprinted.

In order to constrain the timing of garnet growth relative to the history of (U)HP metamorphism and the evolution of the Woodlark Rift, Lu-Hf ages were determined, in conjunction with measurements of Lu and major element zoning, for garnets from three metamorphic rocks. Garnets from the three samples yielded different ages that, instead of recording the spatial and temporal evolution associated with a single metamorphic event, provide information on the timing of three separate plate boundary events.

The youngest Lu-Hf age determined was 7.1 ± 0.7 Ma for garnets in a Late Miocene coesite eclogite. The age is interpreted to record the time when a garnet-bearing partial melt of the mantle crystallized within subducted continental lithosphere at (U)HP conditions. The young Lu-Hf age from the coesite eclogite is in contrast to a 68 ± 3.6 Ma Lu-Hf age obtained on large (1-2 cm) garnet porphyroblasts, from within the Pleistocene amphibolite facies shear zone carapace bounding exposures of (U)HP rocks in the D’Entrecasteaux Islands. This older age records the growth of garnet in response to continental subduction and
ophiolite obduction in the region north and east of Australia during late Mesozoic/early Cenozoic time.

In order to relate the (U)HP metamorphic rocks to the larger-scale tectonic evolution of the Woodlark Rift, a third Lu-Hf age of 11.2 ± 2.1 Ma was determined for dynamically recrystallized garnet from the southeastern margin of the rift. While the tectonic implications of the Lu-Hf age from the southeastern-rifted margin are not fully known, this result suggests that the eastern and western parts of the Woodlark Rift might have a different metamorphic history.

The garnet Lu-Hf results demonstrate that prior to rifting, the Woodlark Rift underwent a complex history of metamorphism involving garnet growth and recrystallization. The ability of the Lu-Hf system to date Late Miocene garnet growth, and at the same time preserve Latest Cretaceous garnets within Pleistocene shear zones in the Woodlark Rift demonstrate that Lu-Hf garnet geochronology is an essential component of investigations where the primary objective is documenting and understanding the full range of plate boundary processes recorded in metamorphic minerals.
1. Introduction

In the past two decades, Lu-Hf garnet geochronology has been used to gain insight into the history of polymetamorphosed terranes (Duchene et al., 1997; Scherer et al., 2000; Blichert-Toft and Frei, 2001; Lapen et al., 2003; Anczkiewicz et al., 2004, 2007; Berger et al., 2005; Skora et al., 2006; Kylander-Clark et al., 2007; Lagos et al., 2007; Cheng et al., 2008; Corrie et al., 2010; and Zirakparvar et al., 2010). It is not uncommon for a particular region to yield a range of Lu-Hf garnet ages, often older than coexisting garnet Sm-Nd, U-Pb zircon, and \(^{40}\)Ar/\(^{39}\)Ar ages. In some cases, the Lu-Hf garnet age range can be interpreted with respect to the spatial and temporal evolution of the metamorphic terrane thereby providing insight into the evolution of complex plate boundary zones.

This paper presents the first Lu-Hf garnet results on eclogite and garnet amphibolites from the youngest known (U)HP terrane in the Woodlark Rift of southeastern Papua New Guinea (PNG) (Baldwin et al. 2004). Late Miocene eclogites, including an ~8 Ma coesite eclogite (Monteleone et al., 2007; Baldwin et al., 2008), are exposed in the lower-plates of active metamorphic core complexes (MCC) in the D’Entrecasteaux Islands. The young age of high pressure and ultrahigh pressure ((U)HP) rocks in southeastern PNG provides an opportunity to study the entire (U)HP metamorphic path—from subduction to exhumation— in a terrane that has not been re-incorporated into an orogen with subsequent tectonic, metamorphic, or thermal overprinting (Baldwin et al., 2004; Monteleone et al., 2007).

The Lu-Hf garnet isochron ages are combined with garnet Lu and major element concentrations to ascertain whether the ages reflect garnet growth or isotopic closure following growth. This approach has been successfully used in other UHP terranes (Skora et al. 2006;
Anczkiewicz et al. 2007; Cheng et al. 2008; Konrad-Schmolke et al. 2008), and allows for a geologically consistent interpretation. Results for two samples analyzed yield the youngest Lu-Hf garnet ages yet reported in the literature.

Results from the western Woodlark Basin support previous geochronological evidence for Late Cretaceous- Paleocene obduction of the Papuan Ultramafic Belt (Davies, 1980; Lus et al., 2004) in southeastern PNG, as well as Late Miocene (U)HP metamorphism in the D’Entrecasteaux Islands (Monteleone et al., 2007). The garnet Lu-Hf isochron age from the Misima Island MCC records dynamic recrystallization during Middle Miocene amphibolite facies metamorphism on the southern-rifted margin of the Woodlark Rift. These results have implications not only for Cenozoic tectonics of southeastern PNG, but also for the interpretation of Lu-Hf garnet ages and Lu concentration profiles from (U)HP rocks exhumed within complex plate boundary zones.

2. Geology of the Woodlark Rift

Much of PNG is thought to have originated as part of the northern Australian rifted passive margin, having separated from Australia during the Late Mesozoic opening of the Coral Sea Basin and Early Cretaceous breakup of East Gondwana. The present northern limit of Australian continental crust is not known, but remnants of Archean basement occur within the fold and thrust belt on the PNG mainland (Hill and Hall, 2003) and in the Trobriand Basin (Baldwin and Ireland, 1995).

Rates of plate tectonic motion in the vicinity of southeastern PNG are generally high (cm/yr), and are related to microplate formation resulting from oblique convergence between the Pacific and Indo-Australian plates (Johnson and Molnar, 1972; Curtis, 1973; Cooper and Taylor, 1987; Taylor et al., 1995; Wallace et al., 2004; Schellart et al., 2006). The Papuan peninsula is an
active zone of arc continent collision, but in the eastern-most part of PNG, where plate motion becomes extensional, eastward extensions of the Papuan peninsula have stretched, separated, and subsided since at least 6 Ma. This has occurred as the seafloor spreading rift tip propagated westward forming the Woodlark Basin (Taylor et al., 1995).

The western apex of the V-shaped Woodlark Basin is a zone where seafloor spreading transitions into continental rifting (Abers, 1991; Taylor et al., 1995; Little et al., 2007). The D’Entrecasteaux Islands (Fig. 1), in the western Woodlark Basin, contain MCCs with complexly deformed (U)HP to amphibolite facies rocks in their lower plates (Davies and Warren, 1988; Baldwin et al., 2008; Little et al., 2011). Lower plate rocks are separated from upper plate rocks, which consist of unmetamorphosed mafic and ultramafic rocks, by km-scale ductile shear zones and brittle detachment faults (Davies and Warren, 1988; Hill, 1994). The lower plates consist of mafic eclogite (including coesite eclogite; Baldwin et al., 2008) boudins and dikes encapsulated in felsic to intermediate gneiss (Davies and Warren, 1988). U-Pb dating of zircon formed at (U)HP conditions in a coesite eclogite (sample 89321 in this study), indicates that the high-grade basement rocks crystallized at ~8 Ma, but the onset, prograde history, or duration of (U)HP metamorphism is unknown.

The polymetamorphic basement exposed in the lower plates of the D’Entrecasteaux Islands MCCs have been recognized as one of the most recently unroofed MCCs on Earth (Baldwin et al., 1993; Hill, 1994). Zircon U-Pb dating and P-T constraints (Baldwin et al., 2004; Monteleone et al., 2007; Baldwin et al., 2008) show that these rocks were at a depth of greater than 90 km as recently as ~8 Ma. K/Ar, \(^{40}\)Ar/\(^{39}\)Ar, and fission track dating techniques applied to the lower plate rocks have documented an extremely rapid (e.g., \(\geq 100^\circ\) C\(\cdot\)m.y.) cooling history occurring within the past five million years (Baldwin et al., 1993). In the D’Entrecasteaux Islands, seismic
activity (Abers et al., 2002), geomorphology, and $^{40}\text{Ar}/^{39}\text{Ar}$ mineral cooling ages, all suggest that final exhumation of lower-plate rocks to the surface occurred during Plio-Pleistocene to Holocene time and may still be active (Baldwin et al., 1993).

Metamorphic grade decreases steadily eastward along strike of the southern-rifted margin of the Woodlark Rift from (U)HP in the D’Entrecasteaux Islands (Baldwin et al., 2008), to upper amphibolite facies on Misima Island (Appleby et al., 1996), and finally to sub-greenschist facies in the Louisiade Archipelago (e.g., Smith, 1973; Smith et al., 1973; and Smith and Pieters, 1973; Davies and Warren, 1988). Misima Island (Fig. 1), located ~150 km southeast of the D’Entrecasteaux Islands, is roughly bisected by a low angle normal fault (Peters et al., 2004). The western half of the island is the footwall of this fault and contains amphibolite-facies felsic to mafic gneisses intruded by granodiorite plutons. The lower plate is juxtaposed against greenschist-facies schists, unmetamorphosed sedimentary and volcanic rocks, and basalts comprising the upper plate. $^{40}\text{Ar}/^{39}\text{Ar}$ apparent ages from the lower plate of Misima Island indicate cooling through Ar closure between 15.1 and 9.8 Ma for amphibole and 8.0-7.6 Ma for biotite (Baldwin et al., 2008). Biotite $^{40}\text{Ar}/^{39}\text{Ar}$ ages are concordant with U-Pb zircon crystallization ages for granodiorites emplaced during MCC exhumation (Appleby et al., 1996). $^{40}\text{Ar}/^{39}\text{Ar}$ data for lower plate rocks of the Misima Island MCC indicates that cooling occurred 4 to 8 Ma earlier than in the D’Entrecasteaux Islands (8 to 7 Ma; Baldwin et al. 2008). The apparent westward younging of $^{40}\text{Ar}/^{39}\text{Ar}$ ages along the southern margin of the Woodlark Rift suggests that MCCs formed in response to westward propagation of the Woodlark Rift (Baldwin et al., 2008; Little et al., 2007).

The (U)HP to sub-greenschist facies rocks in the Woodlark Rift may have first been metamorphosed during an episode of Eocene arc-continent collision (Davies and Warren, 1988).
The primary evidence for this collision is the Papuan Ultramafic Belt (PUB) on the Papuan peninsula, a succession of Cretaceous oceanic crustal slices emplaced southward onto the leading edge of the rifted Australian margin during the early to middle Eocene (Davies, 1970; Milsom, 1973; Jacques and Chappell, 1980; Lus et al., 2004). Constraints for the Late Paleocene - Early Eocene timing of ophiolite emplacement are based on \(^{40}\text{Ar}/^{39}\text{Ar}\) amphibole crystallization ages from the metamorphic sole of the PUB (Lus et al., 2004). The upper-plate rocks of the D’Entrecasteaux Islands MCCs are geochemically and petrographically similar to the basalts and gabbros of the PUB (Davies and Warren, 1988; Little et al., 2007). Gabbro recovered from the Moresby Seamount during ODP Leg 180 yielded a \(^{206}\text{Pb}/^{238}\text{U}\) age of 66.4 ± 1.5 Ma indicating PUB remnants occur in close proximity to the active seafloor spreading rift tip (Monteleone et al., 2001).

3. Analytical Techniques

Three samples from MCCs in the eastern and western parts of the Woodlark Rift were selected for Lu-Hf garnet geochronology (Fig. 1). These were selected to encompass a range of garnet-bearing lithologies from the D’Entrecasteaux and Misima Island MCCs. They include coesite eclogite from the lower plate and garnet amphibolite from the shear zone carapace of the D’Entrecasteaux Island MCCs, and garnet amphibolite in the lower plate of the Misima Island MCC. Garnets were characterized via LA-ICP-MS and WDS to examine the distribution of select trace and major elements as described below.

For Lu-Hf geochronology, multiple garnet and whole-rock fractions, each consisting of 200-250 mg of material, from each sample were dissolved in the radiogenic isotope clean laboratory at Washington State University. All of the garnet and some of the whole-rock fractions in this study were dissolved using partial dissolution, whereas a few of the whole
rocks were also dissolved using high-pressure bombs. Partial dissolution was intended to
dissolve the bulk of the sample without dissolving zircon, whereas full dissolution was
intended to dissolve the entire sample including zircon (King et al., 2007). The partial
dissolution technique, also known as ‘table-top dissolution’, consisted of dissolving the
samples in a mixture of concentrated HF/HNO₃ (10:1) in closed Savillex™ capsules on hot
plates at 150°C for 24 to 48 hours.

Whole-rock fractions subjected to the full dissolution technique were digested in high-
pressure, steel-jacketed, Teflon vessels (‘Parr style bombs’) at 160°C for five to seven days.
The fluoride solutions from both techniques were converted to chlorides using H₃BO₃ and
HCl (Zirakparvar et al., 2010). A mixed $^{176}$Lu-$^{180}$Hf tracer was then added to the solutions
and allowed to equilibrate on a hot plate for a minimum of 24 hours. Samples were then
dried down, re-dissolved, and centrifuged in preparation for chromatographic separations.
The protocols for dissolution, spiking, chemical separations, and analyses used in garnet Lu-
Hf geochronology are fully discussed in Cheng et al. (2008). The protocol used for
separation of Lu and Hf is based on procedures described by Patchett and Tatsumoto (1980),

Lu-Hf isotopic analyses were performed on a ThermoFinnigan™ Neptune MC-ICP-MS
at Washington State University. Hf and Lu were dissolved in dilute (2%) HNO₃ and
aspirated into the instrument via either an Aridus desolvating nebulizer (garnet Hf) or a
standard Teflon low-flow (50uL/min) nebulizer (Lu and whole-rock Hf) in conjunction with
a dual quartz spray chamber. Analyses were made in static mode and consisted of 75, 8-
second integrations for Hf and 30, 8-second integrations for Lu. During the course of this
study the following isotopic values were measured on the JMC475 Hf standard: $^{176}$Hf/$^{177}$Hf =
0.282150 ± 8, \(^{178}\text{Hf}/^{177}\text{Hf} = 1.467221 ± 46,\) and \(^{180}\text{Hf}/^{177}\text{Hf} = 1.886831 ± 74\) (n = 17).

These values compare favorably with the accepted values for JMC 475: \(^{176}\text{Hf}/^{177}\text{Hf} = 0.282160,\) \(^{178}\text{Hf}/^{177}\text{Hf} = 1.467170,\) and \(^{180}\text{Hf}/^{177}\text{Hf} = 1.886660\) (Vervoort and Blichert-Toft, 1999). Final Hf isotope values were adjusted slightly relative to these values using a linear correction following mass bias and interference corrections.

Natural Yb present in the samples was used to correct for mass fractionation in the Lu measurements following the method, and using the values, reported in Vervoort et al. (2004). Lu and Hf concentrations and \(^{176}\text{Lu}/^{177}\text{Hf}\) ratios were subsequently determined by isotope dilution. All regressions for Lu-Hf isochrons were calculated using Isoplot/Ex (Ludwig, 2003) and a \(^{176}\text{Lu}\) decay constant of 1.867 x 10\(^{-11}\) (Scherer et al., 2001; Soderlund et al., 2004). Present-day CHUR values of \(^{176}\text{Hf}/^{177}\text{Hf} = 0.282785\) and \(^{176}\text{Lu}/^{177}\text{Hf} = 0.0336\) (Bouvier et al., 2008) were used in the calculation of epsilon (\(\varepsilon\)) Hf values.

The \(^{176}\text{Hf}/^{177}\text{Hf}, \(^{176}\text{Lu}/^{177}\text{Hf}\) ratios, Lu and Hf concentrations, and present-day epsilon (\(\varepsilon\)) Hf values for all samples analyzed in this study are reported in Table 1. Uncertainty on the \(^{176}\text{Hf}/^{177}\text{Hf}\) ratio given in Table 1 is the internal precision of the Hf isotope measurement determined by within-run statistics (2\(\sigma\) standard error). For the regression of isochrons and calculation of ages, however, we used uncertainties of 0.5\% for the \(^{176}\text{Lu}/^{177}\text{Hf}\) and, for the \(^{176}\text{Hf}/^{177}\text{Hf}\) ratios, within run errors combined in quadrature with a systematic uncertainty of 0.01\% (1 \(\varepsilon_{\text{Hf}}\) unit). Within-run error is a determination of the precision of the measurements during the course of a single analysis, but does not provide an estimate of the accuracy of the measurement. Accuracy is estimated by repeat measurements of a solution of known isotopic composition such as the JMC 475 standard. Solutions from natural geologic samples, however, in reality are not perfectly elementally pure and often have small but measurable
amounts of interfering species and other elements that present difficult to quantify matrix effects. This is particularly true for garnet samples with very low Hf abundances and high REE contents, such as those analyzed in this study. Using error determined only from within-run statistics or repeated measurement of the JMC475 standard underestimates the total uncertainty in the measurement of Hf poor geologic samples and will result in isochrons with unrealistically high MSWD values.

The distribution of Lu in garnet porphyroblasts was determined using laser-ablation ICP-MS at Washington State University. Garnets were analyzed using a NewWave Nd-YAG laser (λ = 213 nm), operating at ~ 10 J/cm² and 10 Hz, with a 10 -15 µm spot size. Ablated material was carried in purified He gas to the plasma source of a ThermoFinnigan Element2 single collector, magnetic sector mass spectrometer. Only Lu results are presented here. Analyses were conducted in low-resolution mode (m/∆m = 300) using analog mode for ²⁹Si and pulse counting for ¹⁷⁵Lu. Data were screened using Glitter software to remove portions of the analysis biased by inclusions. Detection limits were ≤ 0.25 ppm. Analyses were standardized against glass made from USGS rock standard BCR2-G using ²⁹Si as an internal standard.

Element maps of garnet using WDS (Fe, Mg, Mn, and Ca) were obtained using the Cameca SX-100 electron microprobe operating at 15 kV, at the Rensselaer Polytechnic Institute.

4. Results and Interpretations

4.1 Coesite Eclogite

Sample 89321 (Fig. 1) from Tumabaguna Island, off the coast of NW Fergusson Island, is a coesite eclogite that has been previously described (Davies and Warren, 1988; Monteleone et al.,
2007; Baldwin et al., 2008). The coesite eclogite is present as a mafic boudin in a mylonite gneiss. The mafic boudin has a retrogressed amphibolite facies rind, but the sample examined is a pristine eclogite. The mineral assemblage in this sample is: garnet + omphacite + phengite + rutile + zircon + SiO₂. Garnet porphyroblasts are rounded, mostly inclusion-free, ~0.5 mm in diameter. Previous geochronological constraints for this sample come from a ²³⁸U/²⁰⁶Pb zircon in-situ ion microprobe age of 7.9 ± 1.9 Ma (Monteleone et al., 2007). Ti in zircon, Zr in rutile, and garnet-pyroxene thermometry (Monteleone et al., 2007; Baldwin et al., 2008) yielded crystallization temperatures of 650 – 675°C, 695 – 743°C, and 600 - 760°C respectively.

The X-ray maps (Fig. 2a), obtained for a ~0.5 mm diameter garnet, reveal an off-center oscillatory zoning pattern for Ca and Mg that is partially truncated at the garnet grain boundary. This oscillatory zoning is not observed in the Fe or Mn X-ray maps. The concentration of Lu was measured along transects across four separate garnet crystals, including the one for which X-ray element maps were acquired. The Lu transect across the garnet that was characterized by X-Ray mapping passes through the Ca and Mg oscillatory zone, but no significant change in Lu concentration was observed. Lu also appears homogeneously distributed in the three other garnet grains selected for Lu profiling.

A linear regression (Fig. 2b) using six garnet fractions and two whole-rock analyses from sample 89321 (coesite eclogite) yields an age of 7.1 ± 0.7 Ma (MSWD = 1.1) and an initial ¹⁷⁶Hf/¹⁷⁷Hf of 0.283097 ± 20 ($\varepsilon_{Hf(7.1\text{ Ma})} = 11.2$). An important aspect in interpreting the ~7 Ma Lu-Hf garnet age from the coesite eclogite is to determine the origin of the homogeneous distribution of Lu in garnet. If Lu was initially homogeneously distributed the ~7 Ma Lu-Hf age records garnet growth. Alternatively, if the homogeneous distribution is due to re-equilibration of Lu, the ~7 Ma Lu-Hf age records isotopic closure.
In published examples of eclogite garnets lacking Lu zoning due to thermal re-equilibration, major elements are also homogeneously distributed (e.g. Cheng et al. 2008). This is not the case for the ~7 Ma coesite eclogite, as the garnet we examined via X-ray mapping from this sample contains a pronounced off-center oscillatory zone of high Ca and low Mg. Furthermore, this relict off-center oscillatory zone is truncated at the garnet grain boundary, suggesting that garnets in this sample experienced partial resorption during the reaction to form the current mineral assemblage. If re-equilibration is called upon to explain the homogeneous distribution of Lu in the coesite eclogite garnets, this mechanism must have left Ca and Mg zoning undisturbed. Available thermometry indicates crystallization at temperatures that did not exceed ~760°C therefore, re-equilibration seems unlikely.

An alternative explanation to re-equilibration is that Lu zoning was never present in garnets from the coesite eclogite. This could occur if garnet crystallized as a primary igneous phase in a magmatic system above the Lu blocking temperature. The Ca and Mg zoning in the garnet may have formed initially as garnet crystallized from a partial melt of the mantle. If this garnet-bearing partially crystallized melt was intruded into cold continental crust at UHP conditions, its metamorphic assemblage could have crystallized directly from the partial melt when it crystallized rapidly at UHP conditions. This near-instantaneous crystallization could have essentially frozen the distribution of Lu and major elements in garnet. The ~7 Ma Lu-Hf age from the coesite eclogite is concordant with a previously reported U-Pb zircon age from this sample, interpreted as the time when zircon crystallized at UHP conditions (Monteleone et al. 2007), further suggesting rapid crystallization and isotopic closure.

The coesite eclogite’s mafic bulk composition suggests that this partial melt was basaltic, which would have had a solidus above the closure temperature of the Lu-Hf system. The
emplacement of basaltic melts into subducted continental lithosphere at ~7 Ma is also consistent with the tectonic evolution of the Woodlark Rift (see discussion below). Our preferred interpretation of the ~7 Ma Lu-Hf age from the coesite eclogite is that this age records the time of isotopic closure when this sample crystallized at UHP conditions. The ~7 Ma Lu-Hf age records the end of garnet growth, which corresponds to the time when the (U)HP mineral assemblage formed in this sample.

4.2 Garnet Amphibolite in Shear Zone Carapace

Sample 0620d (Fig. 1) is a medium- to coarse-grained amphibolite gneiss from the shear zone carapace on northern Goodenough Island. It contains large (1–2 cm), partially-retrogressed, garnet porphyroblasts in an amphibole + biotite + quartz + feldspar matrix. The large garnets in this sample are mantled by a matrix of preferentially oriented amphiboles and surrounded by pressure shadows composed of quartz and plagioclase (Little et al. 2011). The rims of these large garnet porphyroblasts are partially retrogressed to amphibole.

Major element X-ray maps (Fig. 3a) of a large garnet porphyroblast from this sample do not show strong zoning. There appears to be a gradation from high Ca, Mg, and Fe, but low Mn, from the rim to the core of the imaged garnet. Of these elements, Mg is the most strongly zoned. Mg is enriched in a thin uneven concentric band within the garnet rim. The concentration of Lu was measured along a rim to rim transect through a ~1.5 cm diameter garnet porphyroblast in this sample, revealing an abrupt increase in Lu concentration in the apparent core region of the garnet.

A linear regression for sample 0620d (Fig. 3b) using five garnet fractions and one Savillex-dissolved, garnet-free whole-rock fraction consisting mostly of amphibole yields an
age of 65.8 ± 6.0 (MSWD = 68) and a poorly defined initial \(^{176}\text{Hf}^{/^{177}\text{Hf}}\) of \(~0.282630 (\varepsilon_{\text{Hf}(65.8\text{Ma})} = -4)\). A separate regression excluding the two garnet fractions with the highest \(^{176}\text{Lu}^{/^{177}\text{Hf}}\) ratios, which plot clearly below the regression for the other four analyses yields a more precise age of 68.0 ± 3.6 (MSWD = 8.4) and an initial \(^{176}\text{Hf}^{/^{177}\text{Hf}}\) of \(~0.282560 (\varepsilon_{\text{Hf}(68.0\text{Ma})} = -6)\). This is our preferred age for this sample and the oldest determined in this study.

The Lu enriched garnet core is consistent with prograde growth zoning, giving rise to our interpretation that the ~68 Ma age records garnet growth during a Late Cretaceous /Early Paleocene metamorphic event. The age discrepancy between these two regressions, however, indicates there is some complexity in the Lu-Hf isotope systematics of these large garnet porphyroblasts (see below). While the isochron ages are concordant, these regressions illustrate the sensitivity of results based on the choice of fractions used for isochron age determination.

The large ~68 Ma garnet porphyroblasts in this sample are hosted within the Pleistocene shear zone carapace (Hill 1994; Baldwin et al., 1993) of the D’Entrecasteaux Islands MCCs. These large garnet porphyroblasts are strongly mantled by a mylonitic fabric, a further indication that these porphyroblasts pre-date development of the shear zone gneiss. The outermost regions of the large garnet porphyroblasts in this sample are assumed to be in textural equilibrium with amphibole in the rock matrix, and major element X-ray maps reveal only very weak compositional gradients throughout the large garnet. These two observations support the possibility that thermal equilibration and partial retrogression occurred in these garnets, probably related to formation of the shear zone carapace for the D’Entrecasteaux Islands MCCs during the late Cenozoic. However, the Lu-Hf isotopic systematics remained undisturbed during mylonitization.
The ~68 Ma Lu-Hf garnet age from this sample indicates that the Lu-Hf isotopic systematics in garnets have largely remained closed since the time of initial garnet growth. It is possible that the large size of the garnet porphyroblasts coupled with the concentration of Lu in their cores, shielded the Lu from thermal equilibration during the late Cenozoic formation of the shear zone. It is somewhat surprising that the two garnet fractions with the highest Lu/Hf ratios have Hf isotopic compositions below the regression for the remaining garnet fractions. The cores of the garnets are enriched in Lu, suggesting that garnet fractions with high Lu/Hf ratios might represent the core that should yield an older age. The slightly younger age and higher MSWD obtained if these two high Lu/Hf fractions are included could indicate complex Lu-Hf behavior in these large garnets. This complexity, however, is not severe enough to significantly impact the isochron age, as both regressions yielded Lu-Hf ages that are within error of each other.

4.3 Garnet amphibolite from the southern-rifted margin

Sample 04148a (Fig. 1) is a fine-grained amphibolite gneiss from western Misima Island on the southern rifted margin of the Woodlark Rift. This rock is composed of garnet + quartz + amphibole + minor clinopyroxene + plagioclase. It contains elongate garnet layers (up to 4 mm in length) comprised of aggregates of sub-millimeter polygonal and inequigranular garnet grains. There are currently no P-T constraints for this sample, but results from elsewhere in the lower plate of the Misima Island MCC indicate upper-amphibolite facies to lower eclogite facies metamorphic conditions (i.e., \( P = 12 - 16.5 \) kbar; \( T = 750 - 810^\circ\text{C} \); Peters, 2007).

Major element maps obtained across one of the garnet layers reveal complex zoning (Fig. 4a). Individual garnet grains making up the elongate garnet layers can be grouped into two
categories based on the distribution of Ca, Mg, Fe, and Mn, but cannot be distinguished from one another optically. The upper-right hand region of the layer shown in Fig. 4a is populated by garnets that have rims with higher Mn than their cores. The lower-left hand region of the layer has garnet grains with the opposite relationship (i.e., cores enriched in Mn relative to their rims). Fe, Mg, and Ca exhibit the inverse behavior to Mn. Between these two chemically distinct regions of the layer, there is a zone where major element variation within individual garnet grains is highly variable.

The elongate garnet layers in this sample are part of a penetrative stretching lineation, and these textures are similar to those that have been documented to record high strain conditions where dynamic recrystallization and diffusion creep in garnet accommodated deformation (Terry and Heidelbach 2004; Storey and Prior 2005). This includes the complex sub-millimeter scale major element zoning in individual garnet grains, the presence of elongate layers that define a stretching lineation in the amphibolite, and the overall arrangement of garnet grains in the layers.

The concentration of Lu was measured across three garnet layers, but we observed no systematic variation in Lu (Fig. 4a). The analytical spot size of the LA-ICP-MS Lu analysis is larger than many of the individual garnet crystals making up the garnet layers in this sample. Because of this, it was not possible to determine if concentration variations exist within individual garnet grains.

A linear regression using four garnet separates, a pyroxene and an amphibole separate, and a bomb-dissolved whole rock fraction (sample 04148a; Fig. 4b) yields a Lu-Hf age of 12.8 ± 3.4 Ma (MSWD = 2.0) with an initial $^{176}\text{Hf} / ^{177}\text{Hf}$ of 0.283080 ± 34 ($\varepsilon_{\text{Hf}(12.8 \text{ Ma})} = $
10.7). A second isochron excluding the bomb dissolved whole rock yields an age of \(11.2 \pm 2.1\) Ma (MSWD = 0.64) with an initial \(^{176}\text{Hf}/^{177}\text{Hf}\) of \(0.283099 \pm 23\) (\(\varepsilon_{\text{Hf}(11.2\text{Ma})} = 11.3\)).

The element maps from the Misima Island garnet amphibolite show that individual garnet crystals making up the elongate layers do not all have the same composition. Garnets for Lu-Hf analysis were picked under plane light, where it is not possible to distinguish between garnet grains with different compositions. It is possible that different age populations were mixed together during the picking process. Another difficulty arises from not knowing whether the concentration of Lu changes within the individual garnet grains.

Mixing between two garnet age populations, however, would be expected to result in considerable scatter on the isochron diagram, and should be recognizable. The \(~11\) Ma Lu-Hf age indicates either the Lu-Hf isotopic system of pre-existing garnets was reset during recrystallization, or any pre-existing garnets do not pre-date recrystallization by more than a few million years and is therefore beyond analytical resolution. The \(~11\) Ma Lu-Hf garnet isochron age most likely records the time when the garnet layers formed in this sample, possibly associated with dynamic recrystallization.

5. Discussion

5.1 Tectonic Implications of Lu-Hf ages

Samples in this study yield different Lu-Hf garnet ages, and it is appropriate to address the tectonic implications of the Lu-Hf ages from the two regions (D’Entrecasteaux Islands and Misima Island) separately.

5.1.1 Western Woodlark Rift (D’Entrecasteaux Islands Region)
We interpret the ~68 Ma Lu-Hf garnet age from the shear zone gneiss (0620d) as dating garnet growth during a Late Cretaceous metamorphic event. The Lu-Hf garnet age for this sample is within error of the upper range (~65 to ~56 Ma) of K-Ar and amphibole $^{40}_{39}$Ar apparent ages reported by Lus et al. (2004) from granulites in the metamorphic sole of the early Mesozoic Papuan Ultramafic Belt (PUB) on mainland PNG, and with ~66.4 Ma zircon from inferred PUB remnants in close proximity to the active seafloor spreading rift tip (Monteleone et al., 2001). The presence of ~68 Ma prograde garnet preserved in the Pleistocene shear zone carapace bounding the Goodenough Island MCC (Baldwin et al., 1993) suggests that this sample may have initially formed during metamorphism associated with ophiolite obduction in southeastern PNG (Davies and Jacques, 1984).

Following ~68 Ma garnet growth, the next event recorded by garnets in the D’Entrecasteaux Islands is crystallization of the coesite eclogite at ~7 Ma. We propose that the coesite eclogite crystallized directly from a basaltic partial melt of the upper mantle, and that the ~7 Ma Lu-Hf garnet age records the time of this event (Fig. 5a). How (U)HP metamorphism at ~7 Ma relates to obduction of the PUB is unknown, but (U)HP metamorphism predates sea floor spreading in the Woodlark Basin.

Seismic observations of the MCCs in the D’Entrecasteaux Islands reveals that crustal thinning in the western Woodlark Rift is compensated by upwelling mantle with anomalously low density (Abers et al., 2002). High elevations in the D’Entrecasteaux Islands are currently supported by buoyant mantle, and the seismic Moho beneath this region is elevated by 10-15 km (Abers et al., 2002). P-wave anomalies in the D’Entrecasteaux Islands also indicate that the sub-continental mantle lithosphere has been replaced by asthenosphere (Abers et al., 2002). The early stages of asthenospheric upwelling and consequent partial
melting appears to have occurred at \(\sim 7 \text{ Ma}\); a record of this is preserved in garnets from the coesite eclogite.

Crystallization of basaltic melts at UHP conditions, at \(\sim 7 \text{ Ma}\) in the D’Entrecasteaux Island region, was subsequently followed by the onset of rapid exhumation from (U)HP depths in this region (Fig. 5b). This is supported by the fact that cooling ages from the MCCs in this region are only a few m.y. younger than the \(\sim 7 \text{ Ma}\) Lu-Hf and U-Pb zircon ages recording crystallization of the coesite eclogite (Baldwin et al., 1993). Little is known about the tectonic evolution of the region between the time of obduction at \(\sim 68 \text{ Ma}\) and crystallization of the coesite eclogite at \(\sim 7 \text{ Ma}\).

5.1.2 Eastern Woodlark Rift (Misima Island Region)

The tectonic context of the \(\sim 11 \text{ Ma}\) Lu-Hf garnet age of the Misima Island amphibolite, is less certain. The \(\sim 11 \text{ Ma}\) garnets strongly resemble dynamically recrystallized garnet found in retrogressed eclogites (Storey and Prior, 2005) from the Glenelg-Attadale Inlier of NW Scotland. According to these authors, garnet in rocks with evidence for dynamic recrystallization can be a mixture of the garnet originally present and garnet that formed in response to deformation. This is because the grain boundaries of pre-existing garnet porphyroblasts, experiencing diffusion-accommodated grain boundary sliding, become progressively recrystallized, while at the same time additional garnets form adjacent to the pre-existing grains. It is not possible to rule out the possibility that the \(\sim 11 \text{ Ma}\) Lu-Hf age from this sample is a mixed age, although the preferred interpretation is that this age records dynamic recrystallization of garnet at \(\sim 11 \text{ Ma}\).
Middle Miocene granitic, granodioritic, and dioritic igneous rocks as well as high-K volcanic rocks occur in the southeastern parts of the Papuan peninsula and on the southern- and northern-riifted margins of the Woodlark Basin (Smith and Davies, 1976; Smith, 1972; Smith, 1973; Ashley and Flood, 1983). The widespread occurrence of middle-Miocene igneous rocks in southeastern PNG has been cited as evidence for active subduction of oceanic lithosphere beneath PNG during middle Miocene time, and the ~11 Ma age could result from middle Miocene-aged subduction related metamorphism in this region (Fig. 5c). It is possible that this middle Miocene metamorphism was not widespread, and not recorded elsewhere in the Woodlark Rift. Alternatively, the ~11 Ma age may record recrystallization in response to ongoing oblique convergence between the Australian and Pacific plates during middle Miocene times, and may not be related to subduction (e.g. Hill and Hall, 2003).

Another alternative for producing the ~11 Ma garnet age is that garnet originated in a manner similar to the formation of the ~7 Ma coesite eclogite. The onset of seafloor spreading in the Woodlark Basin occurred at 6 Ma (Taylor et al., 1995), but presumably crustal thinning and asthenospheric upwelling preceded lithospheric rupture by at least a few m.y. The ~11 Ma Lu-Hf garnet age from the Misima Island sample could thus record the arrival of hot asthenospheric material into the lower crust ahead of the earliest phase of rifting in the eastern reaches of the Woodlark Rift. The protolith of the ~11 Ma garnet amphibolite may have experienced recrystallization as a result of asthenospheric upwelling, or it is possible that the ~11 Ma garnet amphibolite crystallized directly from a partial melt associated with the rising asthenosphere. More work is needed to fully understand the tectonic implications of dynamic recrystallization of garnet on Misima Island at ~11 Ma.
5.1.3 Tectonic Summary

Lu-Hf garnet ages indicate that the metamorphic history of garnet bearing rocks exhumed in the Woodlark Rift dates back to an episode of Late Cretaceous ophiolite obduction. Following Late Cretaceous tectonic collision responsible for producing the ~68 Ma garnets, now preserved in the Pleistocene shear zone carapace of the Goodenough Island MCC, little is known about the tectonic evolution of southeastern PNG. Subduction of oceanic lithosphere beneath the region now occupied by the D’Entrecasteaux and Misima Islands may have ceased until the Miocene. Evidence for this comes from the Middle Miocene subduction-related igneous rocks in PNG, as well as the occurrence of ~11 Ma garnet amphibolite on Misima Island.

Rifting in the Woodlark Basin is currently focused in an area of former continental subduction. Asthenospheric upwelling associated with this rifting has resulted in hot material from the mantle becoming entrapped within cold continental lithosphere where it crystallized at UHP conditions (Fig. 5a). In the D’Entrecasteaux Islands, this apparently occurred at ~7 Ma, only a few m.y. before rapid exhumation. The emplacement of hot asthenospheric material into the former subduction complex does not appear to have resulted in widespread recrystallization and isotopic resetting, as is evidenced by the preservation of the large ~68 Ma garnet porphyroblasts currently found in the shear zone carapace (Fig. 5b).

5.2 Implications for the interpretation of Lu-Hf garnet ages

The pristine nature and young age of the coesite eclogite sample provides a rare opportunity to study the Lu-Hf isotopic record of (U)HP metamorphism in a garnet that has not yet been overprinted. The age from this sample corresponds to the time when this sample crystallized at (U)HP conditions. We did not detect prograde Lu zoning in garnets from this
sample, suggesting that there is no record of initial garnet growth, and only the time when the (U)HP mineral assemblage crystallized is preserved in the Lu-Hf isotopic systematics. In assessing the robustness of our assertion that Lu is homogeneously distributed in garnets from the coesite eclogite we compared our results to those of Skora et al. (2006). These authors determined trace element profiles across the exact cores of eclogite garnets that were prepared by X-ray tomography so as to ensure a perfectly central cut. They documented an extremely narrow central peak for Lu (as well as Yb and Tm) that would have gone unrecognized if the garnets had not been cut across their exact centers. Since X-ray tomography was not used in this study to ensure garnets in the coesite eclogite were cut across their exact centers, we cannot rule out the possibility that an exceptionally small zone of high Lu has gone undocumented in garnets from the ~7 Ma coesite eclogite.

It is important to note, however, that the average Lu concentrations for LA-ICP-MS spot analyses (Fig. 2a) and those determined by isotope dilution (table 1) for the garnet fractions used to construct the isochon diagrams are in good agreement with one another. This agreement is an indication that there is probably not an undocumented region of high Lu in garnets from the coesite eclogite, as this region would be expected to raise the Lu concentration for the isochron fractions. The confidence with which garnet trace element profiles can be interpreted is clearly an important consideration in assessing the meaning of Lu-Hf garnet ages. If our interpretation of the ~7 Ma age is correct, the homogeneous distribution of Lu in garnets from the coesite eclogite indicates that thermal re-equilibration is not the only way to produce a garnet lacking Lu zoning.

Another important consideration for interpreting Lu-Hf garnet ages from (U)HP metamorphic terranes arises out of the relationship between the three samples examined from
the Woodlark Rift. The young age and homogeneous distribution of Lu in garnets from the 
~7 Ma coesite eclogite stand in stark contrast to the presence of Lu enriched cores and older 
age of the large garnet porphyroblasts in the shear zone carapace. These ~68 Ma garnets, 
despite being hosted in a Pleistocene amphibolite gneiss, provide a record of garnet growth 
prior to formation of the Woodlark Rift. What is important about the two contrasting ages 
from the D’Entrecasteaux Islands is that they can be related to specific tectonic events: 
ophiolite obduction at ~68 Ma, and crystallization at (U)HP conditions of basaltic melt at ~7 
Ma.

What is also important about the ~68 Ma and ~7 Ma Lu-Hf garnet ages from the 
D’Entrecasteaux Islands is that they record separate events. When the ~11 Ma Lu-Hf age 
from Misima Island is also considered, it is possible to state that possibly three separate 
metamorphic events are recorded by the Lu-Hf system in garnets from the Woodlark Rift. 
The fact that three different Lu-Hf isochron ages were obtained on garnet bearing 
metamorphic rocks exhumed within this active transient plate boundary zone illustrates the 
utility of the Lu-Hf system for piecing together the history of a tectonically complex region. 
However, it is somewhat disappointing that results cannot be used to constrain the onset or 
duration of (U)HP metamorphism, an objective at the outset of this study.

Results illustrate the potential for preservation of multiple generations of garnet in 
metamorphic rocks exhumed within transient plate boundary zones. It is not uncommon for 
Lu-Hf garnet ages from a particular region to span several tens of m.y. (e.g. Cheng et al., 
2008; Kylander-Clark et al., 2007) to hundreds of m.y. (e.g. Zirakparvar et al., 2010). In 
cases where a large gap in the Lu-Hf ages is not apparent between the different samples from 
a particular region, as in the samples from the Woodlark Rift, it might be easy to mistakenly
attribute the ages as all being related to the same tectonic event. This could potentially lead to the development of erroneous tectonic models, especially in the case of terranes that have been re-incorporated into an orogen with the possibility of subsequent tectonic, metamorphic, and/or thermal overprinting.

6.0 Conclusions

Garnet Lu-Hf ages determined in this study provide valuable insights into the tectonic evolution of garnet bearing rocks exhumed within the Woodlark Rift, as well as the interpretation of garnet ages from geologically complex regions:

1) The oldest garnets dated in the Woodlark Rift yielded a Lu-Hf age of ~68 Ma. We infer these garnets grew during Latest Cretaceous metamorphism that occurred during southward obduction of the Papuan Ultramafic Belt (Davies and Jacques, 1984; Lus et al., 2004).

2) The youngest Lu-Hf garnet age of ~7 Ma from a coesite eclogite records crystallization at UHP conditions. This occurred when a garnet-bearing partial melt intruded subducted continental crust during asthenospheric upwelling in the earliest stages of rifting in the western Woodlark Rift.

3) The ~11 Ma Lu-Hf garnet age from the lower plate of the Misima Island MCC, on the southern rifted margin, records dynamic recrystallization during amphibolite facies metamorphism. There are several possible tectonic interpretations of this ~11 Ma garnet Lu-Hf age.

4) Preservation of Lu-Hf isotopic systematics that record garnet growth at ~68 Ma, despite incorporation into a Pleistocene crustal scale shear zone indicates that the Lu-Hf garnet geochronometer can resist isotopic resetting during intense deformation and
metamorphism. Results indicate that unambiguous tectonic interpretations of garnet Lu-Hf isochron ages from metamorphic rocks exhumed within transient plate boundary zones may be difficult to extract, especially in cases where garnet bearing rocks have been subsequently incorporated into a orogen during tectonism unrelated to the events that formed them.

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We thank Garret Hart and Charles Knaack at Washington State University for help with analytical aspects of this project. This work was supported by National Science Foundation (NSF) grant 0709054 to S. Baldwin, P.G. Fitzgerald, and L.E. Webb. The editorial handlings of T. Mark Harrison, and insightful reviews by Ethan Baxter, John Platt, and two anonymous reviewers have significantly improved this manuscript.
References


Figure Captions

Fig. 1  Geologic map of southeastern Papua New Guinea, showing locations of samples considered in this study. Abbreviations are as follows: G.I. = Goodenough Island, F.I. = Fergusson Island, N.I. = Normanby Island, R.I. = Rossel Island, W.I. = Woodlark Island. Modified after Baldwin et al. (2004).

Fig. 2. a) Photomicrograph and X-ray major element maps of a garnet porphyroblast in sample 89321. Area of X-Ray maps indicated by white box on photomicrograph. Concentration of Lu measured via LA-ICP-MS along transects for four garnet porphyroblasts (including porphyroblast shown here) from this sample. b) Lu-Hf isochron diagram for sample 89321 (coesite eclogite).

Fig. 3. a) Photomicrograph of large garnet from sample 0620d in reflected light, X-ray major element maps for a portion of the large garnet, and concentration of Lu in large garnet measured via LA-ICP-MS along a transect (location of transect indicated by dashed white line on photomicrograph). Area of X-Ray maps indicated by black box on photomicrograph. b) Lu-Hf isochron diagram for sample 0620d. The 65.8 Ma age, which is grayed out, results from regression all of the garnet fractions together, whereas the 68.0 Ma age results from omitting the two garnet fractions with the highest Lu/Hf ratios from the regression.

Fig. 4. a) Photomicrograph of garnet layer in sample 04148a, including X-ray major element maps. The concentration of Lu was measured via LA-ICP-MS along transects across five of the garnet layers; however, the LA-ICP-MS spot size is larger than the scale of the major
element zoning in the individual garnet grains. Area of X-Ray maps indicated by black box on photomicrograph. b) Lu-Hf isochron diagram for sample 04148a. Note that two separate regressions are shown on this diagram. The 12.8 Ma regression is grayed out and was constructed by including all of the garnet fractions together with the whole rock. Excluding the whole rock from the regression yields an age of 11.2 Ma.

Fig. 5. a) Schematic diagram illustrating the proposed mechanism for the origin of the ~7 Ma coesite eclogite in the D’Entrecasteaux Islands. We propose that the coesite eclogite crystallized from an asthenospheric melt that was injected into cold subducted continental lithosphere. This occurred during the early phases of rifting in the region. b) Schematic diagram illustrating the structural context of the ~7 Ma and ~68 Ma garnets in the D’Entrecasteaux Islands. Note that the ~7 Ma garnets are found in the core zone, whereas the ~68 Ma garnets are found in the shear zone carapace. The seismic Moho is currently located ~26-29 km below the D’Entrecasteaux Islands region (Abers et al. 2002). c) Series of schematic tectonic cross-sections illustrating the possible tectonic environment for dynamic recrystallization of garnet on Misima Island. The relationship between garnet growth in the eastern Woodlark Rift to those in the western Woodlark Rift is presently unclear.

Table 1. Analytical results from this study. The uncertainties on the $^{176}\text{Hf}/^{177}\text{Hf}$ values shown in this table are the within-run 2 sigma errors ($\times 10^{-6}$). The within-run error was not used for the regression of isochrons. Instead, an uncertainty was applied to the $^{176}\text{Lu}/^{177}\text{Hf}$ and $^{176}\text{Hf}/^{177}\text{Hf}$ ratios (see discussion under analytical techniques).
Amphibolite - greenschist facies metasediments and metabasalts
- Blueschist facies metabasalts
- Greenschist facies metasediments and metabasalts
- Papuan Ultrama/fic Belt; ophiolitic gabbrow and pillow basalt
- Alluvium; also beach sand
- Raised coral
- Colluvium; some alluvium
- Pliocene - recent volcanic rocks
- Pliocene intrusive igneous rocks
- Eocene intrusive igneous rocks
- Blueschist facies metasediments
- Core zone gneiss; eclogite facies metasediments and metabasalts
- Shear zone gneiss; eclogite facies metasediments and metabasalts
- Undifferentiated marine and clastic sedimentary rocks
- Limits of continental crust
- Spreading center
- Fault
Figure 2a. Photomicrograph of garnet in cross polarized light (XPL), X-Ray maps for Fe, Mg, Ca, and Mn, and Lu concentrations measured via LA-ICP-MS in garnets from sample 89321.

Figure 2b. Lu-Hf isochron diagram for sample 89321.
Figure 3a. Photomicrograph of garnet in reflected light (RL), X-Ray maps for Fe, Mg, Ca, and Mn, and Lu concentrations measured via LA-ICP-MS in garnets for a portion of a large garnet porphyroblast in sample 0620d.

Figure 3b. Lu-Hf isochron diagram for sample 0620d.
Figure 4a. Photomicrograph of garnet in plane polarized light (PPL), X-Ray maps for Fe, Mg, Ca, and Mn, and Lu concentrations measured via LA-ICP-MS in garnets from sample 04148a.

Figure 4b. Lu-Hf isochron diagram for sample 04148a.
Tectonic Interpretation for Lu-Hf Garnet Ages from the D’Entrecasteaux Islands Region (Western Woodlark Rift)

Present day tectonic context of ~7 and ~68 Ma garnets:

- ~7 Ma Lu-Hf age garnet porphyroblast in coesite eclogite
- ~68 Ma Lu-Hf age partially resorbed garnet in mylonite gneiss

Tectonic Context of the ~7 and ~68 Ma garnets during earliest phases of rifting:

- ~11 Ma: Formation of garnet on Misima Island, possibly associated with dynamic recrystallization, as a result of several possible tectonic scenarios
  - ophiolite obduction and continental subduction
  - crustal-scale strike slip motion due to oblique PAC-AUS convergence
  - localized tectonic collision in complex plate boundary zone

- ~68 Ma: ~68 Ma garnet now preserved in SZC

Possible Tectonic Interpretations for Lu-Hf Garnet Age from Misima Island (Eastern Woodlark Rift)

- ~11 Ma: Formation of garnet on Misima Island, possibly associated with dynamic recrystallization, as a result of several possible tectonic scenarios
  - ophiolite obduction and continental subduction
  - crustal-scale strike slip motion due to oblique PAC-AUS convergence
  - localized tectonic collision in complex plate boundary zone

- ~7 Ma: ~7 Ma garnet recording crystallization of basaltic melt at UHP conditions

**No Scale Intended**

Chapter 1; figure 5b

Chapter 1; figure 5a

Chapter 1; figure 5c
<table>
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<tr>
<th>Sample</th>
<th>$^{176}\text{Lu}/^{177}\text{Hf}$</th>
<th>$^{176}\text{Hf}/^{177}\text{Hf}$</th>
<th>$\varepsilon$</th>
<th>Lu ppm</th>
<th>Hf ppm</th>
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Chapter 1: table 1
Geochemical and geochronological constraints on the origin of rocks in the active Woodlark Rift of Papua New Guinea: Recognizing the dispersed fragments of an active margin

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Abstract

The western Woodlark Rift provides a natural laboratory to study the surface expression of lithospheric rupture due to rifting in a once active margin, which in turn can be used to identify rocks in the geologic record that were once situated at the site of lithospheric rupture. The western Woodlark Rift also provides an opportunity to roughly assess the rapidity with which a former subduction complex can become the site of lithospheric rupture due to rifting. New geochemical, isotopic, and geochronological data is presented for metamorphic and magmatic rocks from the southern rifted margin of the Woodlark Basin, Papua New Guinea (PNG) that allows a tectonic link to be established between the metamorphic rocks and the Late Cretaceous Whitsunday Volcanic Province (WVP), northeastern Australia. Felsic and intermediate metamorphic rocks in the Woodlark Basin have Nd isotopic compositions that are similar to the Nd isotopic compositions of rocks in the WVP, and also predominantly contain inherited zircons with U-Pb ages that are bracketed by the timing of magmatism in the WVP. None of the metamorphic rocks displayed the highly evolved Hf or Nd isotopic compositions that would be expected of ancient continental crust despite the proximity of the WVP to exposures of ancient crust. Magmas were erupted in the WVP in response to rifting during the breakup of east Gondwana, and the felsic and intermediate rocks in the Woodlark Basin appear uniformly related to the magmatic products produced during this rifting event. Mafic metamorphic rocks exposed in the Woodlark Rift may have also originated as basaltic lavas crystallized from mantle melts at (U)HP depths and as upper plate basalts incorporated via subduction erosion.
1. Introduction

Active margins are typically an amalgamation of exotic terranes, remnants of ocean basins once separating these terranes, arc magmas, and fragments of continental lithosphere. Once an active margin has rifted apart, there is no single or unequivocal geochemical or geochronological signature that can be used to recognize the individual components of a margin that have been dispersed due to rifting. Documenting the current location of fragments of the eastern Australian margin in Gondwana is important for understanding the development of the Australian passive margin as well as the Cenozoic evolution of PAC-AUS plate boundary. Many of these rifted fragments are now located in the New Guinea region, where they have been tectonically overprinted and metamorphosed due to arc-continent collisions during the late Mesozoic and Cenozoic (e.g. Hall, 2002; Schellart et al. 2006). Lithosphere that was once part of the active eastern Australian margin prior to the breakup of Gondwana may also now comprise many of the submarine plateaus and rises in the southwestern Pacific (e.g. the Lord Howe and Chatham rises, Queensland, Challenger, and Campbell plateaus, and Norfolk Ridge) (Lister and Etheridge, 1989; Veevers et al., 1991; Gaina et al. 1998a, 1998b; Veevers, 2000, 2004; Cluzel et al., 1999; Cluzel et al., 2001; Betts et al., 2002; Schellart et al., 2006; Tulloch et al., 2009; Cluzel et al. 2010a, 2010b).

The development of the eastern Australian passive margin was associated with a ~2500 km long magmatically active rift-system (Bryan et al., 1997). Evidence for this rift system occurs as Late Cretaceous aged bimodal volcanic and igneous rocks sporadically exposed along the present-day northeastern Australian coastline, and as thick successions of volcanogenic sediments in eastern Australian basins. While it has been inferred that a large volume of magma
was produced in this Late Cretaceous rift, these magmas do not appear to have acquired the highly evolved geochemical and isotopic characteristics of Australian continental lithosphere (Ewart et al., 1992; Bryan et al., 1997). Therefore the fragments of sialic crust, rifted away from Australia since the Late Cretaceous and now found dispersed throughout the southwest Pacific, need not necessarily have a geochemical, isotopic, or geochronological signature of highly evolved continental lithosphere tectonically related to the former eastern Australian margin in Gondwana.

We present results of an investigation into the tectonic origin of metamorphic rocks in the Woodlark Rift of southeastern PNG, including: 1) the high-pressure and ultrahigh-pressure ((U)HP) rocks comprising the lower plates of active metamorphic core complexes (MCCs) in the D’Entrecasteaux Islands, 2) amphibolite facies lower-plate rocks of an MCC on Misiima Island, and 3) subgreenschist facies Calvados Schists in the Louisiade Archipelago (Fig. 1). Hf and Nd isotopic, trace/REE and major element, and U-Pb zircon geochronological data is used to establish a tectonostratigraphic link between felsic and intermediate metamorphic rocks in the Woodlark Rift rocks and an early Cretaceous volcano-sedimentary succession along the present-day eastern continental margin of Australia (Bryan et al., 1997). Establishing this link allows us to refine the paleogeographic extent of the northern end of a >2,500 km long rift system that occupied the eastern margin of Gondwana during the Cretaceous (Bryan et al. 1997), and identify another dispersed fragment of crust with an eastern Australian tectonic affinity in the southwestern Pacific.

While the primary objective of this study is to establish the tectonic origin of metamorphic rocks exposed along the entire length of the Woodlark Rift’s southern margin, the geochemical and isotopic data for one particular part of this rift provide implications beyond
simply identifying another remnant of the eastern Australian margin. The D’Entrecasteaux Islands are situated in a zone of active extension at the western end of the Woodlark Rift. Here, a wide variety of metamorphic and magmatic rock types, including the felsic volcanic products of rifting, are exposed in an active tectonic setting. Quaternary felsic lavas are being erupted in close proximity to active metamorphic core complexes (MCCs) exhuming metamorphic rocks that have experienced a full range of plate boundary processes but have not inherited the isotopic and trace element signature of the metamorphosed basement. Instead, the Quaternary felsic lavas have juvenile isotopic compositions and are probably highly fractionated mantle melts (e.g. Smith, 1982). When these observations are combined with pre-existing data for the evolution of the mantle beneath the D’Entrecasteaux Islands, it is possible to gain additional insights into the surface expression of mantle processes in an active tectonic setting. This in turn has implications for the manner in which active margins are preserved in the geologic record once they have been rifted apart.

2. Geology of the Woodlark Rift & Samples Analyzed

Metamorphic rocks are exposed on islands along the southern rifted margin of the Woodlark Rift and west of the active seafloor spreading rift tip (Fig. 1). Metamorphic grade decreases steadily eastward along strike of this margin from (U)HP in the D’Entrecasteaux Islands (Davies and Warrant, 1988; Hill and Baldwin, 1993; Baldwin et al., 2008), to upper amphibolite facies on Misima Island, and finally to sub-greenschist facies in the Louisiade Archipelago (e.g., Smith, 1973; Smith et al., 1973; and Smith and Pieters, 1973). For metamorphic rocks in the region, the predominant bulk compositions by volume is felsic to intermediate, but mafic bulk compositions (e.g. eclogite and amphibolite) frequently occur associated with the felsic and intermediate rocks. A more detailed description of the geology of the rocks in the Woodlark Rift is provided below.
Representative lithostratigraphic units in the Woodlark Rift were sampled as part of this study. The rationale for this strategy and details pertaining to these different lithostratigraphic units are provided in section 2.3.

Prior to this study scarce geochronological information is available for the pre-Late Cenozoic history of these rocks (Baldwin and Ireland, 1995), but it is possible that the felsic and intermediate metamorphic rocks originated as part of the eastern Australian margin in Gondwana. In this scenario, these rocks would have been rifted away from Australia during the Early Cretaceous breakup of East Gondwana and Late Mesozoic opening of the Coral Sea Basin (Davies and Warren, 1988). The data gathered as part of this study allows us to explore this possible connection with Australia. Metamorphism of these rocks could have occurred during the Early Cenozoic obduction of the Papuan Ultramafic Belt (PUB), currently exposed on the Papuan peninsula (Davies, 1980; Milsom, 1973; Jacques, 1980; Davies and Warren, 1988; Lus et al., 2004.). This possibility is supported by the fact that an episode of Late Mesozoic garnet growth (Zirakparvar et al., 2011) is preserved in an MCC bounding shear zone in the western Woodlark Rift (Little et al., 2011).

2.1 D’Entrecasteaux and Misima Islands

The D’Entrecasteaux Islands (Fig. 1), in the western Woodlark Rift, contain metamorphic core complexes (MCCs) with complexly deformed (U)HP to amphibolite facies rocks comprising their lower plates (Davies and Warren, 1988; Baldwin et al., 2008). The western apex of the V-shaped Woodlark Rift is a zone of active extension ahead of seafloor spreading in the Woodlark Rift (Abers, 1991; Taylor et al., 1995; Taylor et al., 1999; Wallace et al., 2004; Little et al., 2007). The upper-plate rocks of the D’Entrecasteaux Islands MCCs, which consist of unmetamorphosed mafic and ultramafic rocks, are geochemically and petrographically similar to
the basalts and gabbros of the PUB (Davies and Warren, 1988; Little et al., 2007). Lower plate rocks are separated from upper plate rocks by km-scale shear zones and brittle detachment faults (Davies and Warren, 1988; Hill, 1994; Little et al., 2011). The lower plates consist of mafic eclogite (including coesite eclogite; Baldwin et al., 2008) boudins and dikes encapsulated in felsic to intermediate gneiss (Davies and Warren, 1988). U-Pb dating of zircon formed at (U)HP conditions in a coesite eclogite from the lower plate indicates that the high-grade basement rocks crystallized at ~8 Ma (Monteleone et al., 2007; Baldwin et al., 2008).

The U-Pb zircon geochronological constraints for the timing of (U)HP metamorphism in the western Woodlark Rift (Monteleone et al., 2007) have been corroborated by a ~7 Ma Lu-Hf garnet age determined for a coesite eclogite (sample 89321 in this study; Zirakparvar et al, 2011). This ~7 Ma Lu-Hf garnet age was interpreted by Zirakparvar et al (2011) as recording the time when a basaltic partial melt of the mantle, associated with rising asthenosphere ahead of rifting in the Woodlark Rift, crystallized at UHP conditions within formerly subducted continental crust. The high elevations of the MCCs in the western Woodlark Rift are supported by a seismic Moho that is elevated by 10 – 15 km, and the sub-continental mantle lithosphere in this region has been replaced by asthenosphere (Abers et al. 2002). The emplacement and crystallization of basalt at UHP conditions within subducted continental lithosphere at ~7 Ma probably occurred in response to this asthenospheric upwelling (Zirakparvar et al, 2011).

The polymetamorphic basement exposed in the lower plates of the D’Entrecasteaux Islands MCCs have been recognized as one of the most recently unroofed MCCs on Earth (Baldwin et al., 1993; Baldwin and Hill, 1993; Hill, 1994). Zircon U-Pb dating and P-T constraints (Baldwin et al., 2004; Monteleone et al., 2007; Baldwin et al., 2008) show that these rocks were situated at a depth of greater than 90 km as recently as ~8 Ma. K/Ar, $^{40}$Ar/$^{39}$Ar, and fission track dating
techniques applied to the lower plate rocks have documented an extremely rapid (e.g., ≥ 100° C/m.y.) cooling history culminating within the past five million years (Baldwin et al., 1993). In the D’Entrecasteaux Islands, seismic activity (Abers, 1991), young geomorphology (Miller et al., 2011), and Plio-Pleistocene ⁴⁰Ar/³⁹Ar mineral cooling ages (Baldwin et al., 1993), all suggest that exhumation of lower-plate rocks occurred during Plio-Pleistocene to Holocene time and may still be on-going.

Misima Island (Fig. 1), located ~150 km southeast of the D’Entrecasteaux Islands, is roughly bisected by a low angle normal fault (Peters et al., 2004; Peters, 2007). The western half of the island is the footwall of this fault and contains amphibolite-facies felsic to mafic gneisses intruded by granodiorite plutons. The lower plate is juxtaposed against greenschist-facies schists, unmetamorphosed sedimentary and volcanic rocks, and basalts comprising the upper plate. ⁴⁰Ar/³⁹Ar apparent ages from the lower plate of Misima Island indicate cooling through Ar closure between ~9 and ~12 Ma, which is ~5 to ~7 Ma earlier than in the D’Entrecasteaux Islands (Baldwin et al., 2008). Similarly to the D’Entrecasteaux Islands, partial melting of the metamorphic basement in the Misima Island lower plate appears to have occurred contemporaneously with MCC formation (Appleby et al., 1996). Lu-Hf dating of dynamically recrystallized garnets in an amphibolite gneiss from the lower plate of the Misima Island MCC by Zirakparvar et al (2011) yielded an age of ~12 Ma, but it is unclear whether this episode of dynamic recrystallization in the lower plate is related to convergence or rifting.

Quaternary volcanism in the D’Entrecasteaux Islands is extremely complex (Smith, 1981; Smith, 1982; Smith and Clarke, 1991; Stolz et al., 1993). The arrival of active seafloor spreading the D’Entrecasteaux Islands has produced basaltic and peralkaline rhyolitic eruptive centers throughout the region. However, calc-alkaline rhyolitic lavas are also actively being
erupted on these islands, and are commonly associated with MCC dome bounding faults (Smith, 1982). The close spatial and temporal relationship between the basalt-peralkaline rhyolite suite and the calc alkaline rhyolites is not well understood, but the geochemical data in this paper allows for an exploration of the relationship between calc-alkaline volcanism and partial melting of metamorphic basement rocks.

2.2 Louisiade Archipelago: Subgreenschist facies

The ~45 islands of the Louisiade Archipelago are located along >160 km of the southeasternmost extent of the southern margin of the Woodlark Rift. Exhumation of the subgreenschist facies Calvados Schist found on these islands may be intricately related to continental rifting prior to the onset of seafloor spreading in the Woodlark Rift (e.g. Webb et al., 2008), but this hypothesis is still in the process of being tested. Currently the only geochronological data from the Louisiade Archipelago is a K/Ar whole rock age of ~11 Ma on a pyroxene andesite from the western part of the island chain (Smith, 1973). Geologic investigations in the 1970’s (e.g. Smith, 1973; Smith et al., 1973; and Smith and Pieters, 1973) were conducted by the Australian Department of Minerals and Energy on a reconnaissance basis. The Louisiade Archipelago was also visited in 2009, when samples were collected for this study.

The most abundant rocks in the archipelago are sub-greenschist facies Calvados Schists of the Calvados Schist and the basalt and diorite dikes which occasionally intrude the Calvados Schists. The Louisiade Archipelago is not currently situated near a source of pelitic sediments, but it is possible that the source-region for the sedimentary protoliths of the Calvados Schist was rifted away during the opening of the Coral Sea (Weissel and Watts, 1979). In some locations in the archipelago, notably on Rossel and Sudest Islands, more extensive outcrops of basalt, gabbro, and serpentinite are also present. Detailed investigations of structural and contact
relationships in the Louisiade Archipelago have not yet been performed, so it is currently not possible to fully understand the tectonic context of the magmatic rocks in the Louisiade Archipelago.

2.3 Sampling Strategy

The sampling strategy in this study was to analyze representative examples of all the lithological units in the Woodlark Rift (Fig. 2) to assess whether the low grade and unmetamorphosed units in the Woodlark Rift are the protoliths of the higher grade metamorphic rocks (see discussion). A total of 39 samples (including one sample for which only U-Pb zircon data was acquired) were examined as part of this study (Table 1; Fig. 1). Samples were grouped on the basis of location, lithology, and structural context (Fig. 2). Samples from the D’Entrecasteaux and Misima Islands (western Woodlark Rift) are separated from samples from the Louisiade Archipelago (southeastern Woodlark Rift) due to the fact that rocks exposed in the D’Entrecasteaux Islands attained amphibolite conditions and higher, whereas rocks in the Louisiade Archipelago are only metamorphosed to subgreenschist facies. A brief description of the lithological units and rational for analysis is provided below, and sample descriptions are summarized in Table 1. Symbols used to denote lithostratigraphic units and individual samples in figures 1 and 2 are carried throughout the remainder of the figures.

2.3.1 Samples from the D’Entrecasteaux and Misima Islands

Four lithological units were sampled in the D’Entrecasteaux and Misima Islands (Fig 2a). These include mafic eclogites and amphibolites (n = 6), felsic and intermediate basement host gneisses (n = 7), upper plate rocks (n = 6), and Quaternary aged felsic lavas (n = 4). The upper plate samples consist of basalts, gabbro, and serpentinites that are in fault contact with the lower plate (eclogites, amphibolites, and basement gneisses) rocks. These upper plate rocks were
analyzed to explore the possibility that some of the mafic eclogites originated as part of the upper plate of the former subduction complex that is now being rifted apart. The Quaternary lavas consist of rhyolites, andesite, and obsidian that are found resting unconformably on the lower and upper plate rocks. These lavas were analyzed to explore the relationships between melting of the basement gneisses and the generation of lavas in an active rift.

2.3.2 Samples from the Louisiade Archipelago

Three lithological units were sampled in the Louisiade Archipelago (fig. 2b). These include basalt and gabbro (n = 6), andesite (n = 3), and subgreenschist facies metasediments of the Calvados Schist (n = 5). The basalt and gabbro samples were collected from dikes that cross cut the Calvados Schist, outcrops of basalt whose relationship to the Calvados Schist is unclear, and basalt and gabbro in fault contact with the Calvados Schist. These basalt and gabbro samples were analyzed to explore the possibility that these rocks are the unmetamorphosed or lower grade equivalents of the mafic eclogites and amphibolites in the D’Entrecasteaux Islands. The three samples of intermediate igneous rocks consist of diorites and andesites that occur in conjunction with the Calvados Schist and basalts/gabbros. The contact relationships between these groups of samples is not fully understood at this time, but these samples were analyzed for comparison with some of the basement host gneisses from the D’Entrecasteaux Islands to ascertain whether the intermediate magmatic rocks in the Louisiade Archipelago are the unmetamorphosed equivalents of some of the gneisses.

3. Analytical Methods

3.1 Major and Trace Elements

Rock samples were crushed and chipped using jaw-crushers at the Washington State University (WSU) GeoAnalytical Facility. The procedures of Johnson et al. (1999) and
Knaack et al. (1994) were followed, respectively, for XRF and ICP-MS analyses.

Representative aliquots of the crushed rocks (20 – 30g) were obtained using a rotating splitter and powdered in agate ball mills. Aliquots of powder (3.5g) were mixed with 7g of a lithium tetraborate flux and fused in graphite crucibles at ~1000°C for XRF analyses. The resulting beads were then reground in tungsten-carbide shatterboxes and fused again at ~1000°C. For ICP-MS analyses, 2g powder aliquots were mixed with 2g of lithium tetraborate flux and fused under the same conditions as for the XRF aliquots. These beads for ICP-MS trace element analysis were then reground in Fe shatterboxes and dissolved using a HF/HNO$_3$/HCLO$_4$ acid mixture. Major and trace element data raw data is present in full as an electronic supplement.

3.2 Hf & Nd Isotopes

Samples were powdered in an agate mortar and dissolved using a 10:1 HF/HNO$_3$ mixture in steel-jacketed Teflon bombs in an oven at 160°C for 4 – 5 days at the WSU Radiogenic Isotope Facility. All samples analyzed for isotopic compositions were spiked with a mixed $^{176}$Lu-$^{180}$Hf and $^{149}$Sm-$^{150}$Nd tracer in order to determine accurate parent/daughter ratios. Lu, Hf, Sm, and Nd were separated from the same solutions following methods described in Vervoort and Blichert-Toft, 1999 and Vervoort et al., 2004. The isotopic compositions of these elements were analyzed on a ThermoFinnigan Neptune$^\text{TM}$ multi-collector inductively coupled mass spectrometer (MC-ICP-MS) at WSU according to procedures described in Vervoort et al., 2004.

During the course of this study the following isotopic values were measured for the Aimes Nd and JMC 475 Hf standards: $^{142}$Nd/$^{144}$Nd = 1.141713 ± 40, $^{143}$Nd/$^{144}$Nd = 0.512120 ± 16, $^{145}$Nd/$^{144}$Nd = 0.348418 ± 14, $^{148}$Nd/$^{144}$Nd = 0.241557 ± 20, $^{150}$Nd/$^{144}$Nd = 0.236341 ±
$26 \ (2\sigma\ SD,\ n = 18); \ 176^{\text{Hf/177 Hf}} = 0.282152 \pm 0.02, \ 178^{\text{Hf/177 Hf}} = 1.467243 \pm 0.06, \ 180^{\text{Hf/177 Hf}} = 1.886840 \pm 0.02 \ (2\sigma\ SD,\ n = 24)$. Both the Hf and Nd isotope measurements were normalized to the accepted values for these standards ($^{143\text{Nd}}/^{144\text{Nd}} = 0.512138, \ 176^{\text{Hf/177 Hf}} = 0.282160$). For calculation of epsilon Nd and Hf values we used $^{143\text{Nd}}/^{144\text{Nd}}_{\text{chur(0)}} = 0.512630, \ 147^{\text{Sm}}/^{144\text{Nd}}_{\text{chur(0)}} = 0.1960, \ 176^{\text{Hf/177 Hf}}_{\text{chur(0)}} = 0.282785$, and $^{176\text{Lu}}/^{177\text{Hf}}_{\text{chur(0)}} = 0.0336$ (Bouvier et al., 2008). We were unable to measure the Hf isotopic composition for samples 04119a, 03069b, and 09045c and the Nd isotopic composition for samples 03069b and 09030a due to analytical difficulties. The Hf and Nd isotopic data is presented in figure 3 and tables 2 and 3.

3.3 U-Pb Zircon Dating

Zircons from samples selected for U-Pb geochronology were isolated using standard crushing and mineral separation procedures. Grains handpicked under a binocular microscope were mounted in epoxy with standards, polished to expose their centers, carbon-coated, and imaged with a SEM in CL at Syracuse University. U-Pb zircon analyses of polished grain mounts from samples 03118m and 041198 were conducted using the UCLA Cameca IMS 1270 ion microprobe. The aim of the ion microprobe analyses was to date within-grain spatial relationships for the eclogite to amphibolite facies rocks of the D’Entrecasteaux Islands and Misima Islands. In order to characterize the nature of the protolith of the low-grade Calvados Schist in the Louisiade Archipelago, LA-ICP-MS U-Pb analyses of zircons from samples 0906b and 09053a were made at the WSU Geoanalytical facility (e.g. Kosler et al., 2002).

3.3.1 LA-ICP-MS
All LA-ICP-MS U-Pb measurements were made using a New Wave Nd: YAG UV 213-nm laser coupled to a ThermoFinnigan Element 2 single collector, double-focusing, magnetic sector ICP-MS. Operating procedures and parameters are only outlined briefly here, but are discussed fully in Chang et al. (2006). Laser spot size and repetition rate were 30 µm and 10 Hz, respectively. The sample aerosol was delivered to the plasma in He and Ar carrier gases. Individual analyses were preceded by a short blank, and consisted of 300 sweeps (lasting ~35 s) through masses 204, 206, 207, 208, 232, 235, and 238. Time dependent mass fractionation, which is linear over the short time of the analysis, was corrected by regressing the time series data to the intercept at \( t = 0 \). Time-independent fractionation, which is the largest source of uncertainty in LA-ICP-MS U-Pb measurements, was corrected by normalizing U/Pb and Pb/Pb ratios of the unknowns to standards. In this study, we used two standards to monitor fractionation: Peixe, with an age of 564 Ma (Dickson and Gehrels, 2003), and FC-1, with an age of 1,099 Ma (Paces and Miller, 1993). However, only Peixe was used to correct the LA-ICP-MS U-Pb data reported here, due to the extremely low abundance of Precambrian aged zircons identified in the samples we examined.

3.3.2 Ion Microprobe

Ion microprobe \(^{238}\text{U}/^{206}\text{Pb}\) zircon age measurements were made using a Cameca ims 1270 high resolution, high sensitivity ion microprobe. Resolution of mass interferences within the mass range analyzed was possible due to the high mass resolution (~4500) of this instrument. A 12.5 kV primary \(^{16}\text{O}\) beam with a ~20 nA current and ~25 µm beam diameter were used for ablation of sample material. Zircon standard AS3 was mounted with the unknowns. Prior to analysis, mounts were lightly polished to remove the carbon coating that had been applied for SEM CL imaging, cleaned with dilute HCL, and coated with a ~30 nm Au film.
Intensities of monatomic U\(^+\), Th\(^+\), and P\(^+\) ions and \(^{94}\)Zr\(_2\)O\(^+\) and UO\(^+\) molecular ions were measured with a discrete dynode electron multiplier in peak jumping mode. Individual analyses consisted of 15 cycles with 15s count times. O\(_2\) flooding at 3 x 10\(^{-5}\) Torr was applied to the sample surface to enhance Pb yield. In-house software (ZIPS) was used to reduce the raw data.

4. Overview of Analytical Results

This section serves as an overview of the geochemical characteristics of the groups of samples (Fig. 2) analyzed in this study (Table 1). Major element data is presented in figure 3, Hf and Nd isotopic data is presented in figure 4, REE data is presented in figure 5, and U-Pb zircon data is presented in figure 6. Additional figures presenting geochemical data are composite figures used in the discussion. The geochemistry of individual samples, interpretation of the geochemical data, as well as comparison of the geochemistry of rocks in the Woodlark Rift with the eastern maring of Australia is discussed in subsequent sections.

4.1 D’Entrecasteaux and Misima Islands

4.1.1 Eclogites & Amphibolites

The six mafic eclogite and amphibolite samples, when plotted on igneous discrimination diagrams, have bulk compositions similar to basalt and basaltic andesite (fig. 3a). A few of these samples exhibit a tendency toward high TiO\(_2\) and low CaO (fig. 3b). The eclogites and amphibolites have \(\varepsilon\) Hf between +6.78 and +11.38 and \(\varepsilon\) Nd between +0.25 and +6.24 (fig. 4). The chondrite normalized REE patterns of the eclogites and amphibolites are predominantly flat, although a few of the eclogite (e.g. 89302a) and amphibolite (e.g. 04148a) samples are characterized by marked enrichment of the light and middle REE relative to the heavy REE (fig. 5a).
4.1.2 Basement Host Gneisses

The eight basement gneiss samples exhibit the widest range of bulk compositions of all the sample groups analyzed in this study (figs. 3a, 3b). These samples have ε Hf between -0.63 and +7.42, and ε Nd between +1.7 and +6.42 (fig. 4). The Chondrite normalized REE patterns of the gneisses are variable, and in some cases characterized by extreme depletion of the heavy REE relative to the light and middle REE (fig. 5b).

4.1.3 Upper Plate

Two of the six upper plate samples have basaltic bulk compositions, three have andesitic compositions (fig. 3a), and one sample is a serpentinized ultramafic sample for which major element data is not available. The upper plate samples have ε Hf of +2.61 to +10.11 and ε Nd of +0.8 to +5.43 (fig. 4), and there is no correlation between Hf and Nd isotopic composition and bulk composition in the upper plate samples we analyzed. The two upper plate basalts have completely flat chondrite normalized REE patterns, whereas the three andesites are enriched in the LREE (fig. 5c). The REE patterns of all the upper plate samples are similar except for a serpentinized ultramafic rock sample, which has much lower overall REE abundances than the other upper plate rocks (fig. 5c).

4.1.4 Quaternary Volcanic Rocks

The Quaternary volcanic rock samples all have rhyolitic bulk compositions, with one sample exhibiting greater than 75% SiO₂ (fig. 3a). The Quaternary volcanic rock samples have ε Hf of +9.47 to +10.82 and ε Nd of +5 to +6.11 (fig. 4). All of the chondrite normalized patterns of the Quaternary volcanic rock samples have negative Eu anomalies, enriched LREE, and flat MREE and HREE patterns (fig 5d).
4.2 Louisiade Archipelago

4.2.1 Subgreenschist Facies Calvados Schists

The five samples of subgreenschist facies Calvados Schist from the Louisiade Archipelago have bulk compositions that approximate andesite, dacite, and rhyolite (fig. 3a). These samples have $\varepsilon$ Hf of -1.23 to + 7.67 and $\varepsilon$ Nd of -3.39 to +2.38 (fig. 4). The chondrite normalized REE patterns of these samples are similar to one another with varying degree of Eu depletion, enriched LREE, and flat MREE and HREE (fig. 5e).

4.2.2 Basalt and Gabbro

Four of the six basalt and gabbro samples have basaltic bulk compositions. The other two samples have picrite-basaltic and basaltic-trachyandesitic compositions, respectively (fig. 3a). These samples have $\varepsilon$ Hf of +2.29 to +6.89 and $\varepsilon$ Nd of -0.9 to +2.97 (fig. 4). The chondrite normalized REE patterns of these samples range from flat to slightly LREE enriched; the pattern of the picrite-basalt sample exhibits significantly lower overall abundances of the REE as compared to the other samples, has a positive Eu anomaly, and is slightly HREE enriched (fig. 5f).

4.2.3 Intermediate Magmatic Rocks

Two of the three intermediate magmatic rock samples are basaltic andesites, the other sample is a dacite (fig. 3a). These samples have $\varepsilon$ Hf of +4.59 to +8.62 and $\varepsilon$ Nd of +3.07 to +4.68 (fig. 4). The chondrite normalized REE patterns of these samples are similar to one another, and display enrichment of the LREE relative to the MREE and HREE (fig. 5g).

4.3 U-Pb Zircon Results

For the ion microprobe U-Pb analyses conducted on samples 03118m and 04119a, the $^{206}\text{Pb}/^{238}\text{U}$ ages from zircon cores are presented in figure 6a as probability density curves.
Errors on age calculations are relatively high for some of the zircons in these samples due to low radiogenic Pb relative to common Pb. The data used in the $^{206}\text{Pb}/^{238}\text{U}$ age determinations are uncorrected for common Pb.

Most of the zircon grains from sample 03118m have oscillatory zoned cores that are mantled by dark CL overgrowths. Zircons from this sample were also analyzed as part of an ongoing study exploring under what conditions these young overgrowths form; for the purpose of this study, data presented here are for analyses targeting the zircon cores. The zircon cores from sample 03119m exhibit a range of ages from ~30 to ~110 Ma, but it is possible that the young zircon ages from this sample (e.g. <70 Ma) are the result of mixing between the Cretaceous aged cores and the younger overgrowths formed during metamorphism.

Zircons in sample 04119a, an intermediate gneiss from the lower plate of Misima Island, have an internal grain morphology that is similar to the zircons in sample 03118m. However, zircons in sample 04119a do not exhibit the same dark CL overgrowths as the zircons in sample 03118m. Ion microrobe analyses were targeted at the zircon cores in this sample, revealing a nearly homogeneous ~100 Ma zircon population.

Zircon in two samples of the Calvados Schist (0906b and 0953a) from the Louisiade Archipelago were dated using the LA-ICP-MS technique, and similarly to the ion microprobe U-Pb results, are presented here as probability density curves of the $^{206}\text{Pb}/^{238}\text{U}$ ages single grain ages from these samples (fig. 6b). Sample 0906b is not included in the Hf, Nd, and trace/REE element data set. We were able to obtain >100 zircon grains from 0906b, whereas only ~50 grains were separated from 0953a. Zircon in both of these samples exhibited similar grain morphologies and internal grain structure under CL. The large spot size (~30
micron diameter) of the laser made it difficult to target specific domains within zircons from the two samples of Calvados Schist from the Louisiade Archipelago. However, the results from both samples are similar and are characterized by a prominent early to middle Cretaceous zircon population. Out of the 100 grains analyzed in sample 0906b, only two zircons fell outside the ~90 – 200 Ma age range: one with a ~600 Ma age and one with a ~1.1 Ga age. All of the zircons analyzed in 0953a had ages between ~90 and ~200 Ma, with the bulk of the analyses clustering at ~100 Ma.

5. Tectonic origin of metamorphic rocks in the Woodlark Rift

There are several lines of evidence based on the geochemical and geochronological data supporting a tectonic link between metamorphic rocks with felsic and intermediate bulk compositions (e.g. basement host gneisses in the D’Entrecasteaux and Misima Islands, and the Calvados Schist in the Louisiade Archipelago) exposed along the southern margin of the Woodlark Rift, and the Cretaceous aged Whitsunday Volcanic Province (WVP) in northeastern Queensland (Bryan et al., 1997).

The WVP is currently situated along the rifted Australian passive margin and was active from ~120 to ~90 Ma (Bryan et al., 1997). The WVP lies at the northern end of a >900 km long continental rift that marked the eastern margin of Gondwana during the Cretaceous. Greater than 10^5 km^3 of volcanic rocks were erupted in the WVP, much of which was eroded and transported westward into Australia’s Great Artesian and Otway/Gippsland basin systems (Bryan et al., 1997). The remnants of the WVP currently occupy the Queensland coast of Australia, but the eastern extent of the WVP may have been rifted away from Australia during the Late Mesozoic and Cenozoic.
The primary evidence for the tectonic link between the metamorphic rocks in the Woodlark Rift and the WVP comes from a comparison of their the Nd and Hf isotopic, and trace/REE compositions, as well as from the U-Pb ages of inherited zircons.

5.1 Nd & Hf Isotopic Evidence

The range of Nd isotopic compositions for Cretaceous basalts, andesites, and rhyolites comprising the WVP (Ewart et al., 1992) neatly bracket the range of Nd isotopic compositions displayed by the metamorphic rocks from the Woodlark Rift (Fig. 4). The only notable exception is one of the samples of the Calvados Schist from the Louisiade Archipelago (0945a), which has an εNd of -3.4. This sample also has the most evolved Hf isotopic composition (εHf = -0.79) of all the samples we analyzed, but still does not exhibit the strongly evolved Hf and Nd isotopic signature that would be expected of Archean or Proterozoic continental crust.

The Nd isotopic compositions of all the metamorphic samples from the Woodlark Rift are much more juvenile than any of the Paleozoic and Precambrian metamorphic and magmatic basement rocks currently exposed near the WVP. For example, late Paleozoic to Mesoproterozoic metamorphic and magmatic rocks in the Georgetown Inlier, are located to the northeast of the Cretaceous WVP (Withall and Mackenzie, 1980; Black and McCulloch, 1984; McDonald et al., 1997). Late Paleozoic and early Mesozoic granites are also exposed to the south of the WVP in the New England Fold Belt. The range of present day epsilon Nd values of the Proterozoic granites and volcanic rocks in the Georgetown Inlier (Black and McCulloch 1984, 1990), and granites and associated metasedimentary country rocks in the New England Fold Belt (Hensel et al., 1985) are shown on figure 4 for comparison with the isotopic data from the WVP and the rocks in the Woodlark Rift. It is clear that the Nd
isotopic compositions of the Woodlark Rift samples are much more similar to the WVP than to any of the areas of older crust in eastern Australia.

The relatively primitive isotopic compositions of the WVP granitoids implies that melting of an older, more geochemically evolved crust, was not involved in the production of the Cretaceous rift-related granitoids (Ewart et al., 1992). The more evolved Nd isotopic composition ($\varepsilon$ Nd = -3.4) of one of the samples of the Calvados Schist may indicate that a component of older continental crust, possibly derived from one of the areas of Precambrian basement currently exposed near the WVP (e.g. McDonald et al., 1997), is present in the protoliths of some of the metamorphic rocks in the Woodlark Rift. However, the majority of metamorphic rocks analyzed appear to be isotopically juvenile and similar to rocks from the WVP.

There is currently no whole rock Hf isotopic data for the WVP reported in the literature and it is therefore not possible to directly compare the Hf isotopic compositions of rocks in the Woodlark Rift and the WVP. The Hf isotopic compositions (fig. 4) of the metamorphic rocks from the Woodlark Rift are in agreement with their Nd isotopic compositions (e.g. Vervoort et al., 1999), and similarly to the Nd isotopic results, none of the metamorphic rock samples analyzed display the highly evolved Hf isotopic compositions expected of ancient (i.e Paleoproterozoic or Archean) crust. The overall juvenile Hf isotopic compositions of the metamorphic rock samples from the Woodlark Rift are consistent with a tectonic affinity related to mantle derived magmas produced during a Cretaceous continental rifting event.

### 5.2 Evidence from U-Pb Ages of ‘Inherited’ Zircon Grains

The four samples selected for U-Pb zircon analysis represent both the high grade rocks in the western (D’Entrecasteaux Islands) and central (Misima Island) parts of the Woodlark Rift, as
well as the subgreenschist facies rocks on the southern rifted margin (Louisiade Archipelago) of the Woodlark Rift. The fact that 90 – 100 Ma inherited zircons are the dominant age population in all four of these samples suggests that either the felsic and intermediate metamorphic rocks in the Woodlark Rift originated as part of the WVP, or that some other Cretaceous magmatic province was the dominant source region for the sedimentary protoliths of these rocks. Marine sediments from the Trobriand Basin have produced 2.78 Ga zircons (in igneous and metamorphic clasts; Baldwin and Ireland, 1995), but based on results presented here, a significant component of older continental crust exposed in the Woodlark Rift has yet to be identified.

Volcanic and intrusive activity in the WVP occurred between 132 and 95 Ma, with the peak of activity between 120 and 105 Ma (Bryan et al., 1997). In the two samples of the Calvados Schist that were sampled for U-Pb zircon analysis, zircons are almost exclusively Early Cretaceous (fig. 6b). The only exceptions are two zircon grains with older (Late Precambrian and Grenvillian ages) U-Pb ages from sample PNG 0906b (not shown in the probability plot in figure 6b). The Hf and Nd isotopic compositions of this sample (0906b) were not measured, so it is difficult to further assess the relative contributions of young versus older crust in this sample.

The amphibolite gneiss from Misima (04119b) Island, which also almost exclusively contains ~100 Ma zircon (Fig. 9a), has a Nd isotopic composition (ε Nd = + 6.42; no Hf isotopic data available) that is inconsistent with a significant contribution from ancient continental crust. U-Pb zircon dating of sample 03118b, a felsic gneiss metamorphosed at eclogite facies conditions, also failed to produce any ages older than ~100 Ma (Fig. 6a)

5.3 Trace Element Evidence
The Th/Yb and Ta/Yb ratios (Fig 7) of the Calvados Schist samples from the Louisiade Archipelago are similar to the WVP. The amphibolite and eclogite facies basement host gneisses from the western parts of the Woodlark Rift (D’Entrecasteaux and Misima Islands) exhibit significantly more variable Th/Yb and Ta/Yb ratios that are, for the most part, not similar to the WVP. The relationships between the Th/Yb and Ta/Yb ratios of the metamorphic rock samples from the Woodlark Rift and the WVP (Fig. 7) are mirrored by the REE compositions of these rocks (Fig. 5c, 5f). The subgreenschist facies Calvados Schists lie within the range of REE compositions defined by WVP magmas (Fig. 5f), whereas the (U)HP and amphibolite facies gneisses exhibit a much broader range of REE compositions, and in many cases, fall well beyond the range of compositions defined by the WVP (Fig 5c; Ewart et al. 1992). A few basement host gneiss display extreme HREE depletion (Fig. 5c), and have Th/Yb ratios as high as ~210 (Fig. 7).

The disparity in the trace/REE characteristics of the basement host gneisses and the Calvados Schists in the western and eastern parts of the Woodlark Rift, respectively, does not necessarily indicate that these groups of rocks have different tectonic origins. The gneisses and schists have similar Hf and Nd isotopic compositions, and predominantly contain Cretaceous aged inherited zircons. Partial melting of felsic and intermediate lithologies has accompanied the formation of Pliocene/Pleistocene MCCs in the D’Entrecasteaux and Misima islands (e.g. Hill et al., 1995; Little et al., 2011), and is probably responsible for the disparity in the trace/REE characteristics of the basement host gneisses and the Calvados Schist. If partial melting occurred recently (e.g. within the last few m.y.), it is not surprising that the high and low grade rocks still exhibit similar Hf and Nd isotopic compositions, despite having widely disparate trace/REE characteristics. This is because these two groups
of samples would not have had enough time to evolve drastically divergent isotopic compositions since the time of differentiation due to partial melting.

5.4 Origin of Mafic Eclogites & Amphibolites

Mafic magmas were erupted in the WVP, and it is possible that some of these are the protoliths of the mafic eclogites and amphibolites that occur in the D’Entrecasteaux and Misima Islands. However, the eclogites and amphibolites have REE patterns that are not similar to WVP basalts and gabbros (Fig. 5a), nor is there any overlap between the Th/Yb and Ta/Yb ratios of WVP rocks when compared with the eclogites and amphibolites (Fig. 7). Average WVP basalts and gabbros (~50 % SiO₂) also have \(^{143}\text{Nd}/^{144}\text{Nd}\) of ~0.51295 (Ewart et al., 1992), whereas the eclogites and amphibolites exhibit significantly more variable \(^{143}\text{Nd}/^{144}\text{Nd}\) (fig. 8). These three obvious differences do not necessarily preclude a relationship to the WVP, but indicate that some of the mafic metamorphic rocks may have a different origin (e.g. Davies and Warren, 1992).

Two other possibilities for the protoliths of the mafic eclogites and amphibolites can be explored in light of the geochemical data presented here: a) subduction erosion of the mafic upper plate of an Eocene obducted ophiolite, which now comprises the upper plate of the MCCs in the D’Entrecasteaux Islands, b) intrusion of basalt or gabbro into subducted continental crust, followed by metamorphism prior to Late Miocene exhumation.

An important consideration for these two possibilities is the recent work of Zirakparvar et al. (2011), who used the Lu-Hf system to date metamorphic garnet in two rocks from the D’Entrecasteaux Islands. An amphibolite gneiss in the shear zone carapace on Goodenough Island yielded a Lu-Hf garnet age of ~68 Ma, interpreted by Zirakparvar et al. (2011) as recording garnet growth in response to continental subduction and ophiolite obduction,
followed by incorporation into an MCC bounding shear zone during the late Cenozoic. A separate sample examined by Zirakparvar et al. (2011) was a coesite eclogite from the MCC core zone, yielding a ~7 Ma Lu-Hf garnet age that records the time when when a basaltic partial melt of the mantle was intruded into subducted continental lithosphere and crystallized at UHP conditions. These two Lu-Hf garnet age results indicate that the some of the mafic eclogites and amphibolites have an origin that dates back to at least the late Mesozoic (e.g. subduction erosion of the upper plate), whereas others originated in the late Cenozoic (e.g. intrusion of basalt and gabbro into subducted continental lithosphere).

Before comparing the geochemistry of the mafic eclogites and amphibolites with that of unmetamorphosed mafic rocks in the Woodlark Rift, an important consideration is the that the coesite eclogite examined by Zirakparvar et al. (2011), interpreted as having originated as a basaltic melt that crystallized at (U)HP conditions at ~7 Ma, was also examined as part of this study (sample 89321). This sample can essentially be examined as a known example of an eclogite that originated by emplacement of a mafic magma into subducted continental crust. The Hf and Nd isotopic compositions of this sample are more radiogenic than any of the other eclogite or amphibolite samples. However, this does not preclude a similar origin for the other samples in this group due to the possibility that basalt emplacement into subducted continental crust could have occurred at many different times throughout the history of these rocks.

5.4.1 Subduction erosion of the upper plate

It is possible that some of the mafic eclogites and amphibolites in the D’Entrecasteaux and Misima Islands originated as fragments of the upper plate that were entrapped by subducting continental crust during the late Mesozoic. The MCCs now exposing these
eclogites and amphibolites are flanked by an upper plate of mafic and ultramafic rocks (e.g. Davies and Warren, 1988). The upper plate of these MCCs provides an opportunity to sample the upper plate of the Late Mesozoic subduction complex where the protoliths of the metamorphic rocks in the Woodlark Rift were originally subducted. Several samples of the upper plate rocks were analyzed for comparison with the mafic eclogites and amphibolites.

An important note at the onset of this section is that sample 89321, which has been documented by Zirakparvar et al. (2011) to have originated as a basaltic melt of the mantle that was intruded into subducted continental lithosphere at ~7 Ma, is excluded from this discussion.

The REE patterns of the eclogites and amphibolites are not in complete agreement with the REE patterns of the upper plate samples. With one notable exception (a serpentinized ultramafic rock), all of the upper plate samples have identical HREE behavior. With the exception of the serpentinized ultramafic sample, the REE patterns of the upper plate samples can be described in two ways: negatively sloped (e.g. LREE enriched), or flat (Fig. 5b). The HREE behavior of the eclogites and amphibolites appears more varied than the upper plate samples, but a few of the eclogites and amphibolites do have the completely flat REE patterns exhibited by some of the upper plate samples. Some of the eclogite and amphibolite samples also display negatively sloped REE patterns observed in the upper plate samples, but with more perturbations to the MREE than is observed in the REE patterns of the upper plate samples.

Similarly to the REE, there is some overlap in the Th/Yb and Ta/Yb ratios of the eclogites and amphibolites group and the upper plate group. However, the individual samples within these two groups also display considerable variation in Th/Yb and Ta/Yb, and it is
difficult to make a direct comparison on the basis of their REE and trace element behavior, since the effects of (U)HP metamorphism and metasomatism have probably altered the REE and trace element characteristics of the eclogites and amphibolites.

The most radiogenic Hf and Nd isotopic compositions of the eclogites and amphibolites are more radiogenic than the most radiogenic upper plate samples, but there is also considerable overlap between these groups (fig. 4). On the basis of the trace/REE and isotopic data, it seems reasonable to conclude that some of the eclogites and amphibolites are metamorphosed fragments of the upper plate that were captured during continental subduction.

5.4.2 Basalt & gabbro emplaced into subducted crust

In addition to comparing the eclogite and amphibolite samples with the coesite eclogite sample 89321, which is already documented to have originated as a basaltic melt that was intruded into subducted continental crust (Zirakparvar et al., 2011), basalts and gabbros intruded into the subgreenschist facies Calvados Schists in the Louisiade Archipelago can also be used for this purpose because these mafic magmas may represent the unmetamorphosed equivalents of the eclogites and amphibolites in the D’Entrecasteaux and Misima Islands. The isotopic and trace element data is the only way to establish whether the eclogites and amphibolites in the D’Entrecasteaux and Misima Islands and the basalts and gabbros in the Louisiade Archipelago have a common origin, since geochronologic data pertaining to the timing of metamorphism and deformation in the Louisiade Archipelago, or the timing of basalt and gabbro intrusion into the low grade schists, is currently unavailable.

The eclogites and amphibolites in the D’Entrecasteaux and Misima islands have similar REE compositions (figs. 5a & 5f) and Th/Yb and Ta/Yb ratios (fig. 7) to the Louisiade
Archipelago basalts and gabbros. However, the Hf and Nd isotopic compositions of the eclogites and amphibolites are distinctly more juvenile than the Louisiade Archipelago basalts and gabbros even when sample 89321 is excluded (Fig 4). The contrasting Hf and Nd isotopic compositions, but similar trace/REE characteristics, of the eclogites and amphibolites and the Louisiade Archipelago basalts and gabbros suggests that some of these samples could have similar origins, but that they were separated from their isotopic reservoirs at different times.

5.5 Tectonic Reconstruction

The isotopic and geochronological similarities between the felsic and intermediate metamorphic rocks in the Woodlark Rift and the WVP (Whitsunday Volcanic Province, northeastern Australia) provide compelling evidence that these two regions are tectonically linked. There are three tectonic scenarios (Fig. 9) that could satisfactorily account for the apparent linkage between the WVP and some of the metamorphic rocks in the Woodlark Rift. The possible reconstructions proposed in figure 9 only take into account the tectonic linkage between the felsic and intermediate metamorphic rocks and the Whitsunday Province, and do not consider the multiple possible origins of the mafic eclogites and amphibolites that are associated with the basement host gneisses (section 4.2).

The first possibility is that felsic and intermediate metamorphic rocks in the Woodlark Rift originated as the northern extension of the WVP (Fig. 9a). One problem with this scenario is that no interbedded volcanic rocks were observed in the low grade Calvados Schist in the Louisiade Archipelago. The finely laminated nature of the Calvados Schist suggests that these rocks have a predominantly sedimentary origin. (U)HP and amphibolite facies rocks now exposed in the D’Entrecasteaux and Misima Islands do not preserve primary
textures that would enable identification of a volcanic interbed. It is not possible to rule out the presence of Cretaceous volcanic or igneous protolith for some of the metamorphic rocks in the Woodlark Rift, however the available textural evidence points towards mainly sedimentary protoliths.

The lack of observed volcanic interbeds in the Woodlark Rift opens up a second possibility, which is that the metamorphic rocks in the Woodlark Rift originated as volcaniclastic sediments derived from the Late Cretaceous WVP (Fig. 9b). In this case, the protoliths of the metamorphic rocks exposed in the Woodlark Rift are similar to the volcaniclastic sedimentary successions occupying the Great Artesian and Ottway/Gippsland Basins in eastern Australia (Bryan et al. 1997).

A third possibility is that the protoliths of metamorphic rocks in the Woodlark Rift originated as either sediments derived from, or as part of, a Cretaceous volcanic province that was not a continuous extension of the >2,500 km long rift system in eastern Australia (Fig. 9c). An important consideration in these tectonic scenarios is that the rocks currently exposed in the Woodlark Rift were situated in closer proximity to northeastern Australia until the Early Cenozoic opening of the Coral Sea Basin. Their proximity would have facilitated transport of volcaniclastic sediments derived from the WVP into a Cretaceous depositional basin situated in the present-day location of the Coral Sea.

5.5.1 Complicating factors in the proposed reconstructions

Mortimer et al. (2008) used petrographic, geochemical, K-Ar dating, and tracer isotope data from sedimentary and metasedimentary xenoliths in drill cores recovered from the Queensland Plateau (ODP824-825) and Lord Howe Rise (NORFANZ 85) to assess the tectonic affinity of these offshore provinces. These authors found that the xenoliths from the
Queensland Plateau and Lord Howe Rise have geochemical and geochronological signatures that are consistent with these two provinces being offshore extensions of the New England Orogen. The range of Nd isotopic compositions of metamorphic and metasedimentary xenoliths from the two offshore sites investigated by Mortimer et al. (2008) are shown on figure 4 for comparison with the Woodlark Rift data. It is clear that the Queensland Plateau and Lord Howe Rise xenoliths are not as radiogenic as any of Woodlark Rift samples or the WVP.

The ODP824-825 site on the Queensland Plateau essentially lies on a line between the present-day northern end of the WVP and the western Woodlark Rift, and it is therefore surprising that the metasedimentary and sedimentary xenoliths from ODP824-825 appear related to the New England Orogen, and not the WVP (Fig. 9). However, Mortimer et al. (2008) were not able to rule out the possibility that minor intrusive igneous rocks in the ODP holes are related to the WVP. This problem illustrates the complexity of the arrangement of offshore basement provinces in the region north and east of Australia, which in turn illustrates the challenges of recognizing the dispersed fragments of an active margin (e.g. the eastern margin of Australia in Gondwana).

6. Discussion: implications of Quaternary felsic volcanism in the D’Entrecasteaux Islands for the surface expression of lithospheric rupture in active plate tectonic settings

6.1 Mantle Source of Quaternary Calc-Alkaline Lavas

Quaternary aged rhyolites are commonly found unconformably on the upper and lower plates of the D’Entrecasteaux Islands MCCs, even though rift-related peralkaline basalts currently predominate as the most commonly erupted volcanic rock type in the region (Smith, 1981; Smith, 1982; Smith and Clarke, 1991; Stolz et al., 1993). Partial melting of young-arc
protocrust and metasomatized upper-mantle wedge, as well as extensive differentiation of mantle derived melts in a shallow magma chamber have all been proposed as causes for the range of volcanic rock compositions seen in the D’Entrecasteaux Islands (Stolz et al. 1993). What is especially interesting in the D’Entrecasteaux Islands is that the Quaternary felsic lavas have bulk-compositions that are nearly identical to felsic metamorphic basement rocks through which they are being erupted (Fig. 3). This correlation opens up the possibility that these lavas are simply the extrusive equivalents of the deeply seated metamorphic rocks that have undergone partial melting (Hill et al., 1995; Little et al., 2011).

In order to ascertain whether melting of the metamorphic basement rocks in the D’Entrecasteaux Islands is contributing to the production of Quaternary felsic lavas, four Quaternary silicic volcanic rock samples were analyzed as part of this study. If the Quaternary calc-alkaline volcanic rocks in the D’Entrecasteaux Islands are the extrusive products of partially melted basement gneisses, then the isotopic and geochemical signature of the basement gneisses should be expressed in the Quaternary lavas. Large granodiorite plutons are common in the D’Entrecasteaux Islands, and could have formed by fractional crystallization of mantle-derived magmas coupled with assimilation of deep crustal rocks, by partial melting of the lower crust in response to heat input from the mantle-derived magmas, or by isothermal decompression of felsic and intermediate lithologies during MCC formation (Hill et al., 1995). In some cases, these granodiorite plutons are weakly foliated, indicating that rock exhumation and partial melting are occurring simultaneously (Little et al., 2011). One of the ‘felsic’ D’Entrecasteaux Islands basement gneisses (89327) analyzed in this study is a weakly foliated granodiorite with a 1.91 Ma K/Ar age (Baldwin et al., 1993). The Hf (εHf = +4.34) and Nd (εNd = +1.7) isotopic
compositions of this weakly foliated granodiorite are not as radiogenic as any of the Quaternary lavas (Figs. 5b & 5c).

The four samples of Quaternary lava that we analyzed have greater than 70% SiO₂, and exhibit similar bulk compositions to the felsic basement host gneisses (Fig. 2). However, the felsic basement gneisses from the D’Entrecasteaux Islands are strongly HREE depleted, whereas the Quaternary volcanic rocks are not (Fig. 5f). The Quaternary lavas also display pronounced negative Eu anomalies not observed in the basement gneisses. Another important observation is that the Hf and Nd isotopic compositions of the Quaternary lavas (εHf = +9.47 to +10.82; εNd = +5 to +6.11) are more juvenile than the basement gneisses from the D’Entrecasteaux Islands (εHf = −0.63 to +7.42 ; εNd = +1.7 to +4.34). Figures 3 & 8 which display the variation of ²⁴³Nd/²⁴⁴Nd (fig. 7), and all the major elemental oxides (fig. 3) as a function of silica content, illustrates the fact that the Quaternary volcanic rocks have similar bulk compositions to the felsic basement gneisses in the D’Entrecasteaux Islands, but Hf and Nd compositions resembling those of the eclogites, amphibolites, and upper plate rocks. This is also supported by fig. 10, which is a plot of Th/Lu vs ¹⁷⁶Hf/¹⁷⁷Hf.

The lack of agreement between the isotopic and trace/REE compositions of the basement gneisses and the Quaternary lavas is surprising in light of the fact that partial melting of the basement is occurring in close spatial and temporal proximity to generation of Quaternary silicic lavas in the D’Entrecasteaux Islands. These lavas do not appear to be the extrusive equivalent of melted basement rocks; instead the juvenile isotopic compositions and negative Eu anomalies displayed by the Quaternary felsic lavas suggest that they are highly fractionated mantle melts, and probably the more fractionated equivalents of the peralkaline rift-related basalts in the Woodlark Rift (e.g. Smith, 1982).
6.2 The Surface Expression of Ruptured Lithosphere in the D’Entrecasteaux Islands

The D’Entrecasteaux Islands provide a perfect setting to assess the surface expression of initial lithospheric rupture due to the fact that they are situated in a zone of active extension just ahead of the seafloor spreading rift tip in the Woodlark Rift (e.g. Taylor et al., 1995). Quaternary felsic lavas with mantle Hf and Nd isotopic signatures are one surficial expression that the lithosphere has ruptured in the D’Entrecasteaux Islands region. What is somewhat remarkable is that these lavas have not inherited the isotopic or geochemical signature of the metamorphic basement exposed in close proximity to their eruptive centers—this is despite the fact that the basement rocks have undergone partial melting in response to MCC formation, an ongoing process (Hill et al., 1995). The Quaternary felsic lavas have essentially followed a direct path to the surface from their mantle source, suggesting extremely thin lithosphere beneath the region where these lavas are being erupted. This observation is not surprising given that seismic observation of the D’Entrecasteaux Islands has revealed that the high elevations in the D’Entrecasteaux Islands are not supported by a thick crustal root (Abers et al., 2002). Instead the asthenosphere beneath these islands has upwelled and has essentially replaced the sub-continental mantle lithosphere (Abers et al, 2002).

What is amazing in the D’Entrecasteaux islands is how rapidly asthenospheric upwelling has occurred, and that the arrival of the asthenosphere beneath the D’Entrecasteaux Islands region is already manifest in the magmatic products of rifting ahead of the seafloor spreading rift tip. It is possible to roughly assess the rapidity with which the lithosphere has been ruptured due to some pre-existing constraints for the tectonic development of the
D’Entrecasteaux Islands. A ~7 Ma Lu-Hf garnet isochron age from a coesite eclogite (sample 89321 in this study) has been interpreted by Zirakparvar et al (2011) as reflecting the time when a basaltic partial melt of the upper mantle became entrapped within cold formerly subducted continental lithosphere where it crystallized at UHP conditions. This crystallization event signaled the earliest onset of partial mantle melting in response to the onset rifting in this region, and occurred at a time when continental lithosphere was still subducted to at least ~100 km in order to produce a UHP assemblage (Zirakparvar et al 2011).

It has taken roughly ~7 m.y. for a former active margin to undergo enough thinning such that mantle melts are now reaching the surface without crustal contamination and seismic reflection studies are able to document the presence of upwelled asthenosphere beneath the D’Entrecasteaux Islands. However, it is possible to only roughly assess the rate at which the sub-continental mantle lithosphere has been replaced due to the fact that the geometry of the lithosphere/asthenosphere and related subduction thrust at ~7 Ma is not known.

The possibility that a subduction complex with continental lithosphere entrained to a depth of at least ~100 km can become the site of lithospheric rupture and consequent asthenospheric upwelling within the course of ~7 m.y. implies that this processes can occur relatively rapidly in geologic time. Assuming rifting continues in the D’Entrecasteaux Islands region, the highly fractionated felsic volcanic products of mantle melting will give way to MORB type basalts. The remaining metamorphic rocks, now exposed in MCCs, will also eventually become dispersed. If this future ocean basin is then subducted, the dispersed metamorphic rocks, once situated at the site of lithospheric rupture, will be overprinted during tectonic collisions. However, it will still be possible to recognize rocks that were
situated at the site of initial lithospheric rupture on the basis of the current association of lithostratigraphic units in the D’Entrecasteaux Islands.

In the D’Entrecasteaux Islands, the association of lithostratigraphic units indicating active lithospheric rupture are: A) mafic eclogites (e.g. sample 89321; Zirakparvar et al. 2011) with crystallization ages that predate the eruption of felsic lavas by few m.y., and B) the presence of older felsic and intermediate gneisses that have undergone partial melting with final crystallization occurring between the time of eclogite crystallization and the eruption of felsic lavas with mantle Nd and Hf isotopic signatures. Another important observation is that the eclogites and felsic lavas have similar Hf and Nd isotopic compositions due to the fact that they were separated from the mantle (e.g. Zirakparvar et al., 2011) at roughly the same time, whereas the felsic and intermediate basement host gneisses have slightly more evolved Hf and Nd isotopic compositions reflecting their Late Cretaceous aged volcanoclastic protoliths.

The current tectonic setting of the D’Entrecasteaux Islands is unique on Earth- there are no other known locations where a former subduction complex is actively being rifted apart. Identifying these key relationships requires the use of multiple geochronological and geochemical tools, as well as detailed knowledge of the field relationships and metamorphic history of the region. This work is necessary, however, as the results from the D’Entrecasteaux Islands indicate that lithospheric rupture occurs rapidly- essentially transforming a former subduction complex with a history of continental subduction into an area where the surface products of mantle melting dominate the geology. The observations from the D’Entrecasteaux Islands can be applied to other regions that are no longer active, but where the geologic record is preserved.
7. Conclusions

Isotopic, trace/REE, and geochronological data for the primary lithostratigraphic units (Fig. 2) in the Woodlark Rift of southeastern PNG indicate:

1) Metamorphic rocks with felsic and intermediate bulk compositions appear uniformly related to Cretaceous magmatism in eastern Gondwana. There are three overarching possibilities for this tectonic link:
   a) The rocks are the metamorphosed remnants the northeastern segment of the Whitsunday Volcanic Province (WVP), which was partially rifted away from Australia.
   b) The rocks are the metamorphosed sedimentary products of WVP volcanism, but did not originate as an extension of the WVP.
   c) The rocks are the metamorphosed sedimentary remnants of some other Cretaceous magmatic province in this region.

2) Mafic eclogites and amphibolites have multiple possible origins including:
   a) Basalts and gabbros that were erupted in the WVP and then subducted and metamorphosed along with the felsic and intermediate WVP related rocks.
   b) Fragments of the upper plate entrained in the subduction complex during subduction erosion where these rocks were initially subducted during the Late Mesozoic.
   c) Mafic partial melts of the mantle that experienced crystallization within subducted continental lithosphere in response to the earliest phases of rifting in the Woodlark Rift or some other tectonic events.
3) Quaternary felsic lavas in the D’Entrecasteaux Islands have not inherited the isotopic or geochemical signature of the metamorphic basement despite being erupted within a zone of active continental extension where partial melting of the basement has been documented. This suggests lithospheric rupture has occurred in the region.

4) There is a distinctive association of lithostratigraphic units that can be used to identify rocks that are situated at the site of lithospheric rupture. It is possible for a subduction complex with continental lithosphere entrained to at least 100 km depth to become the site of lithospheric rupture within the span of ~ 7 Ma.

Acknowledgments

Garret Hart, Charles Knaack, and Rick Conrey at Washington State University are thanked for help with analytical aspects of this project. National Science Foundation (NSF) grant 0709054 to S. Baldwin, P.G. Fitzgerald, and L.E. Webb provided support for this work. We thank Paul Fitzgerald, Laura Webb, Tim Little, Brian Monteleone, Katie Pieters, Alec Waggoner, Brad Hacker, and Leigh Castellani for discussions during field work.
References


Cluzel D., Black P., Picard C., and Nicholson K., 2010. Geochemistry and tectonic setting of Matakaoa Volcanics (East Coast Allochthon, New Zealand); supra-subduction zone affinity, regional correlations and origin. Tectonics. (in press)


Figure Captions & Table Headings

Figure 1. Geologic map of southeastern Papua New Guinea with sample locations. Symbols used to denote different lithological units (e.g. sample groups) are used throughout. Refer to figure 2 for schematic diagram illustrating structural relationship of lithostratigraphic units to one another. G.I. = Goodenough Island; F.I. = Fergusson Island; N.I. = Normanby Island; M.I. = Misima Island.

Figure 2. Simplified schematic cross-sections illustrating general structural relationships between groups of samples analyzed in this study. Symbols used in this figure to denote different lithological units are used throughout remaining figures.

Figure 3a. Total Alkalis versus SiO₂. Symbols are as in figure 1. All of the rock samples analyzed in this study are plotted on this figure; non-magmatic rock types are grayed out but are still shown to illustrate their bulk compositions.

Figure 3b. Harker variation diagrams: elemental oxides versus SiO₂. Symbols are as in figures 1 & 2. Major element data is contained in full in the electronic appendix for chapter 2.

Figure 4. Plot of Nd & Hf isotopic compositions for samples in this study. Nd isotopic compositions of rocks in the Queensland Plateau, Lord Howe Rise, Georgetown Inlier, New
England Fold Belt, and Whitsunday Volcanic Province of eastern Australia are also shown for comparison with our data. Hf and Nd isotopic data is presented in tables 2 & 3.

Figure 5. Chondrite normalized (values from McDonough and Sun, 1995) trace/REE variation diagrams for: a) eclogites and amphibolites in the D’Entrecasteaux and Misima islands, b) upper plate rocks from the D’Entrecasteaux and Misima islands, c) felsic & intermediate basement gneisses in the D’Entrecasteaux and Misima islands, d) Quaternary volcanic rocks in the D’Entrecasteaux Islands, e) basalt, gabbro, and serpentinite in the Louisiade Archipelago, f) felsic & intermediate schists from the Louisiade Archipelago, g) intermediate magmatic rocks in the Louisiade Archipelago. REE and trace element data is contained in full in the electronic appendix for chapter 2.

Figure 6a. Results of ion microprobe U-Pb zircon dating. Probability density curves for $^{206}$Pb/$^{238}$U single grain ages for individual analyses of zircons from samples 03118m and 04119b. U-Pb data is contained in full in the electronic appendix for chapter 2.

Figure 6b. Results of LA-ICP-MS U-Pb zircon dating. Probability density curves for $^{206}$Pb/$^{238}$U single grain ages for individual analyses of zircons from samples 0906b and 09053a. U-Pb data is contained in full in the electronic appendix for chapter 2.

Figure 7. Th/Yb vs Ta/Yb for samples in this study. Data from the Whitsunday Volcanic Province and other terrestrial geochemical reservoirs are shown for comparison with the rocks in the Woodlark Rift (Ewart et al. 1992).
Figure 8. $^{143}\text{Nd}/^{144}\text{Nd}$ vs SiO$_2$; representative examples of magmatic rocks from the Whitsunday Volcanic Complex are also shown.

Figure 9. Tectonic scenarios to explain the geochemical similarities between metamorphic rocks in the Woodlark Rift and the Whitsunday Volcanic Province (WVP). Q.P. = Queensland Plateau; L.H.R. = Lord Howe Rise; G.I. = Georgetown Inlier; C.B. = Carpentaria Basin; N.E.O = New England Orogen; G.A.B. = Great Artesian Basin. Samples investigated by Mortimer et al. (2008): 1 = xenoliths in ODP824-825 (Queensland Plateau); 2 = xenoliths in NORFANZ 85 (Lord Howe Rise).

Figure 10. Th/Lu vs $^{176}\text{Hf}/^{177}\text{Hf}$ for samples in this study.

Table 1. Samples analyzed in this study grouped according to lithostratigraphic units described in figure 2. Additional information (e.g. Hf and Nd isotopic compositions determined in this study, lithology, structural context) is provided for each sample. Refer to figure 1 for sample locations.

Table 2. Hf isotopic results. The uncertainties on the $^{176}\text{Hf}/^{177}\text{Hf}$ values shown in this table are the within-run 2 sigma ($\times 10^{-6}$).

Table 3. Nd isotopic results. The uncertainties on the $^{143}\text{Nd}/^{144}\text{Nd}$ values shown in this table are the within-run 2 sigma ($\times 10^{-6}$).

Trace/REE & Major elements in Excel Workbook: “Chapter2TraceandMajorElement.xlsx”.

U-Pb Zircon results in Excel Workbook: “Chapter2_UPb_Suplemental.xlsx”.
Index

- Amphibolite - greenschist facies metasediments and metabasalts
- Greenschist facies metasediments and metabasalts
- Blueschist facies metasediments and metabasalts
- Papuan Ultrama/fic Belt; ophiolitic gabbrow and pillow basalt
- Alluvium; also beach sand
- Raised coral
- Colluvium; some alluvium
- Pilocene - recent volcanic rocks
- Pilocene intrusive igneous rocks
- Eocene intrusive igneous rocks
- Blueschist facies metasediments and metabasalts
- Core zone gneiss; eclogite facies metasediments and metabasalts
- Shear zone gneiss; eclogite facies metasediments and metabasalts
- Undifferentiated marine and clastic sedimentary rocks
- Limits of bathymetric high associated with PNG
- Spreading center

Louisiade Archipelago

Basalt & Gabbro
(1) 09 30a
(2) 09 31a
(3) 09 35a
(4) 09 35b
(5) 09 37a
(6) 09 45c

Calvados Schist
(1) 09 05b
(2) 09 37b
(3) 09 37c
(4) 09 45a
(5) 09 53a

Intermediate Magma
(1) 09 05a
(2) 09 05c
(3) 09 15a

D’Entrecasteaux & Misima Islands

Eclorige & Amphibolite
(1) 89 302a
(2) 08 058a
(3) 08 059b
(4) 08 069b

Basement Host Gneiss
(1) 89 301
(2) 89 303
(3) 89 327
(4) 03 092a
(5) 03 118b
(6) 04 148a
(7) 08 010g

Upper Plate
(1) 03 056
(2) 03 065
(3) 03 069a
(4) 04 186a
(5) ODP Leg 180

Quaternary Volcanic
(1) 89 322
D’Entrecasteaux & Misima Islands

Large MCCs with amphibolite facies to (U)HP lower plate.

Louisiade Archipelago

Tightly folded subgreenschist facies metapelites intruded by, and in fault contact with, mafic and intermediate magmatic rocks.

Rock types analyzed:
(with symbol used in rest of figures)

- eclogites & amphibolites from MCC lower plate
- felsic & intermediate gneisses from MCC lower plate
- basalt, gabbro, serpentinite from MCC upperlate; analyzed for comparison with eclogites & amphibolite
- Quaternary felsic lavas resting unconformably on upper and lower plate rocks; analyzed for comparison with MCC basement to determine source of magmatism (D’Entrecasteaux Islands only)
- intermediate magmatic rocks in fault contact with, and intruding, subgreenschist facies metapelites. Analyzed for comparison with high grade gneisses in western Woodlark Rift
- mafic magmatic rocks in fault contact with, and intruding, subgreenschist facies metapelites. Analyzed for comparison with eclogites and amphibolites
- subgreenschist facies metapelites

**Refer to figure 2 for detailed geologic map & sample locations**
**Refer to table 1 and section 2 for sample details and description of regional geology**
Neodymium Isotopes

Georgetown Inlier
Granitoids:
Poterozoic: $\varepsilon_{Nd} = -12 \text{ to } -25$
Siluro-Devonian: $\varepsilon_{Nd} = -11 \text{ to } -20$
Permian: $\varepsilon_{Nd} = -6 \text{ to } -15$
(Black adn McCulloch, 1990)

New England granitic suite & sediments
$\varepsilon_{Nd} = -20 \text{ to } -10$
(Hensel et al., 1985)

Whitsunday volcanics & granitoids (Ewart et al., 1992)

Lord Howe Rise: schist & graywacke
$\varepsilon_{Nd} = -11 \text{ to } -7$
(Mortimer et al., 2008)

Queensland Plateau: schist & sandstone
$\varepsilon_{Nd} = -10 \text{ to } -8$
(Mortimer et al., 2008)

Lord Howe Rise: schist & graywacke
$\varepsilon_{Nd} = -11 \text{ to } -7$
(Mortimer et al., 2008)

Queensland Plateau: schist & sandstone
$\varepsilon_{Nd} = -10 \text{ to } -8$
(Mortimer et al., 2008)

Neodymium & Hafnium Isotopes

Epsilon (\(\varepsilon\)) Hafnium vs Epsilon Neodymium

chapter 2; figure 4
Chapter 2; Figure 5a through 5d
Chapter 1; Figure 5e through 5g
Chapter 2; Figure 6a

Zircon Analysis by SIMS: Basement Host Gneiss

- Goodenough Island
  PNG 03118
  n = 14

- Misima Island
  PNG 04119
  n = 32

Chapter 2; Figure 6b

Zircon Analysis by LA-ICP-MS: Subgreenschist Facies Calvados Schist

- Louisiade Archipelago
  Panatinani Island
  PNG 0953a
  n = 37

- Louisiade Archipelago
  Wanim Island
  PNG 0906b
  n = 96
Chapter 2; Figure 7
Representative Whitsunday magmatic rocks (Ewart et al., 1992):
- A: basalt
- B: andesite
- C: andesite
- D: ignimbrite
- E: ignimbrite
- F: granodiorite
- G: granite

Chapter 2; Figure 8
Possibility 1:
Metamorphic rocks in S.E. PNG originated as an extension of the Early Cretaceous Whitsunday Volcanic Province in northeastern Australia

Possibility 2:
Metamorphic rocks in S.E. PNG originated as volcaniclastic sediments derived from the Early Cretaceous Whitsunday Volcanic Province

Possibility 3:
Metamorphic rocks in S.E. PNG originated as a Late Cretaceous volcanic province that was separate from Early Cretaceous volcanism in Australia

Legend
- Early Cretaceous Magmatic Province
- Basins Receiving Early Cretaceous Volcaniclastic Sediments
- Late Mesozoic & Cenozoic Ocean Basins
- Australian Continental Crust
- Direction of Sediment Transport

Chapter 2; Figure 9
Chapter 2; Figure 10
## Chapter 2, Table 1: Summary of samples from PNG

<table>
<thead>
<tr>
<th>Sample</th>
<th>Lat/Long</th>
<th>Lithology</th>
<th>Previous P-T-D constraints; Hf &amp; Nd Compositions</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>D’Entrecasteaux &amp; Misima Islands: Elogites &amp; Amphibolites</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>89302E</td>
<td>S 9.323611 E 150.275</td>
<td>Mafic eclogite Goodenough Isl.</td>
<td>Zr in rutile: 718˚ – 825˚ C; U-Pb zircon: 2.94 ± 0.41 Ma (Monteleone et al., 2007). εHf = +7.03; εNd = +2.82.</td>
</tr>
<tr>
<td>03118B</td>
<td>S 9.481301 E 150.250316</td>
<td>Mafic eclogite Goodenough Isl.</td>
<td>Zr in rutile: 677˚ – 817˚ C; U-Pb zircon: 2.09 ± 0.49 Ma; Jadeite barom: &gt; 14 kbar (Monteleone et al., 2007). εHf = +10.82; εNd = +5.15.</td>
</tr>
<tr>
<td>04148A</td>
<td>S 10.642464 E 152.542353</td>
<td>Amphibolite gneiss Misima Isl.</td>
<td>40Ar/39Ar complex amphibole spectrum: 11.3 Ma (unpub.). εHf = +9.58; εNd = +0.25. Lu-Hf garnet age = 11.2 ± 2.1 Ma (Zirakparvar et al., 2011).</td>
</tr>
<tr>
<td>08013A</td>
<td>S 9.383212 E 150.444813</td>
<td>Amphibolite gneiss Fergusson Isl.</td>
<td>εHf = +9.58; εNd = +4.16.</td>
</tr>
<tr>
<td><strong>D’Entrecasteaux Islands &amp; Misima Island: Felsic Gneiss</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>89327</td>
<td>S 9.650534 E 150.738250</td>
<td>Foliated granodiorite Guletatabatu</td>
<td>K/Ar potassium feldspar: 1.91 Ma (Baldwin et al., 1993) εHf = +4.34; εNd = +1.7.</td>
</tr>
<tr>
<td>03092E</td>
<td>S 9.464070 E 150.455992</td>
<td>Felsic gneiss Fergusson Isl.</td>
<td>Felsic host of sample 03092a. εHf = +3.6; εNd = +4.34.</td>
</tr>
<tr>
<td>08010G</td>
<td>S 9.487828 E 150.462938</td>
<td>Felsic gneiss Tomabaguna Isl.</td>
<td>εHf = -0.63; εNd = +3.31.</td>
</tr>
<tr>
<td><strong>D’Entrecasteaux Islands &amp; Misima Island: Intermediate Gneiss</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>89301</td>
<td>S 9.320120 E 150.27642</td>
<td>Felsic gneiss Goodenough Isl.</td>
<td>εHf = +7.42; εNd = +2.82.</td>
</tr>
<tr>
<td>89303</td>
<td>S 9.319203 E 150.274610</td>
<td>Felsic gneiss Goodenough Isl.</td>
<td>εHf = +5.51; εNd = +1.95.</td>
</tr>
<tr>
<td>04119A</td>
<td>S 10.687106 E 152.673239</td>
<td>Amphibolite gneiss Misima Isl.</td>
<td>40Ar/39Ar amphibole: 13.25 +/- 0.81 Ma (unpub.). εNd = +6.42. Hf isotopic data unavailable.</td>
</tr>
<tr>
<td><strong>D’Entrecasteaux Islands &amp; Misima Island: Upper Plate Rocks</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>04186A</td>
<td>S 10.619015 E 152.791931</td>
<td>Phyllite Misima Isl.</td>
<td>40Ar/39Ar amphibole: 9.73 Ma. εHf = +2.61; εNd = +0.8.</td>
</tr>
<tr>
<td>03056</td>
<td>S 10.035194 E 151.090987</td>
<td>Gabbro Normanby Isl.</td>
<td>εHf = +10.11; εNd = +5.43.</td>
</tr>
<tr>
<td>Code</td>
<td>Coordinates</td>
<td>Location</td>
<td>Hf and Nd Isotopic Data</td>
</tr>
<tr>
<td>----------</td>
<td>-------------------</td>
<td>-----------------------------------</td>
<td>-----------------------------------------------</td>
</tr>
<tr>
<td>03058A</td>
<td>S 10.034192 E 151.089816</td>
<td>Pillow basalt, Normanby Isl.</td>
<td>εHf = +8.48; εNd = +2.33.</td>
</tr>
<tr>
<td>03065</td>
<td>S 10.021400 E 151.040114</td>
<td>Olivine basalt, Normanby Isl.</td>
<td>εHf = +8.59; εNd = +4.43.</td>
</tr>
<tr>
<td>03069A</td>
<td>S 9.99461 E 151.014201</td>
<td>Serpentine, Normanby Isl.</td>
<td>No Hf of Nd isotopic data available.</td>
</tr>
<tr>
<td>ODP 180</td>
<td>S 9.507796 E 151.571628</td>
<td>Diabase, Moresby Seamount</td>
<td>εHf = +9.62; εNd = +3.75.</td>
</tr>
</tbody>
</table>

**D’Entrecasteaux Islands: Quaternary Calc-Alkaline Volcanic Rocks**

<table>
<thead>
<tr>
<th>Code</th>
<th>Coordinates</th>
<th>Rock Type</th>
<th>Other Information</th>
</tr>
</thead>
<tbody>
<tr>
<td>89322</td>
<td>S 9.636757 E 150.447834</td>
<td>Andesite</td>
<td>K-Ar whole rock age: 0.79 Ma (Baldwin et al., 1993) εHf = +9.65; εNd = +5.0</td>
</tr>
<tr>
<td>08058A</td>
<td>S 9.594528 E 150.953599</td>
<td>Rhyolitic-dacite</td>
<td>Fergusson Isl.</td>
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**Louisiade Archipelago: Basalt & Gabbro**

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<td>09030A</td>
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<td>Foliated Gabbro</td>
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<td>09031A</td>
<td>S 11 18.625' E 154 7.6571'</td>
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<td>S 11 18.467, E 154 3.501'</td>
<td>Gabbro</td>
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<tr>
<td>09035B</td>
<td>S 11 18.467, E 154 3.501'</td>
<td>Basalt</td>
<td>Basaltic dike intruding gabbro (09035a) εHf = +6.89; εNd = +2.97</td>
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<tr>
<td>09037A</td>
<td>S 11 18.528', E 154 3.482'</td>
<td>Basalt</td>
<td>Collected from basalt underlying metasediments (09037 b &amp; c); contact relationship unclear εHf = +3.78; εNd = +1.21</td>
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<tr>
<td>09045C</td>
<td>S 11 21.930', E 153 58.451'</td>
<td>Serpentinite</td>
<td>Collected from outcrop of serpentinitized ultramafic rocks; in fault contact with 09045A. Hf and Nd isotopic data unavailable</td>
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**Louisiade Archipelago: Intermediate Magmatic Rocks**

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<td>Pyroxene Diorite</td>
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Chapter 2; Table 2
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Chapter 2; Table 3
Zircon growth in rapidly evolving plate boundary zones: Evidence from the active Woodlark Rift of Papua New Guinea

N.A. Zirakparvar\textsuperscript{a*}

S.L. Baldwin\textsuperscript{a}

J.D. Vervoort\textsuperscript{b}

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\textsuperscript{b}Washington State University, School of Earth and Environmental Science, Pullman WA 99164, USA

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Abstract

Conventional analysis of accessory minerals in ‘spot-mode’ in polished sections samples crystal volumes, that even in high-sensitivity secondary ionization mass spectrometry (SIMS), have diameters of ~10-20 µm. These ‘spot mode’ analyses can reveal the ages and trace element compositions of crystal domains if the domains are larger than the lateral beam dimensions. They cannot, however, be used to characterize variations that occur over smaller dimension, or that occur at sharp boundaries between discrete domains often observed in cathodoluminescence (CL) images. Because of this spatial limitation, conventional spot analysis cannot be used to investigate processes operative at scales <10 µm, and the relationship between the U-Pb isotopic and trace element records in the same crystal volume.

In this study SIMS depth-profiling was used to characterize zircons from felsic and intermediate gneisses that host the world’s youngest (U)HP eclogites in southeastern Papua New Guinea. Many zircon CL images show cores overgrown by dark CL rims. SIMS depth profiling perpendicular to unpolished zircon surfaces, and penetrating to a depth of ~15 µm, effectively measured the U-Pb isotopic and trace element composition at a depth resolution of <1 µm. Crystal domains up-to and across the sharp transition from overgrowth into inherited core were investigated. These depth profiles are augmented by U-Pb data collected in ‘spot mode’ on 1) polished cross sections, and 2) external (i.e., unpolished) surfaces. The U-Pb results reveal zircon grew in the Pliocene as individual crystals, and as overgrowths on Late Jurassic to Cretaceous cores. Zircons lacking Cretaceous cores
and with uniformly Pliocene ages also displayed complex internal zoning patterns under CL. The Pliocene zircons are within error of $^{40}\text{Ar}/^{39}\text{Ar}$ mineral ages on the (U)HP host gneisses and indicate zircon crystallization during (U)HP exhumation. The U-Pb zircon systematics in the host gneisses provide information about the protolith age followed by zircon growth at temperatures of ~590 to 690 °C based on Ti thermometry. Small scale geochemical disequilibrium within the host gneisses at the time of zircon crystallization is indicated by contrasting trace element compositions for different zircons crystallizing in the same rock, under the same geologic conditions, and at the same time.
I. Introduction

The recognition and quantification of micron scale chemical and isotopic variations in geologic materials has revolutionized our understanding of processes operating at the scale of individual minerals to the entire solar system (e.g., Ireland and Wlotzka, 1992). The mineral zircon, in particular, has been shown to preserve isotope and trace element variations during growth and recrystallization, that shed light on processes such as the development of plate tectonics on the Early Earth (Harrison and Schmitt, 2007; Trail et al., 2007a, 2007b; Harrison et al., 2005; Watson and Harrison, 2005), magma evolution (e.g. Schmitt et al., 2010a, 2010b, Reid et al., 2011), and UHP metamorphism (Ayers et al., 2002; Walsh et al., 2007; Monteleone et al., 2007; Baldwin et al, 2008).

During (U)HP metamorphism zircon growth can occur as rims on pre-existing zircons as well as by nucleation/crystallization of zircon (Gao et al., 2011; Liati, 2005; Mccelland et al., 2009). U-Pb ages, Ti concentrations, and Hf isotopic compositions determined via SIMS analysis on grains either separated from the host rock or analyzed in situ have been reported (Ireland and Williams, 2003; Harrison et al., 2008; Kinny and Maas, 2003). Typically SIMS data is collected in ‘spot-mode’, over relatively short durations (e.g. ~15 minutes), and sampling small volumes (<1 µm pit depth with diameter of 30 µm) of the zircon. Despite the fact that these analyses are of high spatial resolution and provide information on the ages and trace element compositions of discrete domains within zircon grains, they do not necessarily reveal the chemical and isotopic characteristics of the sharp boundaries between these discreet domains, often observed in Cathode-Ray Luminescence (CL).
images. Because of this, important questions pertaining to the behavior of natural zircon in response to changing conditions in the host, as well as the relationship between the U-Pb isotopic and trace element records of tectonic processes remain unanswered.

To begin to address these questions requires a sampling and measurement technique capable of performing isotopic and trace element analyses up to and across the sharp boundaries separating different aged domains within single zircon crystals. This paper presents the results of an investigation using the SIMS technique to characterize zircons extracted from felsic and intermediate gneisses (fig. 1a) that host the world’s youngest known ultra-high pressure (U)HP eclogites, in southeastern Papua New Guinea (Baldwin et al., 2008). The gneisses were targeted specifically to gain insight into zircon growth relative to (U)HP metamorphism in an area where the timing of tectonic events is relatively well understood. Many of the zircons exhibited internal morphologies consistent with older cores overgrown by younger rims (fig. 1b). The SIMS technique was used to conduct depth profiles of zircons (Schneider et al., 2011; Zhou et al., 2010; Gordon et al., 2009a, 2009b; Trail et al., 2007a; Breeding et al., 2004;) from these gneisses, beginning at the external unpolished grain surface and penetrating to a depth of ~15 µm- effectively measuring the U-Pb isotopic and trace element (Ti, Hf, and Y) composition up-to and across the sharp transition from young zircon overgrowth into the inherited zircon core. Ti, Hf, and Y were chosen because characterizing the behavior of these elements in natural samples under known geologic conditions is important for the interpretation of zircon crystallization temperatures calculated on
the basis of Ti concentration in zircon, and the use of the Lu-Hf isotopic system in zircon.

The zircon U-Pb age and trace element depth profiles are augmented by two other types of SIMS analysis (fig. 1c): 1) U-Pb dating in ‘spot mode’ of polished cross sections of zircon grains, and 2) U-Pb dating in ‘spot mode’ of the external unadulterated surfaces of the zircon grains. When the data from these three types of SIMS analysis is combined with pre-existing constraints for the tectonic evolution of the region, it is possible to directly assess processes responsible for zircon growth in gneisses hosting (U)HP rocks as revealed by the U-Pb isotope and trace element data. These results also provide insight into the spatial-temporal scale and extent of chemical disequilibrium in rocks that are part of a former subduction complex now being rifted apart.

II. Geological Background & Samples Analyzed

A. Geological Overview

The three samples examined in this study are from the core zones (i.e., lower plate) and shear zone carapace of metamorphic core complexes (MCC) in the D’Entrecasteaux Islands, southeastern Papua New Guinea (fig. 1a). The D’Entrecasteaux islands occur in a zone of active extension (i.e., Woodlark Rift) at the western apex of the Woodlark Basin. The oldest seafloor in the Woodlark Basin is c.a. 6 Ma, but extension probably started at c.a. 8.4 Ma based on a regional unconformity (Taylor et al. 1995). The Woodlark Basin seafloor spreading center is actively propagating westward (Taylor et al. 1995, 1999). In the D’Entrecasteaux Islands, Cretaceous volcanlastic sediments and basalts derived from the
Gondwana rifted margin (Zirakparvar et al., to be submitted to Gondwana Research) were subducted during the late Mesozoic, subsequently metamorphosed at (U)HP conditions, (Monteleone et al., 2007; Baldwin et al., 2008; Zirakparvar et al., 2011), and are now being exhumed to the surface during rifting (e.g. Little et al., 2007; 2011; Webb et al., 2008).

Remnants of subducted lithosphere are found in the core zones of large MCCs, where felsic and intermediate gneisses encapsulate mafic eclogites with Miocene–Pliocene crystallization ages. U-Pb zircon and Lu-Hf garnet dating of the mafic eclogites (Monteleone et al., 2007; Zirakparvar et al., 2011), and P-T constraints (Baldwin et al., 2004; 2008), indicate that the mafic eclogites in the D'Entrecasteaux Islands were exhumed from depths of at least 90 km since ~8 Ma. K/Ar, $^{40}$Ar/$^{39}$Ar, and fission track dating techniques applied to the lower plate rocks have also documented an extremely rapid (e.g., ≥ 100º C/ m.y.) cooling history culminating within the past five million years (Baldwin et al., 1993). In the D'Entrecasteaux Islands, seismic activity (Abers et al., 2002), stream profile analysis (Miller et al., 2011), and Plio-Pleistocene $^{40}$Ar/$^{39}$Ar mineral cooling ages, all suggest that exhumation of lower-plate rocks occurred during Plio-Pleistocene to Holocene time and may still be on-going (Baldwin et al., 1993).

B. Samples for SIMS Analysis

Three samples from the western Woodlark Rift were selected for SIMS zircon analysis (fig. 1a). Two of these samples are from the MCC core zone and are the hosts of Miocene–Pliocene mafic eclogites. The third sample is from the Wakonai Shear Zone separating MCC core zone rocks from the unmetamorphosed upper plate
exposed on northern Goodenough Island (Hill, 1994; Baldwin et al., 1993; Little et al., 2011).

1. PNG 03118m

Sample 03118m is a quartz-feldspathic gneiss from southeastern Goodenough Island (Figure 1a). At outcrop scale, this gneiss encapsulates a mafic eclogite that has been previously examined by Monteleone et al (2007) who determined a 2.09 +/- 0.49 Ma (MSWD = 3.3) in-situ U-Pb zircon SIMS age. Trace and REE data was used to interpret this U-Pb age as the time of eclogite facies metamorphism. Zirconium in rutile saturation thermometry for this eclogite gave temperatures ranging from 677 to 817 °C, and jadeite barometry gave pressure of at least 14 kbar (Monteleone et al., 2007). This mafic eclogite is found as a meter scale lenticular boudin within sample 03118m, and unlike its host gneiss, preserves the peak metamorphic assemblage of garnet + omphacite + phengite. ⁴⁰Ar/³⁹Ar apparent ages for minerals extracted from another exposure of quartzo-feldspathic gneiss on Goodenough Island, ~15 km to the north of sample 03118m, range from 2.71 to 3.00 Ma for hornblende, 1.56 to 1.61 Ma for biotite, and 2.29 to 4.17 for plagioclase feldspar (Baldwin et al., 1993). Zircon in this sample ranges in length from 100 to 200 µm, are prismatic to acicular, and exhibit an oscillatory zoned metamict core that is overgrown by a dark CL rim (fig. 1b). Rim thickness varies and ranges from optically invisible up to 5 µm in width.

2. PNG 08010g

Sample 08010g is a quartz-feldspathic gneiss from Tumabaguna Island, offshore of western Fergusson Island (Figure 1a). It is the location of the world’s
youngest known coesite eclogite (e.g. Baldwin et al., 2008; Zirakparvar et al., 2011) and is composed of quartzo-feldspathic gneiss that encapsulate Late Miocene eclogite boudins and dikes. These gneisses are also intruded by concordant and discordant bodies of granodiorite and pegmatite (Little et al., 2011; Brownlee et al., 2011).

Sample 08010g is the felsic host at the locality of the world’s youngest known coesite eclogite (Baldwin et al., 2008). The coesite eclogite has been dated to 9.83 Ma by Baldwin et al (1993) using the $^{40}$Ar/$^{39}$Ar technique for amphibole from the retrogressed rind of the coesite eclogite, 7.0 ± 1.0 Ma by Monteleone et al (2007) using in situ SIMS U-Pb zircon analysis, and 7.09 ± 0.9 Ma by Zirakparvar et al (2011) using the Lu-Hf garnet system. The most recent interpretation by Zirakparvar et al (2011) for the origin of the coesite eclogite is that it crystallized from a basaltic partial melt of the upper mantle that was intruded into formerly subducted continental lithosphere where it crystallized at (U)HP conditions. This is interpreted to have occurred during the earliest phases of lithospheric rupture due to the onset rifting in the region. A 3.52 Ma $^{40}$Ar/$^{39}$Ar age for white mica from a pegmatite in the same outcrop as sample 08010g has also been reported (Baldwin et al., 1993).

Zircons in this sample are prismatic and acicular, in some cases exceeding ~400 µm x ~170 µm. All of the zircons in this sample exhibit complex internal zoning (fig. 1b). Some of the grains have convoluted metamict cores surrounded by concentrically zoned regions, and others exhibit a completely chaotic internal morphology lacking any discernable patterning.
3. PNG 0621a

Sample 0621a is from an outcrop of layered quartzo-feldspathic gneiss in the Wakonai Shear Zone (figure 1a), northern Goodenough Island (e.g. Baldwin et al., 1993; Little et al., 2011). $^{40}\text{Ar}/^{39}\text{Ar}$ mineral ages for a sample locality in the same general area as 0621a are 1.53 Ma for white mica, 1.41 Ma for biotite, and 1.42 Ma for K-feldspar (Baldwin et al., 1993). Apatite fission-track dating for rocks in the Wakonai shear zone produces ages that cluster around 0.8 Ma (Baldwin et al. 1993). These young $^{40}\text{Ar}/^{39}\text{Ar}$ and apatite fission track ages indicate that this shear zone has been active in recent times and is intrinsically related to the exhumation of (U)HP metamorphic rocks in the western Woodlark Basin. However, the Wakonai Shear Zone also preserves a record of the early metamorphic history of the rocks now exposed in the D'Entrecasteaux Islands. Zirakparvar et al. (2011) determined a ~68 Ma Lu-Hf age for a centimeter sized relict garnet porphyroblast in an amphibolite gneiss, also from the same general area as sample 0621a. This older age was interpreted as recording garnet growth in response to initial continental subduction and ophiolite obduction in the region north and east of Australia following the breakup of Gondwana. The presence of relict ~68 Ma garnet porphyroblasts in the Pleistocene/Pliocene shear zone provides evidence for disequilibrium and incomplete resetting of isotopic systematics in the shear zone.

This sample contains both prismatic zircons with lengths of ~60 µm to ~200 µm and subrounded zircons with diameters of~30 µm to ~100 µm. Most of the grains exhibit simple to complex oscillatory magmatic zoning. A thin (e.g. <2 µm) dark CL overgrowth is occasionally present on grains from this sample (fig. 1b).
III. Secondary Ionization Mass Spectrometry (SIMS) Analytical Methods

Ion microprobe measurements were made using the Cameca ims 1270 high resolution, high sensitivity ion microprobe housed at the University of California, Los Angeles. Resolution of mass interferences within the mass range analyzed was possible due to the high mass resolution (~4500). The data reported in this paper was collected during three sessions on the ims 1270 (August of 2008, July of 2010, and May of 2011). The first session consisted of analyses of polished zircon cross sections performed in spot mode (fig 1c). The second analytical session consisted of analyses, also in spot mode, targeting unpolished zircon grain surfaces (fig. 1c). The final analytical session consisted of analyses in depth profiling mode, starting at the exterior surface of unpolished zircon grains (fig. 1c). Only U-Pb analysis was performed during the sessions conducted in spot mode, whereas Ti, Hf, and Y concentrations were also measured during the depth profiling.

Zircons were prepared for SIMS analyses in two ways. For analysis of the polished cross sections, grains were mounted in epoxy with the zircon standard AS3. These mounts were polished, exposing the internal surfaces of the grains, prior to the application of a thin coating of carbon for SEM cathode-ray luminescence (CL) imaging. Prior to SIMS analysis the carbon coating was removed through light polishing, the mount was cleaned with dilute HCL, and finally coated with a ~30 nm Au film. For analyses on unpolished grain surfaces, in both spot and depth profiling modes, zircons were pressed into indium metal along with AS3 zircon standard.
Mounts were cleaned with dilute HCL and coated with a ~30 nm Au film. The analytical parameters for two types of SIMS analyses are described below.

A. Spot Mode

A 12.5 kV primary 160- beam with a ~20 nA current and ~25 µm beam diameter were used for ablation of sample material. Intensities of monatomic U+, Th+, and Pb+ ions and 94Zr2O+ and UO+ molecular ions were measured with a discrete dynode electron multiplier in peak jumping mode. Individual analyses consisted of 15 cycles with 15s count times. O₂ flooding at 3 x 10⁻⁵ Torr was applied to the sample surface to enhance Pb yield. In-house software (ZIPS) was used to reduce the raw data and export isotopic ratios used in age calculations.

B. Depth Profiling

A 12.5 kV primary 160- beam with a ~20 nA current and ~25 µm beam diameter were used for ablation of sample material. Intensities of monatomic U+, Th+, Pb+, Ti+, and Y+ ions and 94Zr2O+, UO+, HfO+ molecular ions were measured with a discrete dynode electron multiplier in peak jumping mode. Each depth profile consisted of 100 cycles with 15s count times. O₂ flooding at 3 x 10⁻⁵ Torr was applied to the sample surface to enhance Pb yield. In-house software (ZIPS) was used to reduce the raw data and export isotopic ratios and elemental intensitites used in age and trace element concentration calculations. Pit depths from the depth profiles were estimated using an optical interferometer at the UCLA.

IV. Results
The $^{238}\text{U}/^{206}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ isotopic ratios of the zircons, uncorrected for common lead, are regressed individually on Terra-Wasserburg concordia diagrams using a fixed upper intercept value of $^{207}\text{Pb}/^{206}\text{Pb}$ of 0.8283 ± 0.05 (fig. 1d) (Sanudo-Wilhelmy & Flegal, 1994). The lower intercept of the regression line between the fixed $^{207}\text{Pb}/^{206}\text{Pb}$ value and the $^{238}\text{U}/^{206}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ratios of the unknowns yield T.W. model ages (fig. 1d).

The significantly longer duration of the depth profile (100 cycles) as compared to the spot mode analysis (15 cycles) necessitates a different approach to correcting raw data for mass fractionation. Our approach for the depth profiles was to generate four separate depth dependent mass fractionation factors (e.g. separate mass fractionation corrections for cycles one through twenty-five, twenty-six through fifty, fifty-one through seventy-five, and seventy-six through one-hundred). This is in contrast to the single mass fractionation correction applied to all fifteen cycles making up the spot mode analysis. For final calculation of isotopic ratios from the depth profiles, individual cycles in each of the twenty-five cycle mass fractionation groups were then subdivided into blocks consisting of five cycles each. The result is that data from the depth profiles is reported as a series of twenty blocks each consisting of five cycles, whereas the data from the spot mode simply consists of all fifteen cycles from that analysis. All of the mass fractionation corrections are derived relative to zircon standard AS3.

In order to calculate trace element concentrations (Ti, Hf, and Y) for the depth profile analyses, individual cycles from each depth profile were grouped into blocks of five cycles, creating twenty blocks for each depth profile. This is to
facilitate direct comparison between the trace element and U-Pb isotopic data for each depth profile. Instead of using zircon standard AS3 for trace element calculations, standard 91500 was used. Data from a depth profile conducted on the 91500 standard was also broken down into twenty blocks, each consisting of five cycles. Relative sensitivity factors (RSF) for each of these five cycle blocks for the elements of interest were calculated by dividing the measured intensities from the standard by the known concentration of the element in the standard (Ti = 4.5 ± 0.4 ppm; Hf = 6250.0 ± 27.2 ppm; Y = 136 ± 2 ppm; from Yong Sheng et al., 2010). This effectively generated twenty RSF factors as a function of depth that were then used to calculate concentration values for each of the corresponding five cycle blocks from the unknowns. Ti concentrations are used to calculate crystallization temperatures based on the formula: \( T_{(°C)} = \frac{-1}{((\log(T_{conc}) - 6.01)/5080))} - 272.15 \) (Watson and Harrison, 2005).

B. Sample Results

Results of the spot mode and depth profile U-Pb analyses are found in the electronic appendix. The Ti, Hf, and Y concentrations from the depth profiles are reported in table 3. Data from the depth profiles are grouped in 5 cycle blocks, for a total of 20 blocks per depth profile (see methods section).

1. PNG03-118m

Fifteen analyses in spot-mode of the polished zircon cross sections, twenty-six analyses in spot-mode of the unpolished zircon surfaces, and four depth profiles were acquired for 03118m (fig. 2a). The range of T.W. model ages for the analyses in spot mode was ~44 to ~110 Ma, with the bulk of ages falling between 80 and 110
Ma. Spot mode analyses on the unpolished external surfaces of the zircon grains resulted in a range of T.W. model ages between 1.1 and 3.8 Ma, with three of the twenty-six analyses falling between 72 and 95 Ma (fig. 2a).

The four depth profiles conducted for this sample are all characterized by initial ages between 1 and 4 Ma, followed by data that indicates older ages (fig. 2a). These older ages fall within the ~90 to ~130 Ma age range from the spot mode analyses on zircon cores. The two depth profiles with the smallest initial zone (profiles three and five with age breaks at blocks six and seven, respectively) of ages between 1 and 4 Ma exhibit a sharp break in age as the depth profile passes from young overgrowth into older core, whereas the two profiles with the largest initial zone (profiles one and two with age breaks at blocks fifteen and eleven, respectively) exhibit a more gradual increase in age as the depth profile passes out of the young rim and into the older parts of the zircon.

For depth profiles 1, 3, and 5, the concentration of Ti decreases steeply and steadily, starting at the unpolished zircon surface until the depth profile reaches the interface between the young overgrowth and the older core (fig. 3). In depth profile 2, the Ti concentration essentially remains unchanged throughout the young overgrowth. Zircon crystallization temperatures for the portions of the depth profiles occurring within the young zircon overgrowths, calculated using the observed Ti concentrations, vary between ~665 and ~590 °C, consistent with previously determined Ti in zircon temperature estimates (e.g. Monteleone et al., 2007).
For depth profiles 1, 2, and 3, the concentration of Hf does not change abruptly at the boundary between the young overgrowth and the older core, whereas in profile 5, the concentration of Hf suddenly begins to increase once the depth profile passes out of the young overgrowth. Hf concentrations for the young overgrowths range from \(~1.8\) wt \% to \(1.4\) wt \%.

In profiles 1, 3, and 5, the concentration of Y remains constant and is \(< 200\) ppm for all of the blocks occurring within the young zircon overgrowth. Once the profiles pass out of the young overgrowths into the older cores, the concentration of Y begins to rise- quite steeply for profiles 3 and 5, eventually attaining values \(> 3,000\) ppm. In profile 2, the concentration of Y at the start of the depth profile is \(~1500\) ppm and then steadily declines over the region of the young overgrowth to \(~500\) ppm at the interface between the young overgrowth and the older core. The concentration then begins to steadily rise, but not as sharply as in profiles 3 and 5.

2. PNG08-10g

Thirteen analyses of the polished zircon cross sections and twenty-five analyses of the unpolished zircon surfaces were obtained (fig. 2b). Three depth profiles were also performed. Despite the complex internal morphology of the zircon grains from this sample, all of the SIMS analyses produce T.W. model ages that are within error of each other and fall between 2.5 and 6 Ma, with most of the analyses falling between 3.5 and 4.5 Ma (fig. 2b).

Ti, Hf, and Y are not homogeneously distributed in zircons from this sample (fig. 4). Ti saturation thermometry for depth profiles two and three indicate temperatures between \(650^\circ\)C and \(700^\circ\)C degrees, whereas depth profile one
exhibits a sudden increase from ~600°C degrees in blocks one through eleven up to >1200°C degrees for blocks fifteen through twenty. The Hf concentration of all the depth profiles from 08010g are similar to one another, but are also not homogeneous throughout the individual profiles. Hf gradually increases during the first five blocks of the profiles, roughly plateaus between blocks six and ten, and then after block ten decreases steadily in profiles one and three but increases in profile two. The concentration of Y is homogeneously distributed in depth profile three, but not in profiles one and two. For the first eleven and five blocks profiles one and two, respectively, exhibit a Y concentration similar to that of profile three (~450 - 600 ppm), however, the concentration of Y in profiles one and two then abruptly increases to values that are >3500 ppm.

3. PNG06-21a

Thirty analyses of the polished zircon cross sections were performed. Two depth profiles were also acquired for zircons in this sample. No spot mode analyses were acquired on the unpolished zircon surfaces. The T.W. model ages for the polished zircon cross sections from 0621a range from 3.2 Ma to 1081 Ma, with most analyses falling between 3 and 5 Ma, and 40 and 120 Ma (fig. 2c). Only a few grains in this sample yielded ages older than 120 Ma, and for the purpose of comparison between the SIMS analyses in spot-mode and the depth profiles from this sample, ages older than 220 Ma are excluded from figure 2c. The first four blocks of the first depth profile yielded T.W. model ages between 3 and 5 Ma. This initial zone of young ages is then followed by a gradual increase in T.W. model ages up to ~70 Ma, where the ages plateau. Depth profile two also has an initial zone that yielded ages
between 3 and 5 Ma, but this zone only consists of two blocks. Following this small initial zone of young ages, there is a rapid increase in age up to ~150 Ma, where there is a plateau.

Ti saturation thermometry for the first depth profile indicates temperatures of ~620°C degrees for the first three blocks of the profile, followed by a sharp increase to a temperature of 670°C for block five, and then a decrease to 650°C for the rest of the profile (fig. 5). The first block of depth profile two yields a temperature of ~640°C. This is followed by a sharp increase over the course of blocks two and three to ~680°C. This sharp rise is then followed by a sharp decline to ~650°C for the remainder of the blocks in profile two.

The first four blocks of depth profile one exhibit Hf concentrations between 1.8 and 1.9 wt %, followed by a sharp decline to ~1.6 wt % for the remainder of the profile. The first block of depth profile two has a Hf concentration of 1.7 wt %, followed by an increase to ~1.9 wt % for block two. Following block two, the concentration of Hf measured in depth profile two steadily declines to ~1.1 wt % where it remains until the end of the profile. The Y concentration in the first two blocks are ~800 and ~600 ppm, respectively, for both depth profiles from 0621a. In depth profile one, concentration of Y continues to decrease after block two to ~500 ppm for blocks three and four. There is then a gradual rise in Y between blocks five and eleven up to ~600 ppm. After block eleven, the concentration of Y rises more sharply to values between 1100 and 1300 ppm for the remainder of profile 1. In depth profile two, the concentration of Y increases after block two to ~1000 ppm. Thereafter, the concentration steadily decreases till it reaches ~600 ppm at block
fourteen. This is then followed by a gradual increase, eventually reaching ~700 ppm, for the remaining six blocks in profile 2.

V. Discussion

The SIMS zircon U-Pb age and trace element measurements for three felsic and intermediate gneisses, two of which host (U)HP eclogites, from the D'Entrecasteaux Islands allow for a robust assessment of the geologic conditions recorded by the U-Pb ages, provide insights into the behavior of Ti, Hf, and Y in zircons from rocks in a transient plate boundary zone with a known tectonic history, and allow for an assessment of coupling between the U-Pb isotopic and trace element records of tectonic processes in natural samples.

A. Geologic Conditions Recorded by U-Pb Zircon Ages

Cathode Ray Luminescence images of the polished zircon cross sections from two of the samples examined in this study reveal an internal structure consistent with new metamorphic zircon that has grown onto a pre-existing grain (samples 03118m and 0621a; fig. 1b). This young zircon appears as a dark CL rims of variable thickness, and the combined results of SIMS spot-mode and depth-profile analysis confirm that these overgrowths are much younger than the zircon they mantle. In samples 03118m and 0621a, the inherited parts of the zircon grains mostly yield ages spanning the mid to late Mesozoic, with a few early Cenozoic, Paleozoic, and Precambrian ages (figs. 2a & 2c). It is not possible to directly address the geologic conditions surrounding zircon growth in the inherited portions of these zircon grains. It is possible, however, to directly assess the conditions during formation of
the metamorphic zircon overgrowths on these inherited grains, as well as the conditions surrounding the crystallization of zircons in sample 08010g (fig. 2b).

The most important observation in understanding the geologic conditions at which these young zircons crystallized is that $^{40}\text{Ar}/^{39}\text{Ar}$ apparent ages from Goodenough and Fergusson Islands are within error of the U-Pb ages from the metamorphic rims in samples 03118m and 0621a and all of zircons in 08010g (Baldwin et al., 1993). The data documents metamorphic zircon growth during rapid exhumation from (U)HP conditions. Depth profiles for the overgrowths in samples 03118m and 0621a produce uniform ages (figs. 2a & 2c), and corresponding U-Pb ages abruptly increase as the ion beam samples the young rim and then passes into the older parts of the grain. The number of analytical blocks corresponding to the limit of the young rim and the time when the ages plateau varies from sample to sample, but this is probably a function of mixing between the young overgrowth and the older core during ion beam sampling, and we do not believe this reflects an age gradient between the young overgrowth and old core in these grains.

Lu-Hf geochronology by Zirakparvar et al. (2011) supports the existing idea that the metamorphic rocks in the D'Entrecasteaux Islands record metamorphism during an episode of Late Mesozoic/Early Cenozoic arc-continent collision involving subduction of the rifted Australian margin and ophiolite obduction (Lus et al., 2004). The Cretaceous ages of the inherited zircon cores in samples 03118m and 0621a indicates that the protoliths of these two gneisses are old enough to have been involved in this Late Mesozoic subduction event- meaning that these two samples
have experienced a full range of plate boundary processes from deposition on a rifted margin in the Cretaceous, to subduction in the Late Mesozoic, to exhumation during the Late Miocene. However, the U-Pb ages of zircons in samples 03118m and 0621a only record two geological ‘events’: the age of the protolith (e.g. the Cretaceous zircon cores), and the timing of final crystallization (e.g. the Late Miocene zircon overgrowths) corresponding to the time of $^{40}\text{Ar}/^{39}\text{Ar}$ bulk closure of amphiboles and muscovites from these rocks.

Samples 03118m and 0621a do not record a U-Pb zircon isotopic record of any geologic events between the protolith age and the time when these rocks were exhumed during Late Miocene rifting. An important note is that there are some ages from the spot-mode analyses of the polished zircon cross sections from samples 03118m and 0621a that fall between the age of the young rims and the ages of the inherited zircon cores. However, our interpretation based on the depth profiling results is that these intermediate ages are due to mixing between the young rim and old core during microbeam sampling and are thus analytical artifacts. It does not appear that metamorphic zircon grew in these samples during (U)HP metamorphism (e.g. ~7 Ma; Baldwin et al., 2008; Zirakparvar et al., 2011).

Zircons in sample 08010g appear to have uniformly crystallized at ~4 Ma and there is no inherited zircon in this rock, suggesting that this gneiss originated as a partial melt or that the availability of Zr was limited until this time (fig. 2b). The ~4 Ma zircon U-Pb ages from this gneiss are within error of a $^{40}\text{Ar}/^{39}\text{Ar}$ cooling age from a granitic pegmatite in the same outcrop, but are younger that the ~8 Ma crystallization age of the coesite eclogite (Monteleone et al., 2007; Baldwin et al.,
2008; Zirakparvar et al., 2011) present as a boudin within this gneiss. The ~4 Ma U-Pb age for zircon in sample 08010g provides no information about the timing of (U)HP facies metamorphism. Instead, and similarly to the young zircon overgrowths in samples 03118m and 0621a, the U-Pb systematics in sample 08010g appear to record crystallization during exhumation from (U)HP conditions.

In other (U)HP terranes, there are still unanswered questions as to whether U-Pb zircon ages from felsic and intermediate gneisses present in association with (U)HP metamorphic rocks provide any information about the timing of prograde and peak (U)HP metamorphism (e.g. Harley and Carswell, 1995; Menold et al., 2009). Based on these results zircon U-Pb ages from the felsic and intermediate gneisses found in association with (U)HP metamorphic rocks appear to reveal the protolith age and ages interpreted as zircon growth during exhumation. However, the U-Pb data set from the felsic and intermediate gneisses we examined does not reveal the time of initial subduction or (U)HP metamorphism.

B. Trace element variability recorded in depth profiles

The depth-dependant progression of U-Pb ages determined for the depth profiles conducted on zircons from samples 03118m and 0621a reveal sharp boundaries between young overgrowths and older cores. In all of the grains examined, this interface is also characterized by a change in the concentration of Ti, Hf, and Y (Figs. 3 & 5). In some instances the concentration change at the interface is significant, whereas in others it is barely observable. Another important observation is that the concentrations of Ti, Hf, and Y vary significantly throughout some of the zircon overgrowths, whereas in others it remains nearly homogeneous.
Numerous studies have shown that cation diffusion proceeds slowly in zircon (e.g., Cherniak and Watson, 2003). Diffusion, by its very nature, is a kinetic process seeking to attain chemical equilibrium, and as such would be expected to manifest itself in the depth profiles as a gradual change in concentration across the sharp boundary between the young zircon overgrowth and old zircon core. The fact that the depth profiles for samples 03118m and 0621a are characterized by abrupt changes in Ti, Hf, and Y concentrations corresponding to the location of the zircon age domain boundaries, strongly suggests that a significant amount of diffusion between the young overgrowths and the older cores has not occurred.

Similar concentrations of Ti, Hf, and Y would not be expected for all of the zircon cores in samples 03118m and 0621a because the spectrum of U-Pb ages for these cores indicates inheritance from multiple sources. It does appear on the basis of U-Pb age, however, that the young overgrowths in samples 03118m and 0621a, and the entirety of the zircon grains in sample 08010g crystallized contemporaneously within their respective samples. On this basis, it would not be unreasonable to predict identical behavior of Ti, Hf, or Y associated with the youngest zircon domains (e.g. young rims in 03118m and 0621a, and all of zircon in 08010g) within a particular sample. However, considerable trace element variability is observed.

There are three possible explanations to account for the trace element variability across different zircons of the same age from a given sample: modification due to the diffusion, progressive enrichment or depletion of the element at the zircon-matrix interface over the interval of crystallization, or within-
sample chemical heterogeneities present during crystallization. Because age-domain boundaries (e.g. the interface between the Late Miocene overgrowth and the Cretaceous inherited core) are all associated with distinct changes in trace element concentration, it is our opinion that diffusion can be ruled out. Modification of trace element concentrations via diffusion would be expected to ‘smooth’ out this abrupt concentration change. Progressive enrichment or depletion at the zircon-matrix interface would be expected to result in an increase or decrease in the concentration of the element throughout the age domain, but this behavior would be observed in all of the zircons from a particular sample if it was in equilibrium at the time of zircon crystallization. The most plausible explanation for the observed trace element variations is that these rocks were not in internal chemical equilibrium for these elements at the time of metamorphic zircon crystallization.

C. Trace element record in zircon and what it reveals about tectonic processes

Comparison of the U-Pb results with the trace element concentrations in the depth profiles reveals an important observation: the U-Pb ages for a given sample are internally consistent (in the case of the young rims in 03118m and 0621a, and all of zircon in 08010g) whereas the trace element concentrations are highly variable. This variation probably reflects hand sample scale geochemical heterogeneities present at the time the zircon overgrowths formed in 03118m and 0621a, and all of zircons in 08010g crystallized and suggests samples did not achieve chemical equilibrium at the time of zircon crystallization. If true, though, this potentially complicates the use of trace element concentrations in zircons from
bulk separates (e.g. the grains are not analyzed in situ) extracted from gneissic rocks that are not in internal equilibrium.

1. Ti in Zircon

Thermometers based on the Ti content of zircon have been widely applied to igneous and metamorphic rocks (e.g. Ferry and Watson, 2007; Fu et al, 2008 ). These thermometers are appealing because they can be used on individual zircons that have been removed from their host rock by artificial (e.g. laboratory separation of zircon) or natural (e.g. detrital zircons) means. These thermometers rely on thermodynamic models for the incorporation of Ti into zircon that have been calibrated in both experimental settings and in natural samples where the temperature of crystallization is already known (e.g. Watson et al., 2006). Since the degree to which Ti is incorporated into zircon depends on the activities of SiO$_2$ ($a_{SiO_2}$) and TiO$_2$ ($a_{TiO_2}$), and temperature, ideal applications of Ti in zircon thermometers would be in situations where $a_{SiO_2}$ and $a_{TiO_2}$ can be measured in conjunction with the concentration of Ti in zircon (Ferry and Watson, 2007). It is quite common, however, for these thermometers to be applied in cases where there is little or no information for $a_{SiO_2}$ and $a_{TiO_2}$. In cases where rutile and quartz are present is common to assume an activity of ‘1’ for these species.

Examination of the results of Ti thermometry calculations for the depth profiles from 03118m shows that for the case of three of the profiles, the calculated temperature decreases steadily and sharply from the grain surface to the interface between the overgrowth and the older core, whereas in one of the profiles, the calculated temperature gradually increases over this interval. It would not be
wholly unexpected for there to have been a change in temperature of the system over the course of young zircon rim formation in this sample. This does not, however, explain the lack of agreement between the three profiles that exhibit a decline in temperature versus the one profile that exhibits an increase over the course of the overgrowth. Instead, this variation is probably due to the fact that the zircons extracted from this sample are probably not in equilibrium. It is quite possible that the \( a_{\text{SiO}_2} \) and \( a_{\text{TiO}_2} \) applicable to the young overgrowth examined in depth profile 2 (the one that exhibits an increase in temperature) are drastically different than those applicable to profiles 1, 3, and 5 (those exhibiting a decrease in temperature). It is also likely that \( a_{\text{SiO}_2} \) and \( a_{\text{TiO}_2} \) changed throughout the time period over which the young rims formed in this sample. It is not unreasonable to posture that, if the \( a_{\text{SiO}_2} \) and \( a_{\text{TiO}_2} \) were known for each point along every depth profile, these grains might yield uniform crystallization temperatures. In fact, they should yield uniform temperatures since they are from the same rock and crystallized contemporaneously. This same argument can also be applied to samples 0621a and 08010g, and suggest that the scale of geochemical heterogeneity at the time of zircon crystallization needs to be taken into account when using the Ti content of zircon to calculate crystallization temperatures.

2. Y in zircon

The geochemical behavior of Y, measured during depth profiling, is often regarded as similar to the HREE (e.g. Lu). Results demonstrate that Y is not homogeneously distributed throughout the portions of the zircons analysed. A good example of this is in the zircons from sample 08010g. These grains apparently
crystallized at the same time and under the same conditions, but the concentration of $Y$ varies by a factor of several thousands of ppm (e.g. from $<500$ to $\sim4500$ ppm) between and throughout the zircons from this sample. When the variation of $Y$, a proxy for Lu, is examined in conjunction with the several weight percent variation of Hf throughout these grains, the blanket assumption that radiogenic in-growth of Hf need not be considered when extending the present day Hf isotopic composition of a zircon to the composition of the melt or rock from which zircon crystallized may not be appropriate. If $Y$ can be used as a proxy for Lu in these samples, its variation suggests that the Lu/Hf ratio can change significantly throughout metamorphic rims that are of the same age. This would likely mean that the Hf isotopic compositions from certain grains would need to be corrected for in-growth, whereas others might not. More work would be necessary to test this hypothesis.

**VI. Conclusions**

U-Pb ages, Ti, Hf and Y was determined via SIMS on zircons extracted from (U)HP gneisses that host the world’s youngest mafic eclogites. The protoliths of two of these gneisses (0621a and 03118m) contain abundant Cretaceous aged zircons, indicating a potentially volcanioclastic or volcanigenic origin, whereas sample 08010g appears to have originated as a partial melt. Polished and unpolished grains, as well as zircon depth profiles reveal the timing of zircon growth and the trace element record of processes related to exhumation from (U)HP conditions in these gneisses. The main conclusions of this study are:
1) The U-Pb ages from zircons in the gneisses provide information about the protoliths of these samples followed by final crystallization of the gneisses in the middle/upper crust during exhumation of the (U)HP terrane and metamorphic core complex formation. There is no information in the U-Pb systematics of these zircons that bears on the timing of initial subduction or the onset/duration of (U)HP metamorphism. This observation comes from comparison of the U-Pb ages from these zircons with pre-existing geochronological information from the Woodlark Rift.

2) Individual zircon grains crystallizing in the same rock, under the same geological conditions, and at the same time as each other can develop contrasting trace element concentrations. This is probably a function of small scale geochemical disequilibrium present throughout the host rock at the time of zircon crystallization.

3) The application of trace element thermometers (e.g. Ti) and isotopic composition measurements (e.g. Hf) in zircons from gneissic rocks need to account for the potential lack of internal equilibrium, as is evidenced by the highly variable Ti, Hf, and Y concentrations in the youngest zircon domains in the three samples we examined.
References


age, Ti thermometry, and O isotopic ion microprobe depth profiling of zircon and monazite. Chemical Geology v. 262, 186 – 201.


Figure Captions

Fig. 1a. Geologic map of D’Entrecasteaux Islands region with sample localities. Major lithological and structural features of the region are from Monteleone et al., 2007 and Little et al., 2011. PNG = Papua New Guinea; G.I. = Goodenough Island; F.I. = Fergusson Island; N.I. = Normanby Island; T.I. = Tumabaguna Island (UHP locale; Baldwin et al., 2008).

Fig. 1b. Representative cathode ray luminescence images for zircons extracted from gneisses in the D’Entrecasteaux Islands.

Fig. 1c. Schematic diagram illustrating different types of SIMS analyses performed in this study.

Fig. 1d. Schematic diagram illustrating construction of Terra-Wasserburg model ages (T.W. model ages).

Fig. 2a. U-Pb zircon results from sample 03118m.

Fig. 2b. U-Pb zircon results from sample 08010g.

Fig. 2c. U-Pb zircon results from sample 0621a.

Fig. 3. Ti, Hf, and Y results for depth profiles from sample 03118m.

Fig. 4. Ti, Hf, and Y results for depth profiles from sample 08010g.

Fig. 5. Ti, Hf, and Y results for depth profiles from sample 0621a.
The full data set used to construct figures in this chapter is contained in an excel workbook in the electronic appendix for chapter 3. This workbook is divided into six worksheets.

**Worksheet 1; “SpotModSamples”:** Results of U-Pb dating in spot mode

**Worksheet 2; “DepthProfilesUPb”:** Results of U-Pb dating in depth profile mode

**Worksheets 3, 4, & 5; “0621ConcentrationDepthProfile”, “03118mConcentrationDepthProfile”, “08010ConcentrationDepthProfile”:** Results of trace element determinations for the depth profiles from samples 0621, 03118m, and 08010g.

**Worksheet 6; “AdditionalSamplesNotinText”:** Contains U-Pb data performed in spot-mode for samples not discussed in the text. Note that data collected in spot mode for samples discussed in this text is also included in this data set.
Chapter 3; Figure 1a

[Map showing geographical locations and geological features]

Chapter 3; Figure 1b

[Images of polished zircon cross sections]

Chapter 3; Figure 1c

Three types of SIMS Analysis:
1) Polished zircon cross sections in spot mode (U-Pb age determination)
2) Unpolished zircon surface in spot mode (U-Pb age determination)
3) Unpolished zircon surface in depth profiling mode (Hf, Ti, Y, & U-Pb age determination)

representative CL images of polished zircon cross sections from three samples examined in this study

Chapter 3; Figure 1d

D'Entrecasteaux Islands
- Blueschist facies metasediment & basalt
- Shear zones bounding MCCs (amphibolite facies)
- Core zones of MCCs (eclogite & amphibolite facies)
- Pliocene and younger volcanic deposits
- Pliocene and younger granodiorite plutons
- Equivalents of the Papuan Ultramafic Belt (basalt, gabbro, & serpentinite)
- Upper plate of MCCs (basalt, gabbro, serpentinite, & sediments)

Papuan Peninsula
- Blueschist facies metasediment & basalt
- Undifferentiated marine chemical & clastic rocks
- Raised coral terrace

Regional Features
- Faults
- Woodlark Basin Seafloor Spreading Center

Sample 0621a
Wakonai Shear Zone Gneiss

Sample 03118m
Quartz-Feldspar Gneiss

Sample 08010g
Quartz-Feldspar Gneiss (T.I.)

Chapter 3; Figure 1a

G.I.
F.I.
N.I.
PNG
Solomon Sea
100 km
Coral Sea

Chapter 3; Figure 1b

PNG 03118m
PNG 08010g
PNG 0621a

15 KV 20.066 nA
15 KV 20.066 nA
15 KV 20.066 nA

100 µm
90 µm
100 µm

Chapter 3; Figure 1c

D'Entrecasteaux Islands

Blue-entrecasteaux Islands

Sample 0621a
Wakonai Shear Zone Gneiss

Sample 03118m
Quartz-Feldspar Gneiss

Sample 08010g
Quartz-Feldspar Gneiss (T.I.)

Chapter 3; Figure 1d

representative CL images of polished zircon cross sections from three samples examined in this study

Three types of SIMS Analysis:
1) Polished zircon cross sections in spot mode (U-Pb age determination)
2) Unpolished zircon surface in spot mode (U-Pb age determination)
3) Unpolished zircon surface in depth profiling mode (Hf, Ti, Y, & U-Pb age determination)
Chapter 3; Figure 2a

U-Pb Zircon Analyses for PNG 03118m

Chapter 3; Figure 2b

U-Pb Zircon Analyses for PNG 08010g

Chapter 3; Figure 2c

U-Pb Zircon Analyses for PNG 0621
PNG 03118m

‘X’ enclosed in point on plots denotes location of boundary between young overgrowth and old core

increasing depth; from external grain surface to ~15 µm
Chapter 3; Figure 4

PNG 08010g

- Ti Saturation Thermometry (degrees celcius)
- weight percent Hf
- ppm Y

Increasing depth; from external grain surface to ~15 µm
Ti Saturation Thermometry

(degrees celsius)

Increasing depth; from external grain surface to ~15 µm

weight percent Hf

depth profile: block number

increasing depth; from external grain surface to ~15 µm

Chapter 3; Figure 5

PNG 0621
Vita

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