

# Petrologic, isotopic, and radiometric age constraints on the origin and tectonic history of the Malino Metamorphic Complex, NW Sulawesi, Indonesia

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## Abstract

The Malino Metamorphic Complex (MMC) is located at the western end of the north arm of Sulawesi. It consists of mica schists and gneisses (derived from proximal turbidite and granitoid protoliths), with intercalations of greenschist, amphibolite, marble, and quartzite, forming an E-W elongated dome-like structure bounded on all sides by faults. The age of the MMC is constrained between Devonian and Early Carboniferous. This Paleozoic age, the presence of Archean and Proterozoic inherited zircons, and the isotopic signature of the mica schists and gneisses indicate that the terrane was derived from the New Guinea-Australian margin of Gondwana. Similarities with basement rocks in the Bird's Head suggests a common origin. Greenschists forming a discontinuous selvage (metamorphic carapace) around the complex were derived from adjacent autochthonous Paleogene formations. The rocks of the MMC show a Barrovian-type progression from greenschist through epidote-amphibolite to amphibolite facies. P–T estimations suggest a depth of burial of up to 27–30 km. K/Ar and <sup>40</sup>Ar/<sup>39</sup>Ar cooling ages of 23–11 Ma, and a 7 Ma age for unconformably overlying volcanic rocks, indicate that the complex was exhumed during the Miocene. Two tectonic scenarios are considered: 1. the continental fragment docked with Sulawesi during the Mesozoic and was exhumed as a metamorphic core complex during the Miocene; 2. it was subducted beneath the north arm during the late Oligocene and then rapidly returned back to the surface.

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## 1. Introduction

The Indonesian region comprises a complex assemblage of continental fragments of Gondwanan origin, which are interspersed with island arcs, accretionary complexes, dismembered ophiolite terranes, and marginal basins. It can be

divided into two main tectonic entities (Fig. 1A) based on the time of accretion of the continental fragments: Sundaland (mainland SE Asia and western Indonesia), which had become a stable craton by the end of the Mesozoic (e.g., Metcalfe, 1996 and references therein), and eastern Indonesia (east Sulawesi and islands to the east and southeast), where the first microcontinents began to arrive during the late Oligocene/early Miocene (e.g., Audley-Charles et al., 1972; Katili, 1978; Hamilton, 1979; Milsom, 2000; Hall and Wilson, 2000).

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Located between these two domains is the Western Sulawesi Province (Fig. 1B), which began to develop at the southeastern margin of Sundaland in the late Mesozoic. A number of authors have suggested that a microcontinent is incorporated in its basement, now (largely) covered by Cenozoic deposits, but opinions differ as to whether it accreted to Sundaland during the late Mesozoic (Audley-

Charles et al., 1988; Metcalfe, 1990, 1996; Priadi et al., 1993; Longley, 1997; Elburg et al., 2003a), or Miocene (Priadi et al., 1994; Bergman et al., 1996; Polvé et al., 1997, 2001; Elburg et al., 2003a). The main evidence published to date in support of the presence of old crustal rocks in Western Sulawesi is provided by the results of radiogenic isotope analyses (Sr, Nd, and Pb) of Middle Miocene to Pliocene

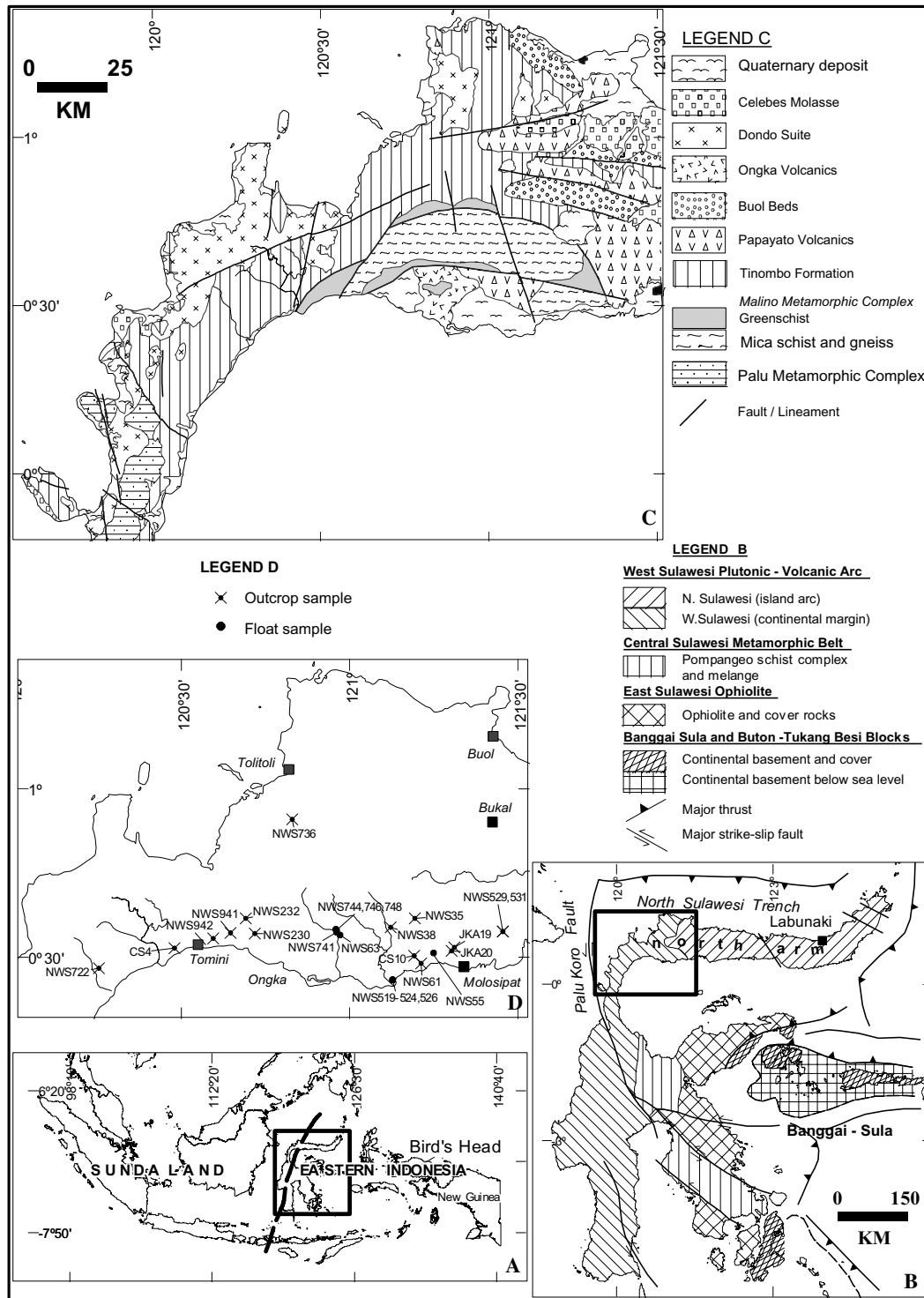


Fig. 1. (A) Geology of NW Sulawesi (modified after Ratman, 1976); (B) Sample locations; (C) Map of Indonesia showing the location of Sulawesi (boxed), boundary between Sundaland and Eastern Indonesia, and the Bird's Head; (D) Tectonic framework of Sulawesi; (A) is boxed.

potassic calc-alkaline to ultrapotassic igneous rocks from the central and northern parts of the province, which suggest involvement of lithospheric mantle and crustal material of Australian affinity in the magma generation (Priadi et al., 1993, 1994; Bergman et al., 1996; Polvé et al., 1997, 2001; Elburg and Foden, 1999a; Elburg et al., 2003a). Additional indirect evidence includes the presence of inherited zircons in the igneous suites displaying Paleozoic to Proterozoic U/Pb and Pb/Pb ages (Bergman et al., 1996), and the occurrence of lower crustal granulites exposed as tectonic fragments along the Palu–Koro Fault, which show an Australian-like isotopic signature (Priadi et al., 1993; Elburg et al., 2003a).

We carried out radiogenic isotope analyses, radiometric age studies, and petrographic examinations of a number of samples collected from the Malino and Palu Metamorphic Complexes, both located in northwest Sulawesi (Fig. 1C), in order to find more direct evidence for the hypothesis that one or more fragments derived from the New Guinea–Australian margin accreted to Western Sulawesi, either during the late Mesozoic or mid-Tertiary. These two areas were chosen as they were considered the most likely sites to contain old allochthonous rocks. Our work unambiguously demonstrates that both complexes indeed contain rocks of New Guinea–Australian origin, as evidenced by the age of the protoliths (Devonian – early Carboniferous and Permian–Triassic, respectively), their isotopic characteristics and the presence of very old inherited zircons.

This paper deals only with the Malino Metamorphic Complex (MMC). The results obtained from the Palu Metamorphic Complex (PMC) will be discussed in a separate paper. The MMC (Fig. 1) underlies an inaccessible rugged, E–W trending mountain range. Malino Mountain is the highest peak (2850 m). Lack of access is probably the main reason that the complex has been subjected to only a few studies until now. These formed part of regional mapping surveys undertaken in NW Sulawesi by Ahlburg (1913), Koperberg (1929) and Ratman (1976), who traversed only the marginal zones of the complex (mainly the southern part). In addition to these three regional surveys, Egeler (1947) carried out detailed petrographic studies on a number of samples collected by Brouwer (1934) from the most southwesterly part of the complex.

In the early 1970s, Rio Tinto undertook a helicopter-supported mineral exploration programme in NW Sulawesi, where float and outcrop observations were made at widely spaced landing sites. A second, smaller programme returned to the area in the mid-1990s, during which time two of the authors (TvL and M) collected over 30 outcrop and float samples from the southern margin of the complex. Selected samples were subjected to the following analyses: petrographic (23 samples), whole rock and trace elements (16), electron-probe micro-analyses or EPMA (10), Sr, Nd and Pb isotopes (7), K/Ar (2),  $^{40}\text{Ar}/^{39}\text{Ar}$  (6), and U/Pb zircon dating (3). Sample locations are shown in Fig. 1D; analytical methods are presented in the Appendix A.

The aim of the present paper is to report the analytical results from the MMC and to integrate them with previously published and unpublished data in an attempt to elucidate the origin and history of the MMC.

## 2. Tectonic setting

Sulawesi is generally subdivided into four lithotectonic units (Fig. 1B; Sukanto, 1975; Hamilton, 1979). These are from west to east: 1. Cenozoic Western Sulawesi Plutono-Volcanic Arc; 2. Central Sulawesi Metamorphic Belt; 3. Eastern Sulawesi Ophiolite Belt; and 4. Microcontinental fragments of Banggai-Sula and Tukang Besi-Buton. The Western Sulawesi Plutono-Volcanic Arc consists of two distinct provinces, viz a continental margin segment ('Western Sulawesi') and an island arc built on oceanic crust ('Northern Sulawesi') (Taylor and Van Leeuwen, 1980; Elburg et al., 2003a; van Leeuwen and Muhandjo, 2005). The MMC is located in the boundary area between these two domains. The name Western Sulawesi Plutono-Volcanic Arc is somewhat of a misnomer as only the Northern Sulawesi segment represents a true volcanic arc (Calvert, 2000; van Leeuwen and Muhandjo, 2005).

The earliest recorded geological event in the Sulawesi region is represented by blueschists and other metamorphic rocks exposed in the South Arm in the Bantimala Mélange Complex (Sukanto, 1975; Wakita et al., 1996) and the Latimojong Mountains (Bergman et al., 1996), and in the Central Sulawesi Metamorphic Belt (Parkinson, 1991, 1998). Parkinson et al. (1998) suggest that they form part of an arcuate zone of dismembered accretionary complexes metamorphosed during the mid-Cretaceous, which extends through southeast Kalimantan to west and central Java along the SE margin of Sundaland. Several continental fragments were incorporated into the Sundaland margin in the Bantimala area (Wakita et al., 1996), the Palu area (authors' unpublished data) and East Java (Smyth, 2005).

During the Late Cretaceous, forearc basins developed in Western Sulawesi on the metamorphic basement during northwesterly directed subduction beneath Sundaland (van Leeuwen, 1981; Hasan, 1991; Sukanto and Simandjuntak, 1983; van Leeuwen and Muhandjo, 2005). Subduction and associated magmatism continued or resumed beneath Western Sulawesi during the Early Tertiary (van Leeuwen, 1981; Yuwono et al., 1988), when Western Sulawesi rifted away from Borneo resulting in lithospheric thinning of the Makassar Strait (e.g. Calvert and Hall, 2003), possibly related to the opening of the Celebes Sea (Hall, 1996). Volcanic arc activity continued intermittently into the Early Miocene. Thick carbonate and siliciclastic deposits formed in the southern and northern parts of Western Sulawesi, respectively.

In Northern Sulawesi, a largely submerged volcanic arc developed along the North Arm during the Early Tertiary on a basement of oceanic crust, producing dominantly basaltic rocks of tholeiitic composition (Papayato Volcanics; Trail et al., 1974; Kavalieris et al., 1992; Pearson and Caira, 1999; Elburg et al., 2003a; van Leeuwen and Muhandjo, 2005).

According to most geodynamic reconstructions, Early Tertiary volcanic activity was related to northerly directed subduction of Indian Ocean lithosphere beneath the North Arm (e.g., Rangin et al., 1997; Daly et al., 1991; Hall, 1996, 2002).

Ophiolite sequences widely exposed in the Eastern Sulawesi Ophiolite Belt may have disparate origins, including Indian Ocean MORB, marginal basin and Sundaland forearc crust (Hall, 2002), or alternatively they may have formed in one tectonic setting (oceanic plateau) and were later overprinted by magmatism in a different environment (Kadariusman et al., 2004). The ophiolites are now intercalated and juxtaposed in a complex manner with Mesozoic and Tertiary sedimentary rocks, as a result of Late Oligocene/Early Miocene collision, subsequent contraction, and later strike-slip faulting (Brouwer, 1934; Sukanto and Simandjuntak, 1983; Hall, 1996, 2002). Collision continued throughout the Neogene, involving continental terranes with stratigraphic affinities to the northern margin of Australia. These terranes were transported as either separate fragments with drift histories independent of Australia, from which they had rifted during the Mesozoic (Pigram and Panggabean, 1984; Silver et al., 1985; de Smet, 1989; Struckmeyer et al., 1993; Villeneuve et al., 1998), or as a single large unit that formed a promontory of the northern Australian margin (referred to by Hall (2002) as the 'extended Bird's Head microcontinent' and was subsequently fragmented along regional left-lateral shear zones (Katili, 1978; Hamilton, 1979; Charlton, 1986, 2001; Audley-Charles et al., 1988; Daly et al., 1991; Hall, 1996, 2002).

As discussed below, an early Miocene tectonic event observed in the western and central parts of the North Arm may be collision related. In Western Sulawesi, with the exception of the northernmost part, the effects of these Tertiary collision events manifested themselves only in the latest Miocene-Pliocene (Calvert, 2000; Hall and Wilson, 2000) and are probably related to the collision of the Banggai-Sula fragment with the East Arm of Sulawesi (Hall, 2002). They include west-directed thrusting, uplift and erosion, which were preceded by, and partly overlapped with, high-K magmatism (~14–1 Ma) that is generally believed to be post-subduction in origin (Yuwono et al., 1988; Bergman et al., 1996; Polvé et al., 1997, 2001; Elburg and Foden, 1999a,b; Elburg et al., 2002, 2003a). In contrast, subduction took place beneath the North Arm throughout most of the Neogene, producing widespread calc-alkaline volcanism, which is still active in the northern tip of the arm (Kavalieris et al., 1992; Polvé et al., 1997).

### 3. Local geology

The MMC is located in the Tolitoli region at the western end of the north arm of Sulawesi, which straddles the Western and Northern Sulawesi provinces (Fig. 1B and C). It is exposed over an area approximately 125 km long and up to 30 km wide. This region was uplifted in the late Cenozoic, leading to exposure not only of the MMC, but also of Paleogene rocks (Tinombo Formation in the western part

and Papayato Volcanics in the east) and Late Miocene–Pliocene granite bodies (Dondo Suite). Neogene sedimentary formations (Buol Beds; Celebes Molasse) and volcanic deposits (Ongka Volcanics) have been preserved locally at lower elevations near the coast and in intermontane basins. The MMC is flanked by the two Paleogene formations and overlain unconformably by the Ongka Volcanics.

The Tinombo Formation (Ahlburg, 1913) consists of a >3 km thick sequence of weakly metamorphosed sedimentary and subordinate volcanic rocks. Paleontological evidence suggests that the sequence was deposited during the Middle Eocene to earliest Miocene (Ratman, 1976; van Leeuwen and Muhandjo, 2005).

The Papayato Volcanics (Trail et al., 1974) are a thick series of basaltic lava flows and breccias and less voluminous felsic volcanics, which contain rare intercalations of pelagic limestone, radiolarian chert and greywacke. These are intruded by co-magmatic stocks and dykes of gabbroic/dioritic and rhyolitic composition. The volcanic rocks are predominantly of tholeiitic composition and of oceanic arc affinity (Elburg et al., 2003a; van Leeuwen and Muhandjo, 2005). The Papayato Volcanics continue eastwards, along the North Arm, forming a volcanic belt with a total strike length of about 275 km. Limited paleontological and radiometric dates obtained from samples collected from the Tolitoli region (van Leeuwen and Muhandjo, 2005) and further to the east (Trail et al., 1974; GRDC, 1997; Polvé et al., 1997; Rangin et al., 1997) suggest that the Papayato Volcanics have a similar age range as the Tinombo Formation, i.e., Middle Eocene to earliest Miocene. The relationship between the two Paleogene formations has not been clearly established. According to Ratman (1976) they may interfinger.

In the northeastern part of the Tolitoli region, the Tinombo Formation and Papayato Volcanics are overlain unconformably by the Early – Middle Miocene Buol Beds, a series of shallow marine sedimentary rocks and volcaniclastics (Koperberg, 1929; Ratman, 1976; van Leeuwen and Muhandjo, 2005). The Pliocene Celebes Molasse overlies these three units unconformably and is composed of weakly consolidated siliciclastics and carbonates deposited in a shallow marine environment (Ratman, 1976; Koperberg, 1929).

The Dondo Suite granitoids (Kavalieris et al., 1992; Elburg et al., 2003a) intrude the Tinombo Formation in the western part of the Tolitoli region and are commonly surrounded by contact metamorphic aureoles. The granitoids are predominantly monzogranite and quartz monzonite of high-K calc-alkaline affinity. Also found in the western part of the Tolitoli region are dykes of shoshonitic composition, which yielded a biotite K/Ar age of 6.7 Ma (Elburg et al., 2003a). The Ongka Volcanics, a series of felsic volcanics dated as ~7 Ma, share a common origin with the Dondo Suite, as indicated by close similarities in whole rock, trace element, isotopic and mineralogic compositions (van Leeuwen et al., 1994; Polvé et al., 1997; Elburg et al., 2003a).

In the eastern part of the region, stocks and dykes of andesite to dacite of calc-alkaline affinity are present, the erosion products of which occur abundantly in the Buol Beds (Koperberg, 1929). They form the western extremity of the Neogene North Sulawesi Arc, which commenced development in the Early Miocene (Kavalieris et al., 1992).

The Tinombo Formation generally shows moderate to steep dips and a dominant northwesterly strike, and is in places thrust and isoclinally folded (Brouwer, 1934; Ratman, 1976). The structure of the Papayato Volcanics is poorly known. In the Tolitoli region, the Mio-Pliocene Buol Beds overlie the more strongly deformed Paleogene formations unconformably. At Bukal, located north of the MMC, the basal part of the Buol Beds consists of massive poorly sorted and immature conglomerates that were deposited in a graben-like structure, suggesting an episode of extension, rapid erosion and deposition. The tectonic features described above, including the presence of a regional unconformity separating the Paleogene formations from the younger deposits, which have also been observed in the central part of the North Arm (Koperberg, 1929; Ratman, 1976; Kavalieris et al., 1992; Carlile and Mitchell, 1994; Pearson and Caira, 1999), point to a significant tectonic event which took place in the North Arm in the earliest Miocene. Kavalieris et al. (1992) suggest that this event may be related to obduction of the East Sulawesi Ophiolite in the East Arm, or more likely the inception of the subduction which generated the Neogene arc in the North Arm. Alternatively, as proposed in this paper, it may be collision-related. Another regional unconformity, roughly at the base of the Pliocene, and the widely scattered occurrence of syn-to late orogenic sediments of Pliocene age (Celebes Molasse) may be linked to the collision of the Banggai-Sula microcontinent with east Sulawesi (e.g., Perelló, 1994).

#### 4. General characteristics of the Malino Metamorphic Complex

##### 4.1. Field relations

The data available, including float observations made in rivers draining the central part of the MMC (Ahlburg, 1913; Koperberg, 1929; Egeler, 1947; Ratman, 1976; Rio Tinto, unpublished data; this study), suggest that the complex consists predominantly of mica-bearing quartzo-feldspathic schists and gneisses, which contain intercalations of quartzite, graphite schist, talc schist, marble, amphibolite, amphibole schist and gneiss. Greenschist-facies rocks ('greenschists') form a discontinuous selvage around the complex, and are also found in the uppermost part of the mica schist/gneiss formation. Along the western margin marble is interbedded with the greenschists and at the eastern end, they appear to be (spatially) closely related to amphibolites (Koperberg, 1929). In the latter area basaltic rocks of the Papayato Volcanics grade into greenschists (Koperberg, 1929; Kavalieris et al., 1992). In the transition zone remnants of relatively pristine basalt, up to 75 cm in

diameter, are engulfed by greenschist (Kavalieris, written communication, 1994). In this paper we refer to the greenschist and amphibolite-bearing rocks as 'metabasites', as they have been derived largely from basaltic rocks.

Interpretation of the structure of the MMC from Shuttle Radar TM (Fig. 4A) combined with field observations (Koperberg, 1929; Ratman, 1976; this study) suggests that the MMC has a dome-like shape, elongated in an E-W direction, and is bounded on all sides by faults, the nature of which is poorly known. A sinuous E-W trending fault zone, up to 100 m wide and filled with pyritic gouge material, separates the complex to the south from a block consisting of the Papayato Volcanics and a small exposure of metamorphic rocks, unconformably covered by the Ongka Volcanics. This block appears to have been down-faulted relative to the complex along a south-dipping (low-angle?) normal fault. The western margin of the complex is bounded by a NE trending fault zone consisting of pods of phyllite and greenschist sandwiched between the Tinombo phyllites to the west and the MMC rocks to the east. No field observations are available for the sinuous E-W fault zone that forms the northern boundary of the MMC. The complex is cut by a series of late NW to NNW trending faults, along which minor right lateral movement has taken place. Dominant strike-dip trends on the MMC and Tertiary cover rocks are shown in Fig. 4B, but vary in detail. Folding at various scales have been observed at several localities (Koperberg, 1929). There is some suggestion that the cover rocks dip away from the complex.

Some of the samples from the MMC examined petrographically, display only weak late-stage deformation and alteration. Others show significant deformation, including shearing, plastic deformation of quartz and plagioclase (suggesting ductile flow), and crenulation folding of muscovite. In these rocks associated retrograde alteration is more strongly developed than in the rocks that are less deformed.

Post-metamorphic dykes and stocks belonging to the Dondo Suite intrude the MMC along its western margin. In the southern part Koperberg (1929) observed a strongly deformed hornblende granite and Ahlburg (1913) a gneissic granite, which may represent syn-tectonic granitoids.

##### 4.2. Petrology

The quartzo-feldspathic mica schists and gneisses comprise metasediments (dominantly meta-psammites) and subordinate meta-granitoids. They generally show a well-defined foliation. Some gneisses have feldspar augen textures. The most common rock types are quartz-mica schist, and quartz-feldspar-mica schist and gneiss. Muscovite is the dominant mica phase. It occurs as flakes mutually interlocking with plagioclase and/or quartz, and also as inclusions in these minerals, and constitutes up to 30 vol. % of the rock. Biotite is deep brown in thin section and commonly chloritized. It occurs either as the only mica type or together with muscovite. Plagioclase is the most common feldspar. It ranges in composition from albite to oligoclase.

Garnet is a significant constituent of some schists and gneisses (5–15 vol. %), occurring as barely visible grains to porphyroblasts up to 2 cm in diameter that contain inclusions of quartz and muscovite. Prograde epidote, and less commonly hornblende, were observed in several samples. Accessory minerals include rutile, apatite, zircon, sphene, magnetite, ilmenite, and tourmaline.

The mineral assemblages of the schists and gneisses can be divided into two groups: 1. Qtz + Ab + Ms ± Bt ± Kfs ± Zo ± Chl (lower grade or biotite zone); and 2. Qtz + Ab + Grt + Ms ± Kfs ± Zo ± Bt (medium grade or garnet zone) (Figs. 2, 3). Koperberg (1929) reports the occurrence of a garnet–muscovite–feldspar–quartz gneiss containing kyanite with cores of staurolite. Thus the quartzo–feldspathic rocks appear to show progressive metamorphism from the biotite zone through the garnet zone to probably the staurolite zone.

The meta-basites show a similar progression from greenschist through epidote–amphibolite to amphibolite facies (Figs. 2, 3). The greenschists are fine-grained, with a well-developed schistosity defined by preferred orientation of chlorite, epidote, and albite. The schists found along the western margin of the complex commonly show layering of chlorite + epidote-rich bands interposed with quartz-rich bands, suggesting a sedimentary origin (Egeler, 1947). In contrast, the most common relict fabrics observed in the southeastern margin schists are igneous. The mineral composition of the greenschists consists of Chl + Ab + Qtz ± Ep ± Act ± Ser ± Mag ± Spn ± Cal. Epidote and magnetite occur as porphyroblastic minerals. Sericite is a significant constituent of the western greenschists (up to 20 vol. %), but much less abundant in the southeastern schists (<5%). Actinolite shows the reversed relationship. The textural and

compositional differences displayed by the western and southeastern greenschists reflect differences in protoliths (see Discussion).

The epidote–amphibolite facies rocks are characterized by the presence of blue–green hornblende and albite. They are fine- to medium-grained and usually show a nematoblastic schistosity defined by the preferred orientation of hornblende, muscovite and epidote crystals, or by the segregation of quartz and hornblende; a few samples show isotropic fabrics. Characteristic prograde mineral assemblages observed in these rocks comprise Ab + Hbl + Ep + Qtz ± Grt ± Ms ± Ksp. Samples CS-10 and NWS-521 differ from the other epidote–amphibolite samples in having significant amounts of muscovite and K-feldspar, suggesting a different protolith. Accessory minerals are rutile, sphene, magnetite, ilmenite, and calcite. The main rock forming minerals occur in the following abundances: hornblende, 20–50 vol. %; quartz, 10–50%; plagioclase, 1–17%; garnet, 4–40%; and epidote, 4–15%.

In the amphibolite facies rocks, green–pale brown hornblende takes the place of blue–green hornblende, epidote decreases in abundance, and plagioclase is of albite to oligoclase composition. Ab–Olig + Hbl + Qtz ± Ep ± Grt ± Ms ± Ksp ± Mag, with calcite, rutile and ilmenite as accessory minerals, form the mineral assemblages observed in these rocks. The amphibolites are medium-grained, with hornblende and plagioclase prisms defining a well-developed nematoblastic schistosity. As in the case of the epidote–amphibolite rocks, plagioclase may be present in small quantities only.

Marble, commonly containing graphite, has been observed as lenses enveloped by mica schist in several localities along the southern margin (Ahlburg, 1913). Retro-

| Sample No.                                             | Rock name | Chl             | Qtz | Ep | Ms | Ab | Bt | Grt | Mag | K-fel | Rut | Cal | Ilm | Tour |
|--------------------------------------------------------|-----------|-----------------|-----|----|----|----|----|-----|-----|-------|-----|-----|-----|------|
| <b>Medium grade metasediment (garnet zone)</b>         |           |                 |     |    |    |    |    |     |     |       |     |     |     |      |
| 1                                                      | NWS-38    | grt-musc schist | x   | x  | x  | x  |    | x   | x   |       |     | x   | x   |      |
| 2                                                      | NWS-55    | grt-musc schist | x   | x  |    | x  | x  | x   | x   |       |     |     |     | x    |
| 3                                                      | NWS-63    | grt-musc gneiss | x   | x  |    | x  | x  | x   | x   |       |     |     |     | x    |
| 4                                                      | NWS-520   | grt-musc schist | x   | x  |    | x  | x  | x   | x   |       | x   |     | x   |      |
| 5                                                      | NWS-522   | grt-musc schist |     | x  | x  | x  | x  | x   |     | x     |     |     |     |      |
| <b>Low grade metasediment/granitoid (biotite zone)</b> |           |                 |     |    |    |    |    |     |     |       |     |     |     |      |
| 6                                                      | NWS-61    | musc-schist     | x   | x  |    | x  | x  |     | x   |       |     | x   |     |      |
| 7                                                      | NWS-942   | musc-schist     | x   | x  | x  | x  | x  |     | x   |       | x   | x   |     |      |
| 8                                                      | NWS-943   | bt-musc gneiss  |     | x  | x  | x  | x  | x   | x   | x     | x   | x   | x   |      |

|   |               |           |     |     |    |    |    |     |      |     |     |     |
|---|---------------|-----------|-----|-----|----|----|----|-----|------|-----|-----|-----|
|   | Metalimestone |           |     |     |    |    |    |     |      |     |     |     |
|   | Sample No.    | Rock name | Chl | Qtz | Ep | Ms | Ab | Cal | Plag | Act | Grt | Mag |
| 9 | NWS-741       | marble    |     | x   |    | x  |    | x   |      |     |     | x   |

| Sample No.                              | Rock name | Chl             | Qtz | Ep | Ab | Plag | Act | Hbl | Grt | Ms | Par | K-fel | Bt | Rut | Mag | Ilm | Spn | Calc |
|-----------------------------------------|-----------|-----------------|-----|----|----|------|-----|-----|-----|----|-----|-------|----|-----|-----|-----|-----|------|
| <b>Metabasite (greenschist)</b>         |           |                 |     |    |    |      |     |     |     |    |     |       |    |     |     |     |     |      |
| 10                                      | NWS-230A  | chlorite schist | x   | x  | x  |      |     |     |     |    |     |       |    | x   |     |     |     |      |
| 11                                      | NWS-744   | chlorite schist | x   | x  | x  | x    |     |     |     |    |     |       |    |     | x   |     | x   | x    |
| 12                                      | CS-04     | chlorite schist | x   | x  |    | x    |     |     |     |    |     |       |    |     |     |     | x   |      |
| <b>Metabasite (epidote amphibolite)</b> |           |                 |     |    |    |      |     |     |     |    |     |       |    |     |     |     |     |      |
| 13                                      | CS-10     | grt amphibolite |     | x  | x  |      | x   |     | x   | x  |     |       | x  |     | x   | x   | x   |      |
| 14                                      | NWS-746   | grt amphibolite |     | x  | x  |      | x   |     | x   | x  | x   |       |    | x   | x   | x   | x   | x    |
| 15                                      | NWS-748   | grt amphibolite | x   | x  | x  |      | x   |     | x   | x  |     |       |    | x   | x   | x   |     |      |
| 16                                      | NWS-521   | amphibolite     | x   | x  | x  |      | x   |     | x   |    | x   |       | x  | x   | x   |     | x   | x    |
| 17                                      | NWS-523   | amphibolite     | x   | x  | x  |      | x   |     | x   |    | x   |       |    | x   |     |     |     | x    |
| <b>Metabasite (amphibolite)</b>         |           |                 |     |    |    |      |     |     |     |    |     |       |    |     |     |     |     |      |
| 18                                      | NWS-232   | amphibolite     |     | x  | x  |      | x   |     | x   |    | x   |       |    |     |     |     |     | x    |
| 19                                      | NWS-519   | grt amphibolite |     | x  | x  |      | x   |     | x   | x  | x   | x     |    | x   | x   | x   |     |      |
| 20                                      | NWS-524   | grt amphibolite | x   | x  | x  |      | x   |     | x   | x  | x   |       |    | x   | x   |     |     | x    |
| 21                                      | NWS-941   | grt amphibolite | x   | x  | x  |      | x   |     | x   | x  | x   |       | x  | x   | x   |     |     |      |

Fig. 2. Mineral assemblages of selected Malino Metamorphic Complex samples.

| Metamorphic Facies |             | Greenschist | EA     | AM |
|--------------------|-------------|-------------|--------|----|
| Metabasite         | Quartz      |             |        |    |
|                    | Albite      |             |        |    |
|                    | Plagioclase |             |        |    |
|                    | Chlorite    |             |        |    |
|                    | Epidote     |             |        |    |
|                    | Actinolite  |             |        |    |
|                    | Hornblende  |             |        |    |
|                    | Garnet      |             |        |    |
|                    | Muscovite   |             |        |    |
|                    | Biotite     |             |        |    |
|                    | K-feldspar  |             |        |    |
|                    | Calcite     |             |        |    |
|                    | Sphene      |             |        |    |
|                    | Rutile      |             |        |    |
|                    | Magnetite   |             |        |    |
|                    | Ilmenite    |             |        |    |
| Metasediment       | Chlorite    |             |        |    |
|                    | Muscovite   |             |        |    |
|                    | Biotite     |             |        |    |
|                    | Garnet      |             |        |    |
|                    | Quartz      |             |        |    |
|                    | K-feldspar  |             |        |    |
|                    | Albite      |             |        |    |
|                    | Zoisite     |             |        |    |
|                    | Rutile      |             |        |    |
|                    | Tourmaline  |             |        |    |
|                    | Magnetite   |             |        |    |
|                    | Ilmenite    |             |        |    |
|                    |             |             |        |    |
|                    |             |             |        |    |
| Zone               |             | Biotite     | Garnet |    |

Fig. 3. Mineral paragenesis of meta-basites, sediments and granitoids.

grade alteration observed in the mica schists/gneisses, amphibolites, and greenschists include chlorite, epidote, quartz, and carbonate, with rare actinolite and sericite. These minerals fill shears/fractures or replace prograde metamorphic minerals. In several epidote-amphibolite samples, a third metamorphic/alteration stage was observed, preceding the peak and retrograde metamorphic events. It is represented by inclusions in garnet that consist of Qtz + Ab + Ep + Ilm + Mag + Rut + Ms, which is a typical greenschist facies assemblage.

### 5. Major and trace element geochemistry

A total of 17 samples were analyzed (7 greenschist; 6 amphibolite; 3 meta-granitoid; and 1 meta-sediment). Selected analyses are shown in Table 1 and plotted on an extended spider diagram (Fig. 5). The main purpose of this work was to identify possible protholiths of the various metamorphic rock types.

The greenschist samples fall into two distinct groups; those collected from the southeastern margin of the complex display features that are typical of island arc tholeiitic basalts, including low K, Sr, Rb, Ba, Zr, Th, and U contents, distinct Nb–Ta negative anomalies, and flat REE patterns. Their chemical composition is very similar to that of

the Papayato basaltic rocks (Fig. 5; Elburg et al., 2003a; van Leeuwen and Muhardjo, 2005). In contrast, the greenschist samples from the western margin do show a wider compositional range for a number of major oxides and trace elements than the greenschists of the former group. Furthermore they have higher  $K_2O$ ,  $Al_2O_3$ , Rb, Zr, Nb, and Ba contents and lower  $Fe_2O_3$  and  $V$  values, and are enriched in LREE. The western schists, which as discussed above are interpreted to have a sedimentary origin, show similarities in their geochemical characteristics with greywacke samples analyzed from the adjacent Tinombo Formation (Fig. 5; van Leeuwen, unpublished data).

The geochemistry of the amphibolite samples is also similar to that of the tholeiitic Papayato Volcanics basalts, with the exception of NWS-523, which is of calc-alkaline affinity and is moderately enriched in LREE, and NWS-521, a biotite–garnet–hornblende gneiss, which differs both petrographically (see above) and chemically (Table 1) from the other samples in this group. Its precursor is possibly an alkali basalt, or alternatively it may be of sedimentary origin.

Meta-granitoid samples NWS-942 and 943 represent medium-K and high-K subalkaline granites respectively; NWS-941 is a low-K diorite showing affinity with the amphibolite samples. The amphibole-rich meta-sediment

Table 1  
Selected whole-rock analyses

|                                | CS04 | NWS 744 | NWS 746 | NWS 523 | NWS 521 | NWS 941 | NWS 942 | NWS 526 |
|--------------------------------|------|---------|---------|---------|---------|---------|---------|---------|
| <i>wt %</i>                    |      |         |         |         |         |         |         |         |
| SiO <sub>2</sub>               | 58.2 | 57.0    | 49.0    | 53.6    | 52.2    | 61.0    | 75.5    | 69.3    |
| Al <sub>2</sub> O <sub>3</sub> | 15.9 | 14.7    | 14.1    | 16.2    | 18.1    | 15.2    | 13.5    | 13.6    |
| Fe <sub>2</sub> O <sub>3</sub> | 8.72 | 10.2    | 14.7    | 8.66    | 6.81    | 8.55    | 1.58    | 4.91    |
| MnO                            | 0.14 | 0.16    | 0.22    | 0.14    | 0.10    | 0.07    | 0.02    | 0.13    |
| MgO                            | 10.9 | 3.94    | 4.75    | 6.57    | 4.22    | 2.69    | 0.46    | 1.41    |
| CaO                            | 9.18 | 7.65    | 9.3     | 7.86    | 7.52    | 5.85    | 1.32    | 1.96    |
| Na <sub>2</sub> O              | 2.49 | 2.30    | 4.17    | 4.18    | 3.34    | 3.67    | 5.24    | 5.34    |
| K <sub>2</sub> O               | 0.67 | 0.15    | 0.17    | 0.32    | 2.60    | 0.14    | 1.50    | 0.17    |
| Ti O <sub>2</sub>              | 1.16 | 1.08    | 1.98    | 0.73    | 0.84    | 0.89    | 0.12    | 0.44    |
| P <sub>2</sub> O <sub>5</sub>  | 0.14 | 0.13    | 0.16    | 0.09    | 0.33    | 0.12    | 0.04    | 0.10    |
| LOI                            | 2.02 | 2.75    | 1.72    | 1.15    | 4.09    | 1.77    | 1.25    | 1.26    |
| Total                          | 98.7 | 99.9    | 100     | 99.5    | 100     | 100     | 100     | 98.6    |
| <i>ppm</i>                     |      |         |         |         |         |         |         |         |
| Sc                             |      |         |         |         |         |         |         |         |
| V                              | 143  | 331     | 439     | 241     | 134     | 237     | nd      | 25      |
| Cr                             | 105  | 52      | 111     | 265     | 245     | 74      | 249     | 182     |
| Co                             | 18   | 27      | 33      | 32      | 19      | 12      | 1       | 2       |
| Ni                             | 57   | 24      | 38      | 54      | nd      | nd      | nd      | nd      |
| Cu                             | 104  | 12      | 41      | 30      | 37      | nd      | nd      | 11      |
| Zn                             | 70   | 42      | 90      | 64      | 83      | nd      | nd      | 74      |
| Ga                             | 19   | 17      | 18      | 18      | 22      | 18      | 13      | 16      |
| Ge                             | 1    | 1       | 2       | 1       | nd      | 2       | 1       | nd      |
| Rb                             | 85   | nd      | nd      | 3       | 101     | nd      | 47      | nd      |
| Sr                             | 154  | 147     | 90      | 145     | 594     | 181     | 173     | 73      |
| Y                              | 35   | 25      | 35      | 19      | 25      | 25      | 10      | 48      |
| Zr                             | 77   | 61      | 82      | 45      | 249     | 109     | 96      | 138     |
| Nb                             | 6    | nd      | 3       | nd      | 7       | nd      | 1       | 2       |
| Cs                             | 3    | nd      | nd      | nd      | 6.2     | nd      | 1.1     | nd      |
| Ba                             | 515  | 18      | 15      | 21      | 846     | 17      | 312     | 9       |
| La                             | 19.6 | 2.6     | 5.9     | 12.7    | 34.6    | 6.7     | 27.6    | 4.9     |
| Ce                             | 41.1 | 7.2     | 14.2    | 27.9    | 64.2    | 16.4    | 52.1    | 14.6    |
| Pr                             | 5.36 | 1.26    | 2.35    | 3.02    | 7.41    | 2.19    | 5.57    | 2.43    |
| Nd                             | 21.2 | 7       | 11.7    | 13      | 30.5    | 10.5    | 20.0    | 13.5    |
| Sm                             | 5    | 2.5     | 3.8     | 3       | 6.1     | 3.1     | 3.6     | 4.2     |
| Eu                             | 1.42 | 1.07    | 1.51    | 0.97    | 1.60    | 1.13    | 0.82    | 1.35    |
| Gd                             | 5.2  | 3.4     | 4.9     | 3.1     | 5.1     | 3.7     | 2.4     | 5.5     |
| Tb                             | 0.91 | 0.7     | 0.9     | 0.6     | 0.8     | 0.6     | 0.3     | 1.1     |
| Dy                             | 5.1  | 4.4     | 5.6     | 3.6     | 4.4     | 4.4     | 1.7     | 7.5     |
| Ho                             | 1.12 | 0.9     | 1.2     | 0.7     | 0.8     | 0.9     | 0.3     | 1.6     |
| Er                             | 2.9  | 2.8     | 3.5     | 2.1     | 2.5     | 2.8     | 1.0     | 5.1     |
| Tm                             | 0.46 | 0.44    | 0.54    | 0.31    | 0.37    | 0.41    | 0.16    | 0.78    |
| Yb                             | 3.1  | 2.8     | 3.4     | 2.1     | 2.2     | 2.7     | 1.0     | 4.9     |
| Lu                             | 0.41 | 0.41    | 0.48    | 0.29    | 0.35    | 0.41    | 0.14    | 0.82    |
| Hf                             | 2.1  | 1.7     | 2.3     | 1.5     | 7.0     | 3.1     | 2.7     | 3.7     |
| Ta                             | 0.44 | nd      | nd      | nd      | 0.6     | 0.1     | 0.5     | nd      |
| Pb                             | 6    | nd      | nd      | 7       | 26      | nd      | 21      | nd      |
| Th                             | 2.9  | 0.2     | nd      | 2       | 12.5    | 1.0     | 9.8     | 0.8     |
| U                              | 0.84 | 0.1     | 0.2     | 0.1     | 2.5     | 0.3     | 2.1     | 0.2     |

nd, not detected. Greenschist: CS04, NWS-744; Amphibolite: NWS-746, NWS-523, NWS-521; Metagranitoid: NWS-941, NWS-942, NWS-943.

sample (NWS-526) has certain features, including low K<sub>2</sub>O, Sr, Rb, Ba, Th, and U, that suggest a similar affinity.

## 6. Radiogenic isotopes

Seven samples from the MMC were subjected to isotopic analysis for Sr, Nd, and Pb with the aim of identifying possible source regions. The Sr, Nd, and Pb isotopic composition as well as the <sup>147</sup>Sm/<sup>144</sup>Nd ratios are given in Table 2 and Fig. 6. There is a clear dichotomy in the analysed sam-

ples, with NWS-526, 941, 523, and IKA-19 (Group 1) having relatively unradiogenic Sr and Pb and radiogenic Nd isotopic signatures, whereas NWS-942, 38, and 63 (Group 2) show the opposite characteristics, having more 'continental' isotopic signatures. The latter group also shows lower <sup>147</sup>Sm/<sup>144</sup>Nd ratios.

Whereas the isotopic analyses of mobile elements such as Sr and Pb for metamorphosed samples such as the ones analysed from the MMC may not always produce reliable estimates of source composition, Nd isotopes are relatively

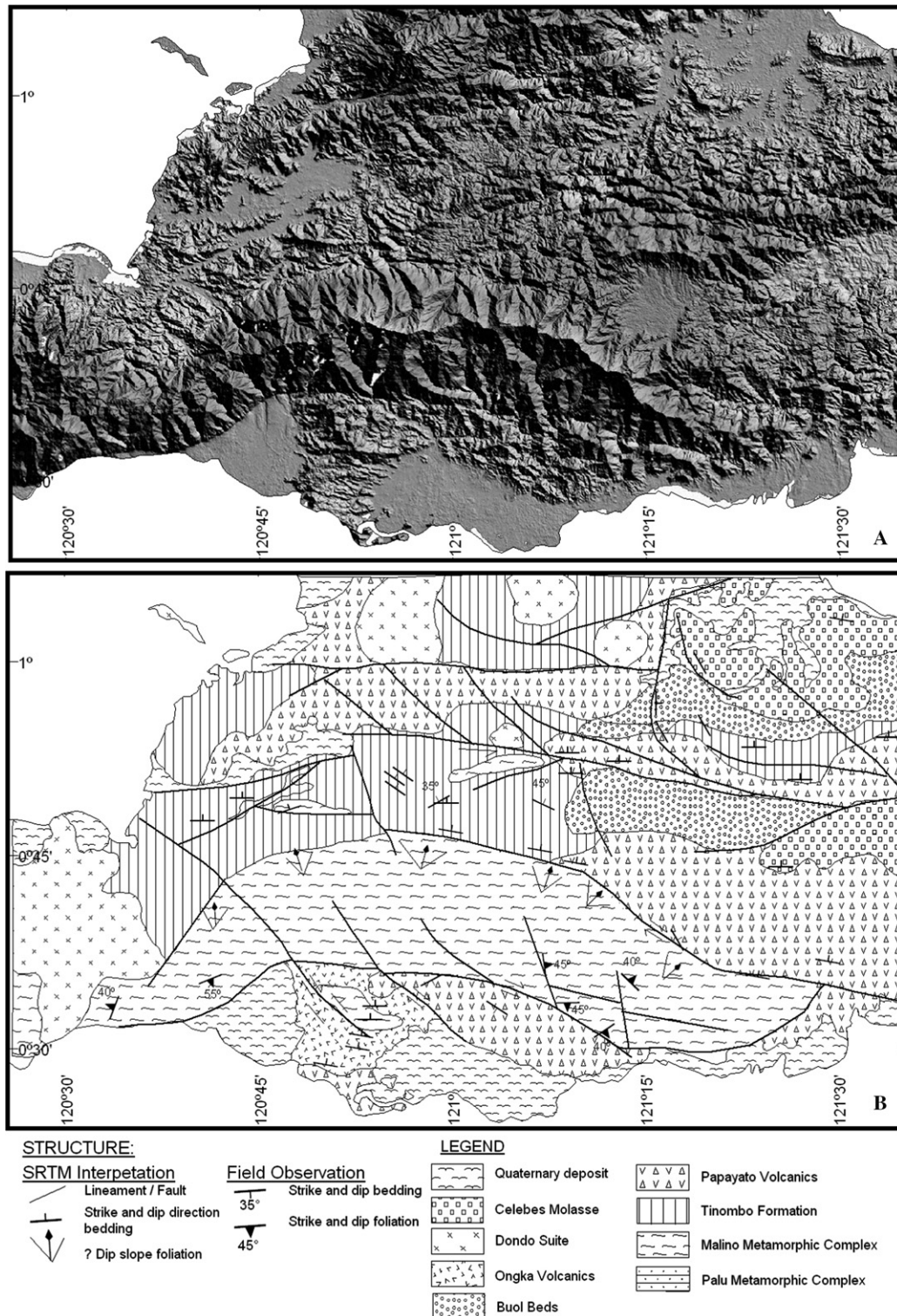


Fig. 4. (A) Shuttle Radar Thematic Mapping Image of the Malino Metamorphic Complex; (B) Interpreted geology.

unaffected by subsequent metamorphic events. The  $^{143}\text{Nd}/^{144}\text{Nd}$  ratio of the samples from Group 1 fall within the field for Mid-Ocean Ridge Basalt of which the majority of analyses for this part of the world lie between 0.5129 and 0.5132 (Hamelin and Allègre, 1985; Hamelin et al., 1985; Ito et al., 1987; White et al., 1987; Dosso et al., 1988; Mahoney et al., 1992; Rehkaemper and Hofmann, 1997). The Nd isotopic

ratios are slightly higher than those of recent island arc tholeiites from the Sangihe arc in NW Sulawesi (Fig. 6), but the data overlap with the Papayato Volcanics, which overlie the eastern side of the MMC. Sr and Pb isotopic compositions of the samples from Group 1 also resemble the Papayato Volcanics, which have been interpreted as island arc tholeiites (Elburg et al., 2003a; van Leeuwen and Muhardjo, 2005).

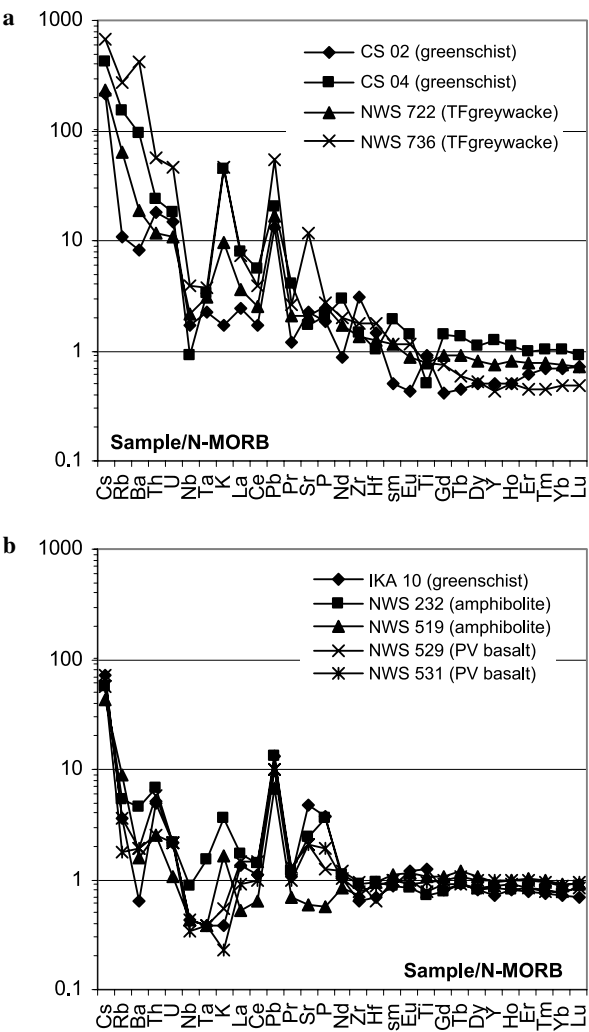


Fig. 5. Chondrite normalized trace element patterns of selected (a) greenschist (western margin) and Tinombo greywacke samples, and (b) greenschist (southeastern margin), amphibolite, and Papayato basalt samples. Normalization factors from Sun and McDonough (1989).

The Nd isotopic ratios of the samples from Group 2 are much lower than either the volcanic rocks (volcanics from the Tinombo Formation and Papayato Volcanics) or the sedimentary rocks (phyllites from the Tinombo Formation and Pompangeo Schists) that are found in areas located further to the south (Fig. 8). In terms of Nd and Sr isotopes, these Group 2 samples bear more resemblance to river sediments from northern Australia (Elburg et al., 2002). The

depleted mantle extraction age ( $T_{DM}$ ) for these samples is 1810–2060 Ma and the age of extraction from a chondrite uniform reservoir ( $T_{CHUR}$ ) varies between 1530 and 1830 Ma. This shows that the Group 1 samples had a long crustal history, with long-term enrichment of Nd over Sm and Rb over Sr, compared to depleted mantle or the chondrite uniform reservoir. Although the Sr and Nd isotopic compositions of Group 2 samples resemble those of northern Australian river sediments, their Pb isotopic ratios are invariably lower. NWS-63 is the only sample of which the Pb isotopic composition comes close to that of the Australian sediments with  $^{206}\text{Pb}/^{204}\text{Pb}=19.04$  and  $^{208}\text{Pb}/^{204}\text{Pb}=39.06$ . This may reflect that the Group 2 samples or their protoliths had only moderate enrichments of Th and U over Pb, compared with the protoliths of the Australian river sediments. The Nd isotopic compositions of Group 2 samples are much lower than for sediments from southern or northern Papua New Guinea at ( $^{143}\text{Nd}/^{144}\text{Nd}$  0.5121–0.5123; Vroon et al., 1996).

When initial Sr and Nd isotope values are calculated for meta-granitoid sample NWS-942 at 320 Ma, its U–Pb zircon age (see below), the ratios are 0.727 and 0.5115, respectively. These age-corrected ratios like the uncorrected ones are closer to those for sediments rather than for mantle-derived material at that time, suggesting that the Carboniferous granite (i.e., the precursor to the mica gneiss represented by NWS-942) was generated primarily from the melting of sedimentary material with a long crustal history.

7. Mineral compositions

Representative chemical analyses of amphibole, plagioclase, garnet, epidote, white mica, K-feldspar, and chlorite, and their calculated structural formula are presented in Tables 3 and 4.

7.1. Metabasites

Calcic amphibole (hornblende) is the main constituent of the Malino amphibolites. In both epidote-amphibolite and amphibolite rocks, it occurs as medium to large prismatic crystals (up to 5 mm). All analysed calcic-amphiboles have Ca with  $M4=1.5$  and low Na + K in the A-site ( $<0.5$ ). Plots of Si (pfu) vs  $\text{Mg}/(\text{Fe}^{2+} + \text{Mg})$  show that the composition of amphibole in amphibolite facies rocks is restricted to tschermakite, whereas amphiboles in epidote-amphibo-

Table 2 Isotopic compositions and  $^{147}\text{Sm}/^{144}\text{Nd}$  ratios of seven MMC samples

| Sample  | Description         | $^{87}\text{Sr}/^{86}\text{Sr}$ | $^{143}\text{Nd}/^{144}\text{Nd}$ | $^{206}\text{Pb}/^{204}\text{Pb}$ | $^{207}\text{Pb}/^{204}\text{Pb}$ | $^{208}\text{Pb}/^{204}\text{Pb}$ | $^{147}\text{Sm}/^{144}\text{Nd}$ |
|---------|---------------------|---------------------------------|-----------------------------------|-----------------------------------|-----------------------------------|-----------------------------------|-----------------------------------|
| NWS 526 | garnet mica schist  | 0.705209                        | 0.513009                          | 18.08                             | 15.51                             | 38.07                             | 0.2052                            |
| NWS 941 | gneiss              | 0.704394                        | 0.512960                          | 18.39                             | 15.59                             | 38.44                             | 0.1732                            |
| NWS 942 | gneiss              | 0.730166                        | 0.511696                          | 18.18                             | 15.61                             | 38.85                             | 0.1060                            |
| NWS 523 | epidote-amphibolite | 0.704367                        | 0.513015                          | 17.99                             | 15.50                             | 38.07                             | 0.1449                            |
| NWS 38  | musc. schist        | 0.743278                        | 0.511685                          | 18.39                             | 15.62                             | 38.73                             | 0.1020                            |
| NWS 63  | musc. schist        | 0.781940                        | 0.511619                          | 19.04                             | 15.76                             | 39.06                             | 0.1117                            |
| IKA19   | greenschist         | 0.705054                        | 0.512969                          | 18.25                             | 15.55                             | 38.30                             |                                   |

The Pb isotope ratios have been corrected for fractionation.

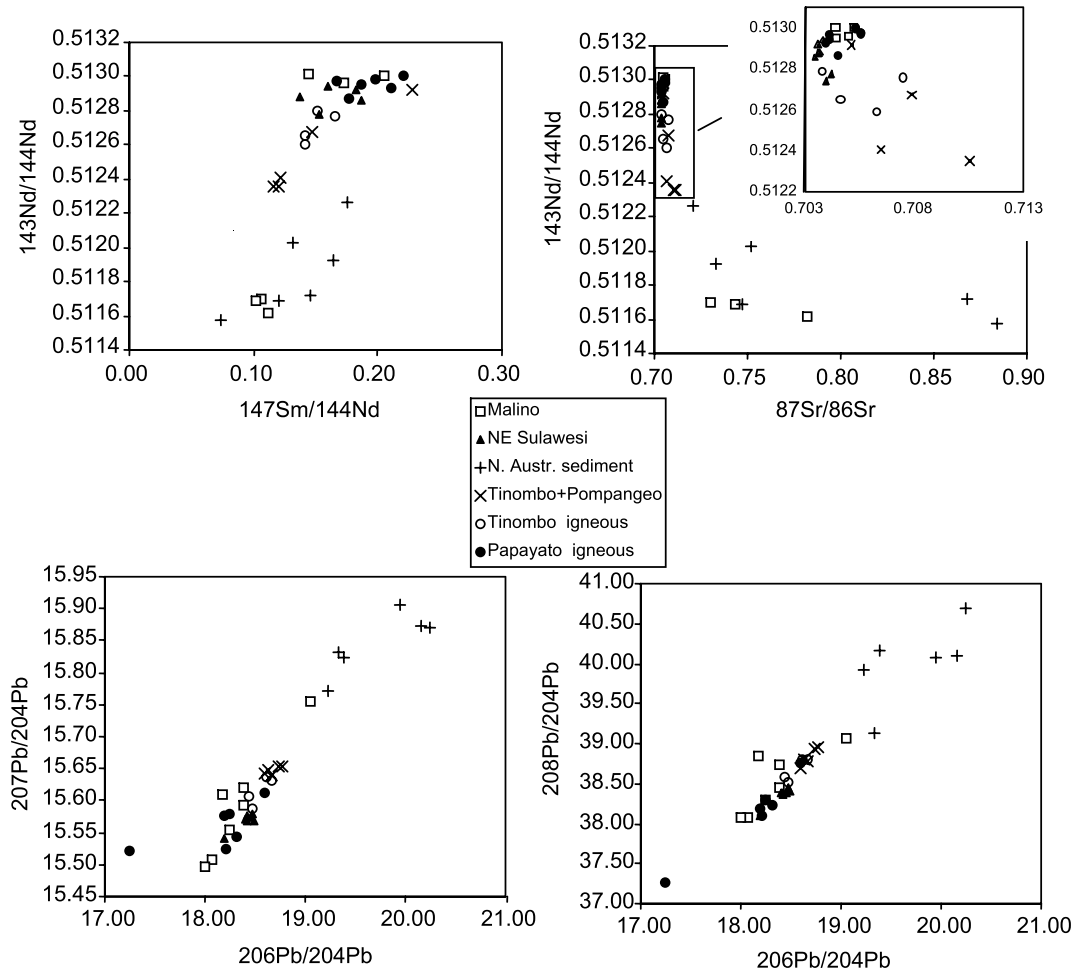


Fig. 6. Sr, Nd, Pb isotopic characteristics of the analysed Malino Metamorphic Complex samples (open squares), compared to recent volcanics from the Sangihe arc in NE Sulawesi (black triangles; [Elburg and Foden, 1998](#)); North Australian sediments (plus signs; [Elburg et al., 2002](#)); phyllites and schists from the Tinombo Formation and Pompangeo Schists (Xs; [Elburg et al., 2003a](#)); and igneous rocks from the Papayato and Tinombo Volcanics (closed and open circles respectively; [Elburg et al., 2003a](#)).

lite facies rocks vary between tschermakite and Mg-hornblende ([Fig. 7](#)).  $\text{Al}_2\text{O}_3$  contents of epidote-amphibolite facies amphiboles vary from 15.0 to 16.8 wt %; and those of amphibolite facies amphiboles from 11.1 to 16.2 wt %.

Albite ( $\text{An}_{<1-3}$ ) is the common plagioclase in the greenschist facies rocks, occurring as anhedral and untwin-

ned grains. Plagioclase in epidote-amphibolite rocks varies between  $\text{An}_1$  and  $\text{An}_{10}$ . A still wider range ( $\text{An}_{5-28}$ ) is shown by plagioclase in amphibolite rocks, as some samples have coexisting pairs of albite and oligoclase. The anorthite content of plagioclase in the meta-basites increases systematically with increasing metamorphic grade. A possible solvus gap (peristerite gap; [Maruyama et al., 1982](#)) between albite and oligoclase can be observed in two amphibolite samples ([Fig. 8](#)).

Epidote is a common mineral in greenschist and epidote-amphibolite facies rocks, but only a minor component in amphibolite facies rocks. Some epidote grains in epidote-amphibolite and amphibolite rocks show zoning, with cores that have higher birefringence than their rims, indicating a higher  $\text{Fe}^{3+}$  content in the cores. In greenschist facies rocks, most epidote is fine-grained, whereas in epidote-amphibolite rocks it is coarse grained (up to 3.0 mm). The pistacite component ( $100 \times \text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Al})$ ) of epidote-amphibolite and amphibolite rocks vary from 16 to 24 and 15 to 25, respectively, whereas in greenschist facies rocks it varies from 20 to 30. A trend of decreasing pistacite content towards zoisite can be observed from greenschist to

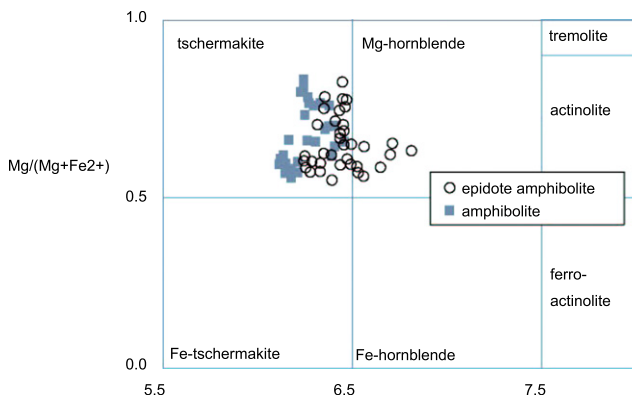


Fig. 7. Composition of amphibolite in metabasite samples ([Leake et al., 1997](#)).

Table 3  
Representative chemical analyses of amphiboles (23-oxygen basis), plagioclase (8-oxygen basis), epidote (12.5-oxygen basis), garnet (12-oxygen basis), chlorite (28-oxygen basis), and white mica (22-oxygen basis) for six MMC metabasite samples

| Amphibole                                                  |       |       |       |       |       |       |       |       | Plagioclase                    |       |        |       | Epidote                                     |       |       |       | Garnet                         |        |       | White mica                     |       |       | Chlorite                       |       |       |       |  |
|------------------------------------------------------------|-------|-------|-------|-------|-------|-------|-------|-------|--------------------------------|-------|--------|-------|---------------------------------------------|-------|-------|-------|--------------------------------|--------|-------|--------------------------------|-------|-------|--------------------------------|-------|-------|-------|--|
| Zone                                                       | EA    | EA    | EA    | EA    | AM    | AM    | AM    | AM    | Zone                           | GS    | EA     | AM    | Zone                                        | GS    | EA    | AM    | Zone                           | EA     | AM    | Zone                           | EA    | AM    | Zone                           | GS    | EA    | AM    |  |
| Sample no. NWS74 NWS74 NWS52 NWS52 NWS51 NWS51 NWS52 NWS52 |       |       |       |       |       |       |       |       | Sample no. NWS74 NWS748 NWS51  |       |        |       | Sample no. NWS74 NWS74 NWS51                |       |       |       | Sample no. NWS74 NWS51         |        |       | Sample no. NWS52 NWS51         |       |       | Sample no. NWS74 NWS74 NWS51   |       |       |       |  |
| Point no.                                                  | 8     | 8     | 1     | 1     | 9     | 9     | 4     | 4     | Point no.                      | 4     | 28     | 9     | Point no.                                   | 4     | 8     | 9     | Point No                       | 8      | 9     | Point no.                      | 3     | 9     | Point no.                      | 4     | 8     | 9     |  |
|                                                            | 8     | 10    | 7     | 8     | 2     | 16    | 2     | 27    |                                | 17    |        | 4     |                                             | 1     | 10    | 4     |                                | 11     | 1     |                                | 17    | 3     |                                | 2     | 3     | 8     |  |
| SiO <sub>2</sub>                                           | 42.94 | 42.93 | 44.72 | 43.97 | 44.14 | 42.06 | 41.54 | 45.74 | SiO <sub>2</sub>               | 68.77 | 68.44  | 62.26 | SiO <sub>2</sub>                            | 37.38 | 38.61 | 39.27 | SiO <sub>2</sub>               | 38.58  | 39.24 | SiO <sub>2</sub>               | 48.36 | 47.48 | SiO <sub>2</sub>               | 26.74 | 27.00 | 24.66 |  |
| TiO <sub>2</sub>                                           | 0.41  | 0.41  | 0.43  | 0.45  | 0.61  | 0.54  | 0.44  | 0.27  | Al <sub>2</sub> O <sub>3</sub> | 19.58 | 20.04  | 23.42 | Al <sub>2</sub> O <sub>3</sub>              | 21.88 | 23.24 | 25.29 | Al <sub>2</sub> O <sub>3</sub> | 0.07   | 0.06  | Al <sub>2</sub> O <sub>3</sub> | 0.04  | 0.14  | Al <sub>2</sub> O <sub>3</sub> | 0.01  | 0.05  | 0.06  |  |
| Al <sub>2</sub> O <sub>3</sub>                             | 16.22 | 15.46 | 14.29 | 15.63 | 15.70 | 16.16 | 16.56 | 11.11 | TiO <sub>2</sub>               | 0.00  | 0.00   | 0.02  | TiO <sub>2</sub>                            | 0.11  | 0.07  | 0.12  | TiO <sub>2</sub>               | 21.48  | 20.92 | TiO <sub>2</sub>               | 31.27 | 39.29 | TiO <sub>2</sub>               | 18.94 | 20.92 | 22.19 |  |
| Fe <sub>2</sub> O <sub>3</sub>                             | 3.51  | 4.42  | 7.10  | 6.65  | 3.07  | 6.71  | 2.72  | 6.60  | FeO <sup>+</sup>               | 0.16  | 0.26   | 0.00  | Fe <sub>2</sub> O <sub>3</sub> <sup>+</sup> | 12.91 | 11.38 | 9.58  | FeO <sup>+</sup>               | 27.82  | 26.82 | FeO <sup>+</sup>               | 2.13  | 0.83  | FeO <sup>+</sup>               | 21.28 | 22.02 | 19.95 |  |
| FeO                                                        | 11.80 | 12.14 | 6.27  | 6.70  | 10.51 | 9.25  | 12.79 | 7.10  | MnO                            | 0.00  | 0.00   | 0.02  | MnO                                         | 0.39  | 0.33  | 0.20  | MnO                            | 1.36   | 0.77  | MnO                            | 0.00  | 0.00  | MnO                            | 0.43  | 0.00  | 0.37  |  |
| MnO                                                        | 0.00  | 0.00  | 0.50  | 0.49  | 0.13  | 0.08  | 0.07  | 0.16  | MgO                            | 0.00  | 0.00   | 0.01  | MgO                                         | 0.01  | 0.01  | 0.04  | MgO                            | 5.00   | 7.00  | MgO                            | 2.38  | 0.12  | MgO                            | 18.98 | 17.76 | 18.54 |  |
| MgO                                                        | 9.44  | 8.95  | 11.66 | 10.95 | 10.27 | 9.73  | 9.21  | 13.15 | CaO                            | 0.13  | 0.33   | 4.97  | CaO                                         | 23.75 | 22.14 | 23.16 | CaO                            | 5.93   | 4.30  | CaO                            | 0.03  | 0.84  | CaO                            | 0.00  | 0.00  | 0.00  |  |
| CaO                                                        | 10.32 | 10.04 | 8.95  | 9.07  | 10.74 | 10.30 | 11.35 | 11.41 | Na <sub>2</sub> O              | 10.94 | 11.00  | 8.29  | Na <sub>2</sub> O                           | 0.00  | 0.02  | 0.02  | Na <sub>2</sub> O              | 0.02   | 0.02  | Na <sub>2</sub> O              | 0.82  | 6.24  | Na <sub>2</sub> O              | 0.02  | 0.00  | 0.01  |  |
| Na <sub>2</sub> O                                          | 3.30  | 3.18  | 3.77  | 3.56  | 2.38  | 2.91  | 2.91  | 1.61  | K <sub>2</sub> O               | 0.05  | 0.03   | 0.06  | K <sub>2</sub> O                            | 0.00  | 0.01  | 0.00  | K <sub>2</sub> O               | 0.00   | 0.04  | K <sub>2</sub> O               | 7.95  | 0.59  | K <sub>2</sub> O               | 0.00  | 0.00  | 0.00  |  |
| K <sub>2</sub> O                                           | 0.05  | 0.10  | 0.11  | 0.16  | 0.20  | 0.00  | 0.00  | 0.34  | Total                          | 99.63 | 100.10 | 99.05 | Total                                       | 96.41 | 95.80 | 97.68 | Cr <sub>2</sub> O <sub>3</sub> | 0.00   | 0.00  | Total                          | 92.98 | 95.52 | Total                          | 86.40 | 87.74 | 85.78 |  |
| Total                                                      | 97.98 | 97.62 | 97.79 | 97.62 | 97.75 | 97.74 | 97.37 | 97.49 |                                |       |        |       |                                             |       |       |       | Total                          | 100.24 | 99.18 |                                |       |       |                                |       |       |       |  |
| Si                                                         | 6.28  | 6.32  | 6.44  | 6.35  | 6.40  | 6.38  | 6.15  | 6.14  | Si                             | 3.01  | 2.98   | 2.78  | Si                                          | 3.03  | 3.11  | 3.08  | Si                             | 3.02   | 3.06  | Si                             | 6.52  | 6.04  | Si                             | 5.61  | 5.56  | 5.18  |  |
| Al(IV)                                                     | 1.72  | 1.68  | 1.56  | 1.65  | 1.60  | 1.62  | 1.83  | 1.86  | Ti                             | 0.00  | 0.00   | 0.00  | Al                                          | 2.09  | 2.21  | 2.34  | Al                             | 2.00   | 1.92  | Al                             | 1.48  | 5.89  | Al                             | 4.68  | 5.08  | 5.49  |  |
| Al(VI)                                                     | 1.07  | 1.00  | 0.87  | 1.01  | 1.09  | 1.08  | 0.94  | 1.03  | Al                             | 1.01  | 1.03   | 1.23  | Ti                                          | 0.01  | 0.00  | 0.01  | Cr                             | 0.00   | 0.00  | Ti                             | 0.00  | 0.01  | Ti                             | 0.00  | 0.01  | 0.01  |  |
| Ti                                                         | 0.05  | 0.05  | 0.05  | 0.05  | 0.07  | 0.07  | 0.06  | 0.05  | Fe <sup>2+</sup>               | 0.01  | 0.01   | 0.00  | Fe <sup>3++</sup>                           | 0.79  | 0.69  | 0.57  | Ti                             | 0.00   | 0.00  | Fe <sup>2+</sup>               | 0.24  | 0.09  | Fe <sup>2+</sup>               | 3.73  | 3.79  | 3.51  |  |
| Fe <sup>3+</sup>                                           | 0.39  | 0.49  | 0.77  | 0.72  | 0.34  | 0.85  | 0.74  | 0.30  | Mn                             | 0.00  | 0.00   | 0.00  | Mn                                          | 0.03  | 0.02  | 0.01  | Fe <sup>2+</sup>               | 1.82   | 1.75  | Mn                             | 0.00  | 0.00  | Mn                             | 0.08  | 0.00  | 0.07  |  |
| Fe <sup>2+</sup>                                           | 1.44  | 1.50  | 0.76  | 0.81  | 1.27  | 0.74  | 1.13  | 1.58  | Mg                             | 0.00  | 0.00   | 0.24  | Mg                                          | 0.00  | 0.00  | 0.00  | Mn                             | 0.09   | 0.05  | Mg                             | 0.48  | 0.02  | Mg                             | 5.94  | 5.45  | 5.81  |  |
| Mn                                                         | 0.00  | 0.00  | 0.06  | 0.06  | 0.02  | 0.00  | 0.01  | 0.01  | Ca                             | 0.01  | 0.02   | 0.72  | Ca                                          | 2.07  | 1.91  | 1.95  | Mg                             | 0.58   | 0.81  | Ca                             | 0.00  | 0.11  | Ca                             | 0.00  | 0.00  | 0.00  |  |
| Mg                                                         | 2.06  | 1.97  | 2.50  | 2.36  | 2.22  | 2.26  | 2.12  | 2.03  | Na                             | 0.93  | 0.93   | 0.00  | Na                                          | 0.00  | 0.00  | 0.00  | Ca                             | 0.50   | 0.36  | Na                             | 0.21  | 1.54  | Na                             | 0.01  | 0.00  | 0.00  |  |
| Ca                                                         | 1.62  | 1.58  | 1.38  | 1.40  | 1.67  | 1.42  | 1.61  | 1.80  | K                              | 0.00  | 0.00   | 4.97  | K                                           | 0.00  | 0.00  | 0.00  | Na                             | 0.00   | 0.00  | K                              | 1.37  | 0.10  | K                              | 0.00  | 0.00  | 0.00  |  |
| Na(M4)                                                     | 0.38  | 0.42  | 0.62  | 0.60  | 0.33  | 0.58  | 0.39  | 0.20  | Total                          | 4.96  | 4.97   | 4.97  | Total                                       | 8.02  | 7.94  | 7.96  | K                              | 0.00   | 0.00  | Total                          | 13.79 | 13.81 | Total                          | 20.05 | 19.89 | 20.06 |  |
| Na(A)                                                      | 0.55  | 0.49  | 0.43  | 0.40  | 0.34  | 0.10  | 0.44  | 0.63  |                                |       |        |       |                                             |       |       |       | Total                          | 8.01   | 7.97  |                                |       |       |                                |       |       |       |  |
| K                                                          | 0.01  | 0.02  | 0.02  | 0.03  | 0.04  | 0.03  | 0.00  | 0.00  | XAn                            | 0.01  | 0.02   | 0.25  | XFe <sup>3+</sup>                           | 0.27  | 0.24  | 0.19  |                                |        |       | Ms                             | 86.2  | 5.4   | *Total Fe as FeO               |       |       |       |  |
| Total                                                      | 15.56 | 15.51 | 15.43 | 15.43 | 15.37 | 15.14 | 15.44 | 15.63 | XAb                            | 0.99  | 0.98   | 0.75  | *Total Fe as Fe <sub>2</sub> O <sub>3</sub> |       |       |       | Alm                            | 60.9   | 58.8  | Pg                             | 13.5  | 88.0  |                                |       |       |       |  |
|                                                            |       |       |       |       |       |       |       |       | XOr                            | 0.00  | 0.00   | 0.00  |                                             |       |       |       | Prp                            | 19.5   | 27.4  | Mrg                            | 0.3   | 6.5   |                                |       |       |       |  |
| XFe <sup>3+</sup>                                          | 0.27  | 0.33  | 0.47  | 0.42  | 0.24  | 0.44  | 0.44  | 0.23  | *Total Fe as FeO               |       |        |       |                                             |       |       |       | Grs                            | 16.6   | 12.1  | *Total Fe as FeO               |       |       |                                |       |       |       |  |
| XMg                                                        | 0.59  | 0.57  | 0.77  | 0.74  | 0.64  | 0.65  | 0.65  | 0.56  |                                |       |        |       |                                             |       |       |       | Sps                            | 3.0    | 1.7   |                                |       |       |                                |       |       |       |  |
|                                                            |       |       |       |       |       |       |       |       |                                |       |        |       |                                             |       |       |       | *Total Fe as FeO               |        |       |                                |       |       |                                |       |       |       |  |

Table 4

Representative chemical analyses of plagioclase/K-feldspar (8-oxygen basis), epidote (12.5 oxygen basis), garnet (12-oxygen basis), and white mica (22-oxygen basis) for four MMC mica schist/gneiss samples

| Plagioclase                    |         |          |          | Epidote                                     |         |          |          | Garnet                         |          |          | White mica                     |         |          |          | Chlorite                       |          |          | K-feldspar                     |         |
|--------------------------------|---------|----------|----------|---------------------------------------------|---------|----------|----------|--------------------------------|----------|----------|--------------------------------|---------|----------|----------|--------------------------------|----------|----------|--------------------------------|---------|
| Zone                           | Ms-zone | Grt-zone | Grt-zone | Zone                                        | Ms-zone | Grt-zone | Grt-zone | Zone                           | Grt-zone | Grt-zone | Zone                           | Ms-zone | Grt-zone | Grt-zone | Zone                           | Grt-zone | Grt-zone | Zone                           | Ms-zone |
| Sample no.                     | NWS-942 | NWS-38   | NWS-522  | Sample no.                                  | NWS-942 | NWS-38   | NWS-522  | Sample no.                     | NWS-38   | NWS-522  | Sample no.                     | NWS-942 | NWS-38   | NWS-522  | Sample no.                     | NWS-520  | NWS-520  | Sample no.                     | NWS-522 |
| Point no.                      | 4       | 4        | 21       | Point no.                                   | 1       | 3        | 3        | Point no.                      | 1        | 12       | Point No                       | 2       | 2        | 2        | Point no.                      | 10       | 11       | Point no.                      | 1       |
| SiO <sub>2</sub>               | 68.78   | 69.14    | 69.34    | SiO <sub>2</sub>                            | 38.62   | 39.05    | 34.14    | SiO <sub>2</sub>               | 38.77    | 38.75    | SiO <sub>2</sub>               | 51.48   | 49.34    | 50.92    | SiO <sub>2</sub>               | 24.66    | 24.72    | SiO <sub>2</sub>               | 65.22   |
| Al <sub>2</sub> O <sub>3</sub> | 20.17   | 19.67    | 20.44    | Al <sub>2</sub> O <sub>3</sub>              | 25.36   | 26.31    | 21.46    | Al <sub>2</sub> O <sub>3</sub> | 0.06     | 0.06     | Al <sub>2</sub> O <sub>3</sub> | 0.26    | 0.25     | 0.52     | Al <sub>2</sub> O <sub>3</sub> | 0.00     | 0.00     | Al <sub>2</sub> O <sub>3</sub> | 18.57   |
| TiO <sub>2</sub>               | 0.00    | 0.01     | 0.00     | TiO <sub>2</sub>                            | 0.21    | 0.02     | 0.06     | TiO <sub>2</sub>               | 19.95    | 21.21    | TiO <sub>2</sub>               | 28.57   | 27.87    | 28.75    | TiO <sub>2</sub>               | 21.33    | 21.30    | TiO <sub>2</sub>               | 0.07    |
| FeO                            | 0.00    | 0.01     | 0.01     | Fe <sub>2</sub> O <sub>3</sub> *            | 9.76    | 7.54     | 9.23     | FeO*                           | 13.79    | 19.17    | FeO*                           | 2.08    | 4.25     | 5.05     | FeO*                           | 32.50    | 33.68    | FeO*                           | 0.00    |
| MnO                            | 0.00    | 0.00     | 0.01     | MnO                                         | 0.25    | 0.15     | 0.04     | MnO                            | 6.95     | 2.25     | MnO                            | 0.02    | 0.12     | 0.04     | MnO                            | 0.02     | 0.01     | MnO                            | 0.00    |
| MgO                            | 0.01    | 0.01     | 0.00     | MgO                                         | 0.05    | 0.03     | 0.19     | MgO                            | 0.35     | 0.43     | MgO                            | 3.10    | 1.85     | 2.07     | MgO                            | 9.54     | 8.36     | MgO                            | 0.00    |
| CaO                            | 1.03    | 0.15     | 0.67     | CaO                                         | 23.71   | 23.93    | 16.78    | CaO                            | 18.14    | 18.10    | CaO                            | 0.00    | 0.00     | 0.00     | CaO                            | 0.01     | 0.02     | CaO                            | 0.00    |
| Na <sub>2</sub> O              | 11.04   | 11.53    | 10.85    | Na <sub>2</sub> O                           | 0.03    | 0.05     | 0.03     | Na <sub>2</sub> O              | 0.04     | 0.00     | Na <sub>2</sub> O              | 0.20    | 0.08     | 0.18     | Na <sub>2</sub> O              | 0.01     | 0.00     | Na <sub>2</sub> O              | 0.69    |
| K <sub>2</sub> O               | 0.05    | 0.09     | 0.17     | K <sub>2</sub> O                            | 0.00    | 0.01     | 0.00     | K <sub>2</sub> O               | 0.00     | 0.01     | K <sub>2</sub> O               | 6.99    | 10.43    | 7.03     | K <sub>2</sub> O               | 0.00     | 0.01     | K <sub>2</sub> O               | 14.29   |
| Total                          | 101.08  | 100.59   | 101.49   | Total                                       | 97.99   | 97.08    | 81.91    | Cr <sub>2</sub> O <sub>3</sub> | 0.02     | 0.00     | Total                          | 92.71   | 94.20    | 94.56    | Total                          | 88.06    | 88.10    | Total                          | 98.84   |
|                                |         |          |          |                                             |         |          |          | Total                          | 98.07    | 99.98    |                                |         |          |          |                                |          |          |                                |         |
| Si                             | 2.97    | 3.00     | 2.98     | Si                                          | 3.03    | 3.07     | 3.16     | Si                             | 3.09     | 3.04     | Si                             | 6.87    | 6.72     | 6.77     | Si                             | 5.38     | 5.42     | Si                             | 3.01    |
| Ti                             | 0.00    | 0.00     | 0.00     | Ti                                          | 2.35    | 2.44     | 2.34     | Al                             | 1.88     | 1.96     | Ti                             | 4.49    | 4.48     | 4.50     | Ti                             | 5.48     | 5.50     | Ti                             | 0.00    |
| Al                             | 1.03    | 1.01     | 1.03     | Al                                          | 0.01    | 0.00     | 0.00     | Cr                             | 0.00     | 0.00     | Al                             | 0.03    | 0.03     | 0.05     | Al                             | 0.00     | 0.00     | Al                             | 1.01    |
| Fe <sup>2+</sup>               | 0.00    | 0.00     | 0.00     | Fe <sup>3+</sup>                            | 0.58    | 0.45     | 0.64     | Ti                             | 0.00     | 0.00     | Fe <sup>2+</sup>               | 0.23    | 0.48     | 0.56     | Fe <sup>2+</sup>               | 5.92     | 6.17     | Fe <sup>2+</sup>               | 0.00    |
| Mn                             | 0.00    | 0.00     | 0.00     | Mn                                          | 0.02    | 0.01     | 0.00     | Fe <sup>2+</sup>               | 0.92     | 1.26     | Mn                             | 0.00    | 0.01     | 0.00     | Mn                             | 0.00     | 0.00     | Mn                             | 0.00    |
| Mg                             | 0.00    | 0.00     | 0.00     | Mg                                          | 0.01    | 0.00     | 0.03     | Mn                             | 0.47     | 0.15     | Mg                             | 0.62    | 0.38     | 0.41     | Mg                             | 3.10     | 2.73     | Mg                             | 0.00    |
| Ca                             | 0.05    | 0.01     | 0.03     | Ca                                          | 2.00    | 2.02     | 1.66     | Mg                             | 0.04     | 0.05     | Ca                             | 0.00    | 0.00     | 0.00     | Ca                             | 0.00     | 0.01     | Ca                             | 0.00    |
| Na                             | 0.93    | 0.97     | 0.90     | Na                                          | 0.00    | 0.01     | 0.01     | Ca                             | 1.55     | 1.52     | Na                             | 0.05    | 0.02     | 0.05     | Na                             | 0.00     | 0.00     | Na                             | 0.06    |
| K                              | 0.00    | 0.00     | 0.01     | K                                           | 0.00    | 0.00     | 0.00     | Na                             | 0.01     | 0.00     | K                              | 1.19    | 1.81     | 1.19     | K                              | 0.00     | 0.00     | K                              | 0.84    |
| Total                          | 4.98    | 4.99     | 4.96     | Total                                       | 7.99    | 7.99     | 7.85     | K                              | 0.00     | 0.00     | Total                          | 13.48   | 13.93    | 13.54    | Total                          | 19.89    | 19.83    | Total                          | 4.93    |
|                                |         |          |          |                                             |         |          |          | Total                          | 7.97     | 7.98     |                                |         |          |          |                                |          |          |                                |         |
| XAn                            | 0.05    | 0.01     | 0.03     | XFe <sup>3+</sup>                           | 0.20    | 0.15     | 0.22     |                                |          |          | Ms                             | 95.8    | 98.8     | 96.2     | *Total Fe as FeO               |          |          | XAn                            | 0.00    |
| XAb                            | 0.95    | 0.99     | 0.96     | *Total Fe as Fe <sub>2</sub> O <sub>3</sub> |         |          |          | Alm                            | 30.9     | 42.2     | Pg                             | 4.2     | 1.2      | 3.8      |                                |          |          | XAb                            | 0.07    |
| XOr                            | 0.00    | 0.01     | 0.01     |                                             |         |          |          | Prp                            | 1.4      | 1.7      | Mrg                            | 0       | 0        | 0        |                                |          |          | XOr                            | 0.93    |
| *Total Fe as FeO               |         |          |          |                                             |         |          |          | Grs                            | 52.0     | 51.1     | *Total Fe as FeO               |         |          |          |                                |          |          | *Total Fe as FeO               |         |
|                                |         |          |          |                                             |         |          |          | Sps                            | 15.8     | 5.0      |                                |         |          |          |                                |          |          |                                |         |
|                                |         |          |          |                                             |         |          |          | *Total Fe as FeO               |          |          |                                |         |          |          |                                |          |          |                                |         |

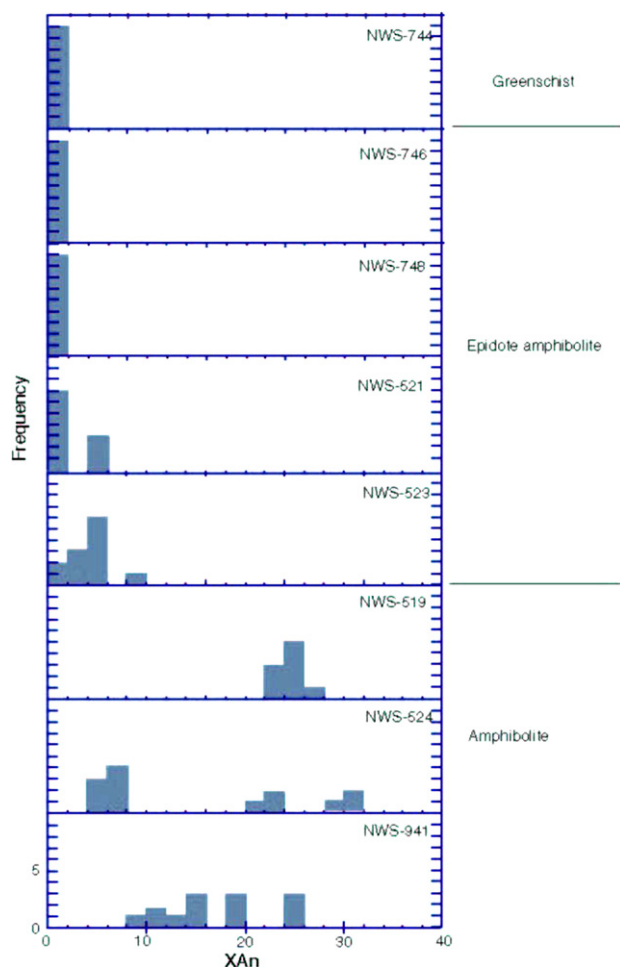


Fig. 8. An content of plagioclase in metabasite samples.

epidote-amphibolite facies rocks. This trend is similar to that shown by meta-basites in other metamorphic terranes (Maruyama et al., 1983).

Chlorite is an ubiquitous mineral in the greenschist facies rocks, and is also present in some epidote-amphibolite facies rocks. It occurs as oriented tabular crystals and prismatic crystals. The modal amount of chlorite is highly variable, with up to 25% in greenschists, between 5–10% in epidote-amphibolites and less than 3% in amphibolites (mostly retrograde). Retrograde chlorite replaces garnet and hornblende in epidote-amphibolite and amphibolite facies rocks. The composition of chlorite in greenschist, epidote-amphibolite and amphibolite meta-basites has been calculated based on 28 oxygen atoms, and assuming all  $\text{Fe}^* = \text{Fe}^{2+}$ . Plotted on Hey's (1954) diagram, it shows a relatively wide range in composition, i.e., ripidolite, pycnochlorite, and clinochlorite.

Garnet porphyroblasts were observed in most amphibolite samples. They are up to 3.0 mm in diameter, and their modes range between 5 and 20%. The garnets are almandine-rich and pyrope-poor (57.7–70.0% and 9.5–29.6%, respectively). There is an increase in pyrope content and decrease in almandine and grossular contents from epidote-amphibolite to amphibolite facies rocks. This trend has also

been observed in many other progressive metamorphic terranes. Most of the garnet crystals display internal (prograde) normal compositional zoning with Ca + Mn-rich cores (bell-shaped Mn-profiles) and Mg-rich rims, whereas the almandine profile is rather flat. Rapid compositional changes occur within the inner three-fourth of the garnet, which contains inclusions; rims are inclusion-free.

White mica is coarse-grained and forms less than 5% of the meta-basite samples that were subjected to EPMA analyses. It has a muscovite composition and in some samples may also have an appreciable paragonite component. The muscovite in epidote-amphibolite rocks has a highly variable range in Si (pfu) (6.48–6.55) and  $\text{Fe}^* + \text{Mg}$  (0.63–0.72), whereas in amphibolite rocks it has both lower Si (pfu) (5.97–6.05) and  $\text{Fe}^* + \text{Mg}$  (0.09–0.54). K-feldspar was observed in a few samples. Its occurrence may be related to protolith composition. Opaque minerals are mostly magnetite and ilmenite, both being up to 0.5 mm in diameter; some ilmenite occurs as inclusions in garnet. Titanite was found as a matrix constituent up to 3% in modal composition in both epidote-amphibolite and amphibolite facies rocks. Rutile was observed in amphibolites only.

## 7.2. Metasediments/granitoids

Garnet occurs as subhedral, generally subequant crystals, up to 1.0 cm in diameter. All of the medium-grade rock samples contain porphyroblastic garnet. Only a few garnet porphyroblasts display snowball or sigmoidal inclusion fabrics due to rotation during crystal growth. The garnets contain inclusions of quartz, and more rarely ilmenite, zoisite, and muscovite. The garnets are almandine-rich and spessartine-poor; grossular and pyrope composition contents vary in each sample, due to differences in protolith types. The garnets do not show any regular compositional zoning. Grade compositions are as follows:  $\text{Alm}_{30-79}$ ,  $\text{Sp}_{0.0-17}$ ,  $\text{Pyr}_{1.4-16}$ ,  $\text{Grs}_{6.2-52}$ .

Mica consists of muscovite and minor biotite. Muscovite defines the primary schistosity in all samples. Subhedral flakes with long dimensions up to 3 mm were observed. Biotite is present in several samples, and apparently formed as a late stage mineral. Muscovite in the garnet zone has lower  $\text{Fe} + \text{Mg}$  values than muscovite in the biotite zone. In sample NWS-520, white mica has a higher paragonite component than in the other four samples.

Zoisite is a significant constituent of the mica schist and gneisses. It has a wide compositional range. In the garnet zone the pistacite component ( $100 \times \text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Al})$ ) is low (12–22), whereas in the biotite zone it varies from 17 to 29. A trend of decreasing pistacite content toward clinozoisite can be observed from low grade to medium grade metasediments. This trend is consistent with that shown by meta-sediments in other metamorphic terranes.

Plagioclase is one of the main rock-forming minerals together with quartz and mica. Most crystals are untwinned, subhedral to anhedral shaped, and are albitic:  $\text{An}_{2-9}$  in the biotite zone (NWS-942 and NWS-943) and  $\text{An}_{1-9}$  in the

garnet zone (NWS-38 and 522). Some plagioclase in NWS-943 has a higher An content (up to An<sub>14</sub>), which may reflect the primary composition of the protolith (meta-granite) plagioclase. In the quartzo-feldspathic rocks there is no regular increase in the An content from the muscovite zone (low grade) to the garnet zone (medium grade).

K-feldspar was observed in only a few samples. Its presence may be related to the protolith composition (feldspar-rich granitoid or arkose). It forms perthitic subhedral to anhedral grains, up to 1.0 mm in size. Some K-feldspar occurs in veinlets closely associated with calcite. Ilmenite and magnetite occur both in the matrix and as inclusions in garnet. Chlorite appears to be a minor constituent, occurring mostly as an alteration product of garnet and muscovite. Quartz is ubiquitous. It occurs as subhedral grains up to 3 mm across with modal amounts ranging between 15% and 40%. Quartz is also a common inclusion phase in garnet and muscovite porphyroblasts.

## 8. Geothermobarometry

P–T (pressure–temperature) estimations for greenschist facies rocks cannot be determined precisely due to the lack of index minerals. The assemblage of Chl + Ab + Zo + Qtz ± Cal in samples analyzed by EPMA represents the lowermost greenschist facies (300 °C to 350 °C, and a maximum of 5 kbar pressure (Spear, 1995)). However, as mentioned above, actinolite has been observed in samples not analysed by EPMA, indicative of temperatures up to 450 °C.

The P–T conditions can be more rigorously determined for the epidote-amphibolite and amphibolite facies rocks because of the presence of suitable geothermobarometers. Three stages of metamorphism can be distinguished in several epidote-amphibolite samples: 1. An early low-temperature stage, which is represented by Qtz + Ab + Ep + Ilm + Mag + Rut + Ms inclusions in garnet, a mineral assemblage that is typical of greenschist facies; 2. Main peak P–T stage (main mineral assemblage); and 3. Late retrograde stage, which is present as intensive Qtz + Ep + Preh + Cal veining. The late stage is probably the equivalent of the prehnite–actinolite facies (Fig. 9).

Geothermobarometry was performed on several garnet–amphibolite epidote-amphibolite and amphibolite facies samples (NWS-746, 748 and NWS-519, 524, 941, respectively; Table 5, Fig. 9). P–T conditions of the main peak stage were obtained by using Fe–Mg partitioning between garnet and hornblende, with calibrations of Graham and Powell (1984) for temperature estimation, and the garnet–hornblende–plagioclase–quartz geobarometer of Kohn and Spear (1989). The two epidote-amphibolite facies samples yielded temperatures ranging between 440 and 522 °C at pressures of 4.8–6.2 kbar. The three amphibolite facies samples show three different sets of P–T conditions, varying from 646 to 671 °C at pressures of 7.5–9.6 kbar. On this basis two samples (NWS-524, 941) fall outside the P–T stability field of the amphibolite facies and reach the granulite field. The estimated P–T conditions for the MMC metabasites match well a progression from early pro-grade

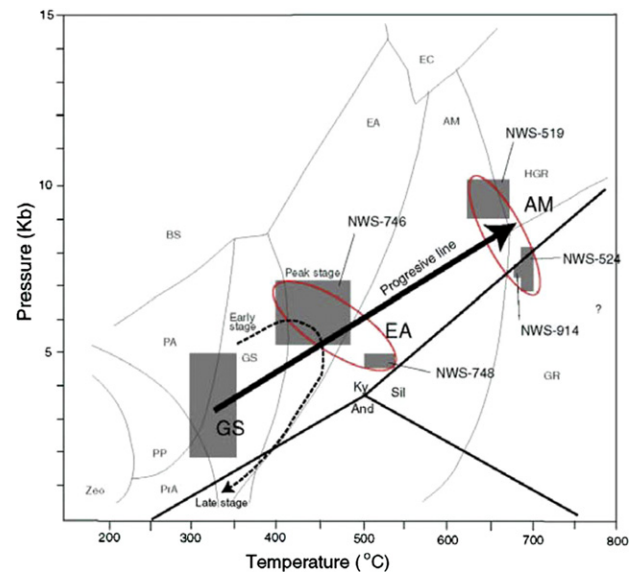


Fig. 9. P–T trajectory of progressive metamorphism in Malino metabasites. P–T petrogenetic grid for facies metamorphism are based on Maruyama et al. (1996). Abbreviations of facies names: BS, blueschist; Zeo, zeolite; PP, prehnite-pumpellyite; PrA, prehnite-actinolite; PA, pumpellyite-actinolite; GS, greenschist; EA, epidote-amphibolite; AM, amphibolite; HGR, high-pressure granulite; GR, granulite; EC, eclogite.

Table 5  
P–T estimation for five MMC amphibolite samples

| Epidote-amphibolite facies |          |         |          | Amphibolite facies |          |         |          |         |          |
|----------------------------|----------|---------|----------|--------------------|----------|---------|----------|---------|----------|
| NWS-748                    |          | NWS-746 |          | NWS-519            |          | NWS-524 |          | NWS-941 |          |
| T (°C)                     | P (kbar) | T (°C)  | P (kbar) | T (°C)             | P (kbar) | T (°C)  | P (kbar) | T (°C)  | P (kbar) |
| 503                        | 4.9      | 390     | 6.4      | 678                | 8.6      | 685     | 7.0      | 672     | 7.6      |
| 539                        | 5.0      | 516     | 4.5      | 649                | 9.5      | 696     | 6.6      | 668     | 7.6      |
| 539                        | 4.6      | 434     | 6.1      | 625                | 9.9      | 704     | 8.1      | 675     | 7.8      |
| 508                        | 4.7      | 415     | 7.0      | 653                | 10.1     | 694     | 8.1      |         |          |
|                            |          | 400     | 7.0      | 625                | 9.9      | 685     | 7.4      |         |          |
|                            |          | 451     | 6.0      |                    |          |         |          |         |          |
|                            |          | 474     | 6.6      |                    |          |         |          |         |          |
| Average                    | 522      | 440     | 6.2      | 646                | 9.6      | 693     | 7.5      | 671     | 7.6      |
| STDEV                      | 19       | 44      | 0.9      | 22                 | 0.6      | 8       | 0.7      | 4       | 0.1      |

through main peak to retrograde stages as suggested by the observed mineral assemblages (Fig. 9).

P–T conditions in the quartz–feldspathic metasediments and metagranitoids cannot be precisely determined, because their mineral assemblages have a wide temperature and pressure stability range. The maximum temperature for the garnet zone is estimated to be 470 °C (Spear, 1995), as this is the minimum temperature for the first appearance of staurolite in the staurolite zone. By using the geothermometer of garnet and muscovite of Hynes and Forest (1988), the garnet zone samples NWS-522 and NWS-38 yield temperatures of 410–440 °C.

## 9. Geochronology

A total of 12 samples were selected for K/Ar and  $^{40}\text{Ar}/^{39}\text{Ar}$  analysis of muscovite, biotite and hornblende separates, and U/Th/Pb analysis of zircons.

### 9.1. K/Ar analyses

The results of K/Ar determinations on muscovite from two mica schist samples are shown in Table 6. NWS-38 yielded a mean age of  $14.4 \pm 0.3$  Ma, and NWS-61 an age of  $11.3 \pm 0.1$  Ma. These ages fall within the upper end of the range of  $^{40}\text{Ar}/^{39}\text{Ar}$  ages (see below).

### 9.2. $^{40}\text{Ar}/^{39}\text{Ar}$ analyses

Six samples were analysed. The analytical results are summarized in Table 7 and illustrated in Fig. 10 as plateau diagrams that show one representative grain analysis per sample. The following account is based on Vasconcelos (2000).

The two muscovite grains analyzed for sample NWS-55, a garnet–muscovite metasediment, yielded two well-defined plateau ages within 2 sigma error from each other ( $19.3 \pm 0.3$  and  $19.7 \pm 0.2$  Ma). The ideogram age for this sample is 19.6 Ma.

The three muscovite grains analyzed for muscovite schist sample NWS-63 yielded three well-defined plateau ages within 2 sigma error from each other ( $14.1 \pm 0.2$ ,  $14.1 \pm 0.2$ , and  $14.7 \pm 0.4$ ). The sharp and well-defined ideogram obtained for this sample gives a 14.1 Ma age, consistent with the plateau ages obtained from the three grains analysed.

Sample NWS-232 is an amphibolite. Five hornblende grains analyzed yielded mostly disturbed spectra with plateau and pseudo-plateau ages ranging from  $17.1 \pm 0.1$  to

$33.0 \pm 1.2$  Ma. The grain that yielded the youngest result also gave the most precise measurements, which is reflected in the size of the ideogram peak. The only two grains that yielded reproducible results gave ages in the 32–33 Ma range. Chloritization of the hornblende crystals may account for the disturbed spectra.

Five biotite and four hornblende grains from sample NWS-521 which, is a meta-siltstone were analyzed. The biotite grains yielded meaningless results (not illustrated in this paper) because they are completely chloritized. The hornblende grains on the other hand yielded reproducible, similar spectra with relatively consistent plateau ages ranging from  $20.7 \pm 1.1$  to  $25.0 \pm 1.7$  Ma. The ideogram shows a peak age of 22.6 Ma.

The four hornblende grains analyzed for amphibolite sample NWS-523 yielded reproducible, similar spectra that show relatively consistent plateau ages varying from  $20.9 \pm 0.5$  to  $24.6 \pm 1.5$  Ma. These results when plotted together indicate an ideogram age of 23.1 Ma for this sample.

Table 7  
Summary results for the  $^{40}\text{Ar}/^{39}\text{Ar}$  analysis of six MMC samples

| Sample  | Material   | Plateau age $\pm$ (Ma) |     | Ideogram age (Ma) |
|---------|------------|------------------------|-----|-------------------|
| NWS55   | Muscovite  | 19.3                   | 0.3 | 19.6              |
|         | Muscovite  | 19.7                   | 0.2 |                   |
| NWS 63  | Muscovite  | 14.1                   | 0.2 | 14.1              |
|         | Muscovite  | 14.2                   | 0.2 |                   |
|         | Muscovite  | 14.7                   | 0.4 |                   |
| NWS 232 | Hornblende | 24.9                   | 1.0 | 17.1, 30.1        |
|         | Hornblende | —                      | —   |                   |
|         | Hornblende | 31.9 <sup>a</sup>      | 1.2 |                   |
|         | Hornblende | 33.0                   | 1.2 |                   |
|         | Hornblende | 17.1                   | 0.1 |                   |
| NWS 521 | Hornblende | 23.0                   | 1.0 | 22.6              |
|         | Hornblende | 22.1                   | 1.3 |                   |
|         | Hornblende | 20.7                   | 1.1 |                   |
|         | Hornblende | 25.0                   | 1.7 |                   |
| NWS 523 | Hornblende | 24.6                   | 1.5 | 23.1              |
|         | Hornblende | 20.9                   | 0.5 |                   |
|         | Hornblende | 23.0                   | 2.0 |                   |
|         | Hornblende | 23.2                   | 0.7 |                   |
| NWS 526 | Hornblende | 11.0                   | 0.4 | 11.1              |
|         | Hornblende | 13.0                   | 1.7 |                   |
|         | Hornblende | 9.30                   | 0.6 |                   |
|         | Hornblende | 10.0                   | 0.3 |                   |
|         | Hornblende | 15.3                   | 0.9 |                   |
|         | Hornblende | 10.0                   | 6.0 |                   |

<sup>a</sup> Pseudo-plateau.

Table 6  
K–Ar data for two mica schist samples from the Malino Metamorphic Complex

| Sample no. | Material  | wt% K | $^{40}\text{Ar} (\times 10^{-11} \text{ mol/g})$ | $^{40}\text{Ar}^*/^{40}\text{Ar}$ total | Age (Ma)       |
|------------|-----------|-------|--------------------------------------------------|-----------------------------------------|----------------|
| NWS 38     | Muscovite | 7.063 | 17.3672                                          | 0.530                                   | $14.1 \pm 0.3$ |
|            |           |       | 18.0496                                          | 0.570                                   | $14.7 \pm 0.3$ |
| NWS 61     | Muscovite | 7.104 | 1.4000                                           | 0.773                                   | $11.3 \pm 1$   |
|            |           | 7.116 |                                                  |                                         |                |

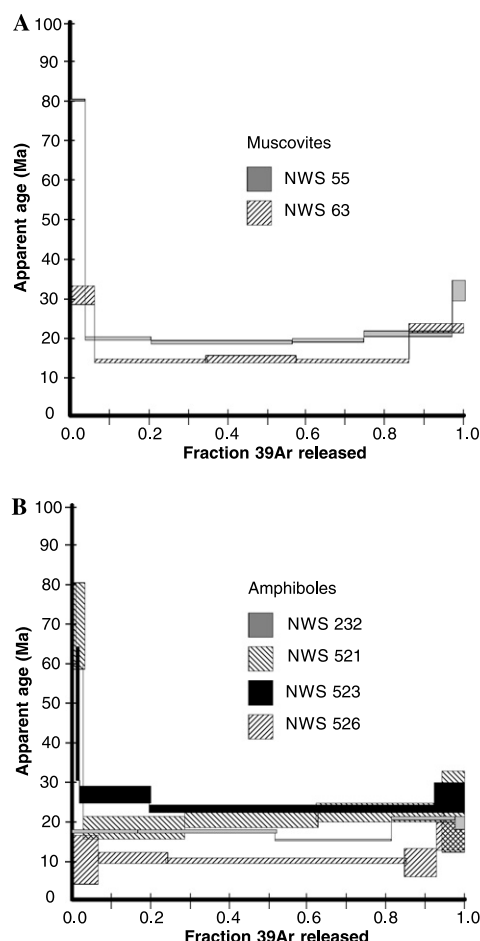


Fig. 10. Selected  $^{40}\text{Ar}/^{39}\text{Ar}$  spectra for muscovites (A) and amphiboles (B) from the Malino Metamorphic Complex.

Six hornblende grains separated from meta-siltstone sample NWS-526 yielded reproducible similar spectra. Five of the grains produce plateau ages within 2 sigma error from each other ( $10.0 \pm 0.3$  to  $13.0 \pm 1.7$  Ma) and an ideogram age of 11.1 Ma. The sixth grain yielded a slightly older age ( $15.3 \pm 0.9$  Ma), possibly reflecting incomplete resetting during metamorphism.

In summary, the  $^{39}\text{Ar}/^{40}\text{Ar}$  and K/Ar ages obtained for eight MMC samples fall within a range of 23–11 Ma, with the exception of an outlier of 30 Ma. These age are interpreted to reflect cooling ages resulting from the exhumation of the metamorphic complex, based on the metamorphic P–T conditions discussed above and the Ar closure temperatures for the respective minerals.

### 9.3. U–Th–Pb Zircon dating

A summary of the analytical results for three samples (NWS-942, NWS-943, and NWS-55) is presented in Table 8.

Sample NWS-942 is a biotite-grade meta-granitoid. Thirty analyses were made of 24 individual zircons and the concordant grains range in age from 331 to 17.5 Ma (Fig. 11). Cathodoluminescence images show moderately narrow rhythmic banding with a few distinct cores, and

Table 8  
Summary of ages interpreted from LA-ICP-MS U/Pb dating of single zircons

*NWS-942*. Biotite grade meta-granitoid. Igneous age =  $321.0 \pm 3.8$  Ma.

Proterozoic inheritance. Metamorphic rim ages of 19.2 and 17.5 Ma.

Rims have very low Th contents, and low Th/U.

*NWS-943*. Biotite grade meta-granitoid. Extensive Pb loss. Igneous age =  $332 \pm 17$  Ma (upper intercept age). Proterozoic inheritance.

Modern (Miocene?) Pb loss (lower intercept =  $33 \pm 84$  Ma).

*NWS-55*. Metasediment. Detrital zircon age peaks at about: 3560, 1820, 1600–1470, 1120, and 400 Ma. There is a discrete large population with an age of  $1543 \pm 29$  Ma (MSWD = 2.28) for 9 of 30 grains analysed.

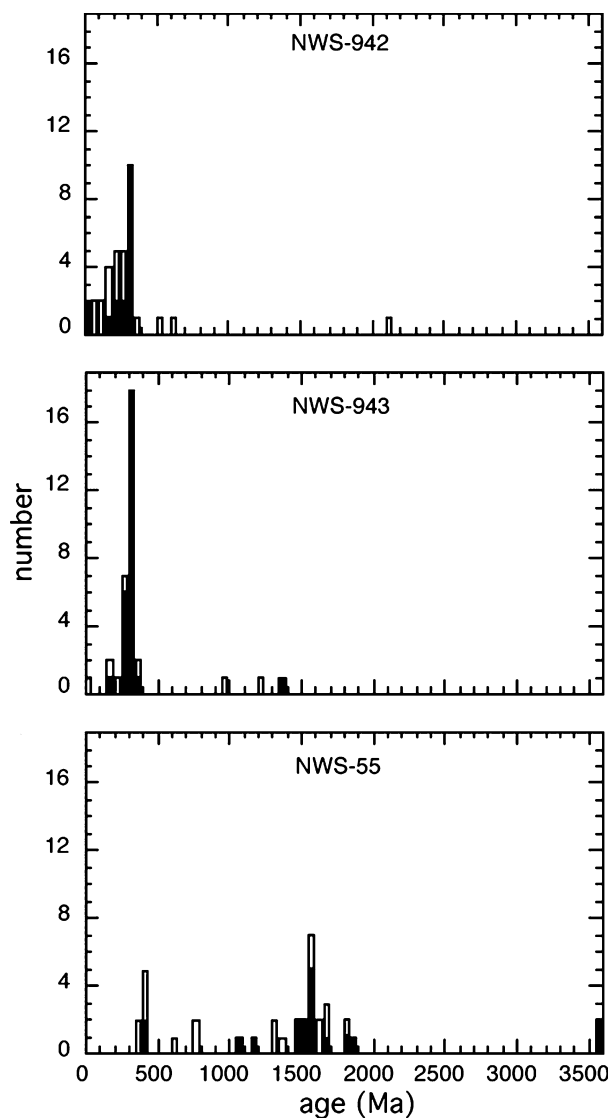


Fig. 11. Zircon age histograms for the three samples from the Malino Metamorphic Complex. Filled bars represent robust age results (see text). Open bars are discordant or noisy age results (see text). Carboniferous ages dominate samples NWS942 and NWS943. NWS55 contains concordant zircons ranging in age from ~3550 to 420 Ma.

some bright rims on somewhat irregularly shaped substrate grains (Fig. 12A). The cathodoluminescence patterns are typical of zircons known to be igneous. Most analyses pierced the reverse sides of the zircons and the resulting

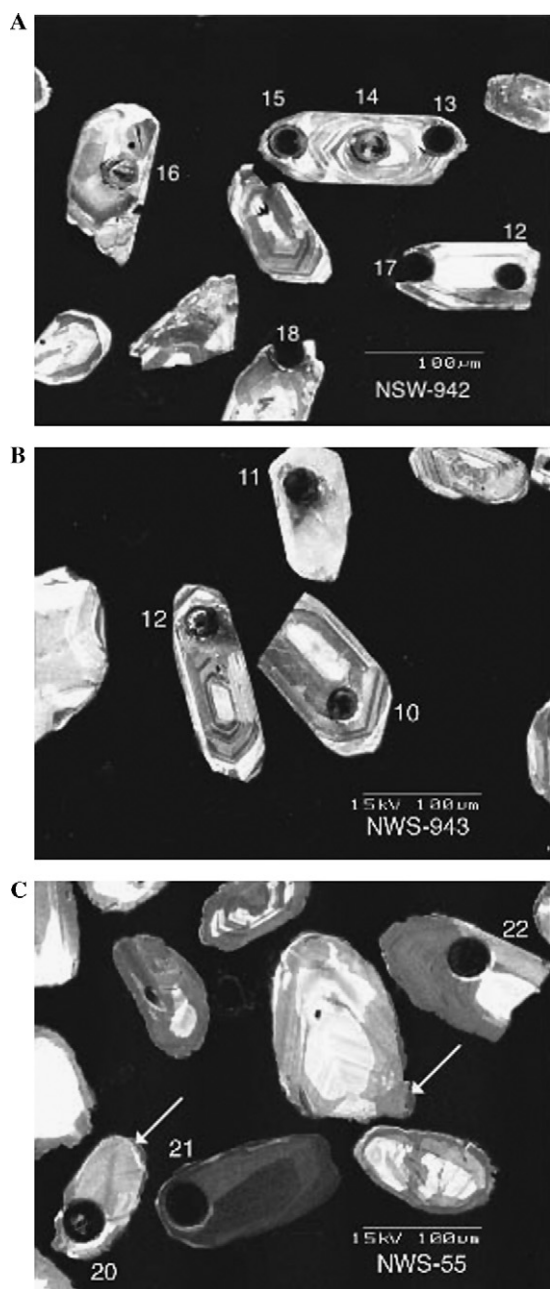


Fig. 12. Cathodoluminescence images at 300 magnification for U–Pb zircon grain mounts. Laser ablated holes are 32  $\mu\text{m}$  across. Generally those ablations that penetrated the zircon appear black, whereas those that ended within the zircon appear brighter (grain containing analysis #16 in A for instance). (A) NWS-942. This sample is dominated by grains with fine concentric banding. (B) NWS-943. Most grains from this sample have concentric banding but others have broad cathodoluminescence zones (grain #11). (C) NWS-55. Zircons from this sample vary from subhedral to subrounded. Cathodoluminescence patterns are more varied. Many grains have fine concentric zoning like # 22. Others have broad zoning (#20 and 21) and some appear to have healed cracks (lower right of figure C). Moreover some of these grains have discontinuous narrow rims, two instances indicated by arrows.

patterns were generally a plateau in  $^{206}\text{Pb}/^{238}\text{U}$  ages at the start of the ablation, followed by a steady age decrease towards the grain margin (Fig. 13). That part of the ablation used for data compilation (plateau) occurs between the

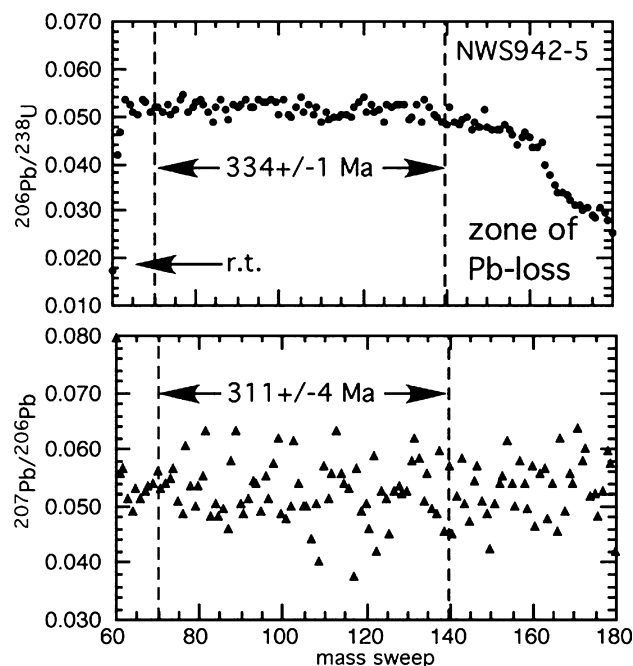


Fig. 13.  $^{206}\text{Pb}/^{238}\text{U}$  and  $^{207}\text{Pb}/^{206}\text{Pb}$  displayed as a function of mass sweep (or depth or time) during ablation of NWS942 grain #5 in which the back of the grain was pierced. The ablation consists of 20 s of background, laser-off, (not pictured) followed by 40 s of laser-on. During the period of 10 mass sweeps called 'rise-time' ('r.t.') the signal stabilizes. In this sample the core of the grain used for age calculation (between dashed lines) returned a concordant Carboniferous age, but the rim of the grain is a zone of Pb-loss as demonstrated by no change in  $^{207}\text{Pb}/^{206}\text{Pb}$  but a decline in  $^{206}\text{Pb}/^{238}\text{U}$ .

vertical dotted lines and in this case corresponds to a robust Carboniferous age. The region used for compilation occurs after 10 mass sweeps of rise-time or 'r.t.' during which the signal stabilizes after the start of ablation, and before a region of Pb-loss at the grain margin. A diagnosis of Pb-loss is straightforward, because whereas the  $^{206}\text{Pb}/^{238}\text{U}$  age is changing in the top panel, the  $^{207}\text{Pb}/^{206}\text{Pb}$  age is constant across the spectrum. Pb-loss does not affect Pb isotope composition.

Twelve of these plateaux yielded a consistent, concordant age of  $315.1 \pm 6.3$  Ma (MSWD = 4.67) but the MSWD is high and we interpret that the three youngest grains suffered some Pb loss making  $321.0 \pm 3.8$  Ma for 9 grains (MSWD = 1.20) the best age for the population. This is interpreted as the time of primary igneous crystallization. Three grains display a continuation of this Pb loss behavior in that they yielded steadily decreasing  $^{206}\text{Pb}/^{238}\text{U}$  ages suggesting continuous age zoning (Pb-loss) at the micron scale over the entire 20  $\mu\text{m}$  of laser drilling. From two grains we can interpret an age and one is displayed in Fig. 14. This grain yields a  $^{207}\text{Pb}/^{206}\text{Pb}$  age of  $428 \pm 8$  Ma across most of the spectrum but the  $^{206}\text{Pb}/^{238}\text{U}$  ratio is changing except for the last 20 mass scans before the back of the zircon was pierced. This short plateau yields a  $^{206}\text{Pb}/^{238}\text{U}$  age of  $19.2 \pm 0.3$  Ma, and  $^{207}\text{Pb}/^{235}\text{U}$  age of  $20.2 \pm 1.4$  Ma. For this interval the  $^{207}\text{Pb}/^{206}\text{Pb}$  age, on average is lower (140 Ma) than for the entire spectrum, but it is

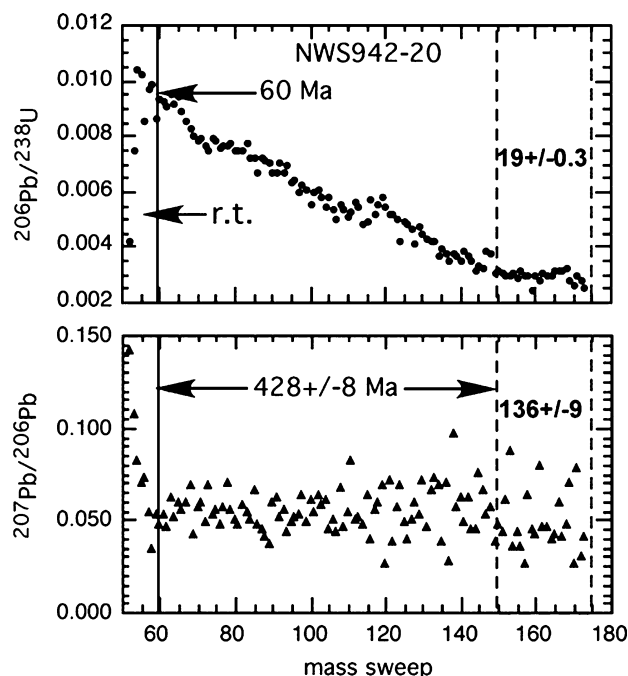


Fig. 14.  $^{206}\text{Pb}/^{238}\text{U}$  and  $^{207}\text{Pb}/^{206}\text{Pb}$  displayed as a function of mass sweep during ablation of NWS942 grain #20 in which the back of the grain was pierced. We interpret this as a possible continuation of the style of zircon behaviour in Fig. 13. Most of the grain is a zone of Pb-loss ( $^{206}\text{Pb}/^{238}\text{U}$  ratio decreases from an age-equivalence of 60 Ma) and the ages interpreted from the two ratios are very different. A zone of new zircon growth  $\sim 19$  Ma in age is interpreted for the last 25 mass sweeps where  $^{206}\text{Pb}/^{238}\text{U}$  forms a plateau between the vertical dashed lines. The  $^{206}\text{Pb}/^{238}\text{U}$  and  $^{206}\text{Pb}/^{238}\text{U}$  ages agree ( $19.1 \pm 0.3$  and  $20.2 \pm 1.4$  Ma, respectively). The weighted mean  $^{207}\text{Pb}/^{206}\text{Pb}$  age of 140 Ma is substantially lower than for the rest of the analysis but does not agree with the U–Pb systems. See text for possible explanations.

much greater than the Miocene U–Pb age. Ideally if this is new zircon growth on a wide Pb-loss mantle, then the grain would be concordant with respect to all three isotope ratios, however it is possible that the young age, the presence of common Pb, and the low count rates on  $^{207}\text{Pb}$  all have prevented a true measurement of the  $^{207}\text{Pb}/^{206}\text{Pb}$  age for this section of the grain. We provisionally interpret this, and a small segment of one other grain (#15 at 17.5 Ma), to be fine rims of new zircon growth that correspond to bright rims in cathodoluminescence. These two rims have atomic Th/U of 0.008 and 0.004 whereas the nine Carboniferous grains used in compiling the igneous age have an average atomic Th/U of  $0.40 \pm 0.11$ . These fine rims have outstandingly low Th contents which is a feature of many metamorphic zircons.

Two grains are strongly discordant with calculated Proterozoic upper intercept  $^{207}\text{Pb}/^{206}\text{Pb}$  ages, the oldest having a minimum age of 2100 Ma and these are interpreted as relict igneous cores of Proterozoic age. Because of the discordance, and behaviour of the zircon (strong age gradients), we can be no more certain than to say that there is a strong probability of Proterozoic precursors that were inherited into the Carboniferous magma. A total of 13 analyses give  $^{206}\text{Pb}/^{238}\text{U}$  ages of post-Carboniferous and pre-Miocene

age and 3 are concordant (250, 236, and 193 Ma) Ma. These ages could represent either mixed ages (zircons zones straddled by the laser pit) or zircons that have suffered Pb loss. Given the prominent evidence for Pb loss in most zircons in this rock, we suggest that the ages intermediate between the Carboniferous igneous age, and the Miocene are the product of Pb loss.

We interpret that these zircons are Carboniferous igneous zircons with some Proterozoic relict cores that were affected by high grade metamorphism, a metamorphism that ended at about 17–19 Ma (exhumation?). Diffusion of Pb in zircon under dry conditions is very slow, and to achieve such broad diffusion zones by a thermal mechanism alone demands a very high temperature ( $>900^\circ\text{C}$ ) and many tens of millions of years (Lee et al., 1997). We favour, instead, fluid-present Pb-loss mechanism that was effective even in a biotite-grade rock.

Sample NWS-943 is a biotite–muscovite gneiss, a meta-igneous rock also of biotite-grade. The zircons are euhedral, subhedral, and broken and the vast majority display compositional banding on a very fine scale (Fig. 12 B). Thirty five spots were analyzed and one failed to give an age solution. Twenty-seven analyses are concordant and range in age from 1372 to 151 Ma however for these extremes, the rest of the concordant points have ages between 361 and 272 Ma (Fig. 11). This continuous range of ages cannot be a single igneous age population (MSWD = 37.5) and we suggest that the range is the product of Pb loss from a Carboniferous population. The kind of zircon behavior shown in Fig. 13 is also displayed in many of the ablations from this sample. If a discordia is fitted to all ablations less than 360 Ma, the probability of fit is 0, but if eight grains that appear to lie to the right of the discordia are removed the upper intercept is  $332 \pm 17$  Ma and the lower is  $33 \pm 84$  Ma (Fig. 15; MSWD = 1.9, Isoplot, Ludwig, 2003). This upper intercept overlaps in age with that of sample NWS-942. All zircons have Th/U of between 0.1 and 1 except two very low U grains (Th/U > 1) and one young grain (Th/U = 0.02). There is no concrete evidence of new zircon growth. We interpret this as a Carboniferous rock that has suffered modern (Miocene?) Pb loss and that the zircon cores that range in age from approximately 1200–1550 Ma represent an inheritance in the igneous rock. Note that only one of the three inherited cores is concordant at the 5% level ( $1372 \pm 28$  Ma).

Sample NWS-55 is garnet–muscovite schist that yielded a 19.6 Ma muscovite Ar–Ar age. It is a metamorphosed clastic sedimentary rock containing concordant detrital zircons that range in age from Archaean to Siluro-Devonian (Fig. 11). The zircons range in shape from euhedral to sub-rounded. Some have tight concentric cathodoluminescence zoning patterns but most have distinct broad cores; some have terminated structures and substantial rims (Fig. 12C). Thin rims were observed on some grains, but they are not ubiquitous, as one might expect if metamorphic growth of zircon had occurred in the Miocene. Of the 30 analysed zircons, about half of them are concordant. The concordant

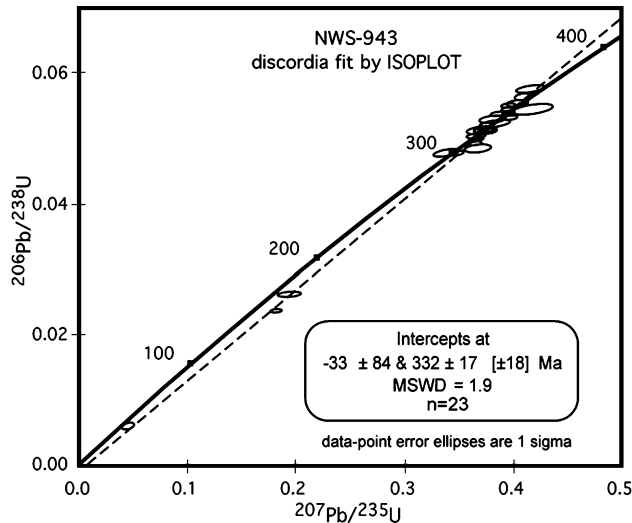


Fig. 15. Discordia plot for all grains <360 Ma from sample NWS-943 except #8 which suggest a discordia parallel to that displayed. These 23 points suggest a modern Pb loss event that, statistically, could be as old as Mesozoic. The upper intercept of  $332 \pm 17$  is interpreted as igneous crystallization of zircon.

grains indicate five populations:  $\sim 3560$ , 1820, 1600–1470 (with a dominant peak at 1550), 1120, and  $\sim 400$  Ma (Fig. 11). Most discordant grains lie between the dominant  $\sim 1550$  group and the  $\sim 400$  group. A discordia fit to all grains in this age interval save one discordant one, and two concordant  $\sim 1100$  Ma grains gave an upper intercept of  $1554 \pm 23$  and lower intercept of  $423 \pm 38$  Ma (Fig. 16; MSWD = 14; ISOPLOT) suggesting Pb loss in the Siluro-Devonian from a 1550 Ma population. A first interpretation is that the young grains that cluster at the low end of

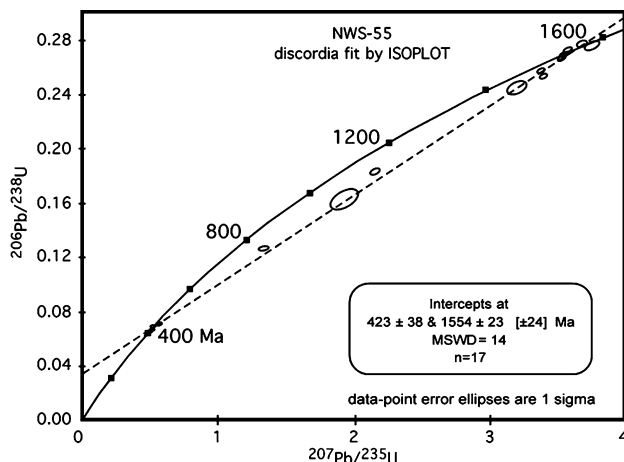


Fig. 16. Discordia fit to all zircons from sample NWS-55 less than 1700 Ma save two concordant grains about 1100 Ma in age, and one discordant one that clearly does not fall on the discordia. This interpretation indicates that the grains are somewhat younger than 1550 Ma, and the clearly discordant ones could be the result of Pb loss in the Devonian. Although the Devonian ages are used to construct the discordia, they are interpreted not as the result of Pb-loss, but of new zircon growth (see Fig. 14). Without the Devonian grains, the lower intercept for the remaining 11 is  $530 \pm 100$  Ma.

discordia lost all of their Pb at  $\sim 420$  Ma. Zircon textures do not support this idea nor do zircon age relationships like that displayed in Fig. 17. In the ablation of this relatively large grain, the young rim (start) is concordant (area between vertical bars). Note that the  $^{206}\text{Pb}/^{238}\text{U}$  and  $^{207}\text{Pb}/^{206}\text{Pb}$  roughly track one another but whereas the end of the ablation occurs in a Mesoproterozoic-aged ( $\sim 1280$  Ma)  $^{207}\text{Pb}/^{206}\text{Pb}$  plateau, the Neoproterozoic-aged  $^{206}\text{Pb}/^{238}\text{U}$  continues to rise. Interpretation of this single ablation indicates that the Devonian rim is new zircon growth over a diffusion zone on a Mesoproterozoic core ( $\sim 1280$  Ma) that cannot be strictly dated in this ablation, because truly concordant core zircon was not penetrated. Based on observations like this, there are 6 analyses with concordant  $^{206}\text{Pb}/^{238}\text{U}$  ages between 390 and 442 Ma that are new igneous zircon growth. Only the oldest 3 of the 6 are robust; the younger zircons have observed uncertainties more than twice that expected. The 3 robust grains give an age of  $428 \pm 14$  Ma and MSWD of 3.11. More work is needed to gain a precise age of this Devonian event. Because this new zircon growth is about that of the Pb loss event ( $530 \pm 100$  Ma if only pre-Palaeozoic grains are considered in fitting a discordia), we interpret that the zircons in this clastic rock are dominated by zircons from igneous (volcanic?) rocks of Devonian age that had zircon-rich sources, or that assimilated continental crust. The Pb loss event for the Mesoproterozoic zircons was heating and incorporation

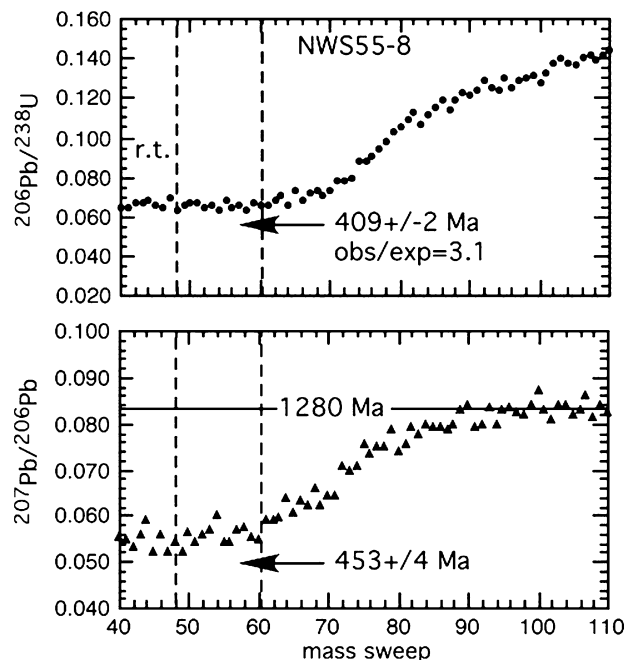


Fig. 17.  $^{206}\text{Pb}/^{238}\text{U}$  and  $^{207}\text{Pb}/^{206}\text{Pb}$  displayed as a function of mass sweep during ablation of NWS55 grain #8. In this grain, a rim was encountered before the core was ablated. The agreement of the two ages for the zone used for age calculation suggests a Devonian rim of new zircon growth on a Pb-loss zone mantling a Mesoproterozoic core, although a concordant core was not encountered and a rigorous core age cannot be calculated. The calculated age of 409 Ma, for this ablation is not considered robust although it is concordant due to the observed uncertainty in  $^{206}\text{Pb}/^{238}\text{U}$  being more than twice that expected from counting statistics.

Table 9

Salient features of the MMC (this study) and basement of the Kemum Terrane, Bird's Head (Pieters et al., 1983; Bladon, 1988; Pigram and Sukanta, 1988)

|                            | Malino Metamorphic Complex                                                      | Basement Kemum Terrane                                                                    |
|----------------------------|---------------------------------------------------------------------------------|-------------------------------------------------------------------------------------------|
| Protoliths                 | Predominantly quartz-rich sediments intruded by granitoids                      | Pelitic and quartz-rich sediments (distal and proximal turbidites) intruded by granitoids |
| Age sedimentary protoliths | (At least in part) Devonian                                                     | Silurian – Devonian                                                                       |
| Age intrusive rocks        | $321 \pm 3.4$ Ma (U/Pb zircon dating)                                           | 324 Ma (K/Ar dating)                                                                      |
| Age of protosource rocks   | Inherited zircons predominantly Proterozoic, including an age of $\sim 1280$ Ma | Granite pebble dated at 1255 Ma by K/Ar method                                            |

into magma that produced new zircon growth. The youngest robust zircon age from NWS-55 is  $417 \pm 7$  Ma and this is a maximum age of sediment deposition.

The largest concordant zircon population in NWS-55 comprises 12 grains, 9 of which are concordant at better than the 5% level, yielding a  $^{208}\text{Pb}$ -corrected  $^{207}\text{Pb}/^{206}\text{Pb}$  age of  $1543 \pm 29$  Ma (MSWD = 2.28) corresponding to the upper intercept of Fig. 16. The two Archaean grains are both well-behaved and have overlapping  $^{208}\text{Pb}$ -corrected  $^{207}\text{Pb}/^{206}\text{Pb}$  ages of  $3550 \pm 38$  and  $3585 \pm 38$  Ma. The full ablation intervals were used to calculate these ages and thus we can make no inferences about their relation to younger geologic events. The same is the case for grains with  $^{208}\text{Pb}$ -corrected  $^{207}\text{Pb}/^{206}\text{Pb}$  ages of  $1856 \pm 30$ ,  $1838 \pm 28$ ,  $1158 \pm 20$ , and  $1091 \pm 18$  Ma, none of which gave indication of age zoning in their analyses.

We interpret this clastic rock to be comprised of Devonian, 1100, 1550, 1840, and 3560 Ma zircons. The pre-Devonian zircon population (Fig. 11) has greatest regional significance because zircon age populations like these occur on the Australian craton.

## 10. Discussion

The mica-bearing quartzo–feldspathic schists and gneisses, which form the bulk of the MMC, are predominantly derived from sedimentary rocks. The dominance of muscovite over biotite suggests that Mg–Fe rich pelitic material was not a dominant component. We interpret the sedimentary sequence to have been composed of arkoses and greywackes (? proximal turbidites) with intercalations of quartzose sandstone, mudstone limestone and volcanics. Some of these rocks originally contained appreciable amounts of carbonaceous material, now converted into graphite. The quartz-rich nature of the protholiths indicates that the succession was deposited on or adjacent to continental crust, most likely in a continental shelf environment. When and where are explored in the discussion below.

The sedimentary formations were intruded by granitoids during the Early Carboniferous ( $321 \pm 3.8$  Ma) as represented by samples NWS-942 and 943. The youngest detrital zircon in a low-grade meta-psammite sample NWS-55 is  $417 \pm 4$  Ma, constraining the maximum depositional age of this and related rocks to be Devonian. Thus the limited zircon dating results indicate that the MMC is, at least in part, of Devonian to Early Carboniferous age. Ar–Ar dating, and U–Pb dates on rims of new zircon (sample NWS-942),

indicate metamorphism and exhumation of these rocks in the Miocene.

The igneous zircon ages clearly demonstrate that the complex is an allochthonous terrane as the region that is now Sulawesi started to form only in the Cretaceous. The presence of inherited Proterozoic and Archaean zircons suggests that the terrane had its origin probably in the New Guinea–Australian region. Further support for a New Guinea–Australian origin of the Malino meta-sediments and meta-granitoids or their protosources is provided by their Sr and Nd isotope characteristics, which resemble those of northern Australian river sediments, and by mantle extraction model ages between 1500 and 2000 Ma. The MMC has several features in common with the basement rocks of the Kemum Terrane in the Bird's Head, Irian Jaya (Fig. 1B), including mid-Carboniferous intrusive activity (Table 9), suggesting derivation of the MMC from rocks originally adjacent to the Bird's Head. A whole rock K/Ar age of  $398 \pm 30$  Ma obtained from a gabbro suggests that the Bird's Head region may also have witnessed a late Silurian–early Devonian magmatic event (i.e., a possible source for the detrital Devonian igneous zircons in the MMC), but the low potassium content of the sample suggests that this age may not be reliable (Bladon, 1988). Regional tectonic models developed by Pigram et al. (1985) and Struckmeyer et al. (1993) assumed that the Bird's Head was derived from Papua New Guinea or eastern Australia, whereas the model of Hill and Hall (2003) assumed a relatively fixed position of the Bird's Head relative to Australia.

The presence of inherited Proterozoic and Archaean zircons in the MMC may indicate that the complex is underlain by Precambrian basement, as may be also the case for parts of Papua New Guinea (Baldwin and Ireland, 1995; Munroe and Williams, 1996). This would suggest that the MMC and Bird's Head may be an accreted exotic Precambrian terrane derived from beyond the Australian craton. Alternatively, during the Paleozoic, the MMC and Bird's Head may have been located relatively close to Australia, perhaps in a continental margin setting, and were receiving sediments derived from one or more of the Precambrian basement areas. To test the latter scenario we compared the provenance ages of detrital zircon within the MMC with the ages of the major basement areas in Australia (cf. Pell et al., 1997). This identified the Archaean Pilbara Block and Proterozoic Paterson Province in Western Australia as potential source areas for the sedimentary protoliths of the MMC. The ages for the eastern Australian Craton given by

Pell et al. (1997) show a poor correlation with the MMC zircon ages, making it unlikely that the MMC was originally positioned near the eastern margin of Australia, even although this area is the only source of Devonian and Carboniferous zircons in continental Australia, forcing the interpretation that during the Paleozoic the MMC and Bird's Head did not form part of the eastern Australian Craton, but instead developed in the northern Australian craton-New Guinea region. From the above we may conclude that it is not necessary to explain the MMC and Bird's Head as exotic terranes derived from a distant location, but that they could have developed during the Paleozoic as an active continental margin positioned to the northwest of the Australian Craton.

The greenschists facies rocks that form a discontinuous selvage around the MMC have a different origin. The schists found along the western margin appear to have been derived from sedimentary rocks of the adjacent Paleogene Tinombo Formation, based on the following evidence: 1. they have a sedimentary precursor; 2. their geochemistry is similar to that of Tinombo greywackes; and 3. the (recrystallized) limestone intercalations that occur in the greenschists are similar to those found in the Tinombo Formation (Brouwer, 1934). The southeastern margin greenschists are derived from the Papayato Volcanics, as evidenced by: 1. their dominantly igneous origin; 2. gradational relationships observed between the two units; and 3. similar geochemical and isotopic characteristics. The origin of the amphibolites and related rocks is less certain, although our isotope and geochemistry data suggest, that in part they may be related to the Papayato Volcanics. The close spatial relationship between greenschists and amphibolites observed in the Marissa area supports this interpretation. The greenschist selvage appears to be similar to the metamorphic carapaces observed in several metamorphic core complexes in the SW USA. These carapaces consist of upper plate (cover rock) strata that are commonly metamorphosed to upper greenschist and amphibolite grade. The rocks form a relatively thin, tabular sheet and appear to be welded or plated to underlying crystalline rocks (Davis, 1980).

The rocks of the MMC show mineralogical changes that are typical of medium P/T regional metamorphism, progressing from greenschist through epidote-amphibolite to amphibolite facies in the meta-basites, and chlorite to biotite zones in the meta-sediments/meta-granitoids. The equivalents of the amphibolite facies, i.e., the staurolite and kyanite zones, have not been identified in the latter rocks, with the possible exception of the kyanite (with cores of staurolite) -bearing gneiss reported by Koperberg (1929). This may be due to the prevalence of arenaceous over pelitic protoliths. Estimated PT conditions in meta-basite samples representing the greenschist, epidote-amphibolite and amphibolite facies, assuming that the mineral parageneses were formed under ambient conditions, are 300–350 °C at 3–5 kbar, 440–522 °C at 4.8–6.2 kbar, and 646–617 °C at 7.5–9.6 kbar, respectively. The geothermal gradi-

ent estimated from the P–T curve (Fig. 6) is 20 °C/km, which is at the lower end of geothermal gradient ranges of medium P/T regional metamorphic belts (20–30 °C/km). The maximum pressures of 9–10 kbar estimated for the amphibolite rock indicates burial up to 27–30 km. The decompressional and cooling P–T path reflects subsequent uplift of the MMC, which may have been accompanied by granitoid intrusions, as suggested by relatively high temperatures shown by some of the amphibolite samples.

The tectonic history of the MMC is still contentious, due to the lack of sufficient constraining data. Several authors have suggested that the complex is an exposure of autochthonous Pre-Tertiary metamorphic basement rocks in the Cenozoic volcanic arc (Koperberg, 1929; Ratman, 1976). This hypothesis is not consistent with the following observations: 1. the contact between the MMC and Paleogene formations is tectonic rather than depositional; 2. no metamorphic detritus has been observed in the adjacent Tinombo Formation; and 3. radiometric isotope ages obtained in this study indicate that metamorphism and subsequent exhumation of the complex took place during the Miocene and not during the Pre-Tertiary. Rangin et al. (1990) proposed that the MMC originated in south China from which it rifted away during the Tertiary, with a volcanic arc developing on it while it travelled southwards. This model is not supported by available paleomagnetic data that indicate that the North Arm of Sulawesi did not undergo any significant latitudinal shift during the Tertiary (Sasajima et al., 1980; Surmont et al., 1994; Walpersdorf et al., 1998). Kavalieris et al. (1992) briefly discussed the possibility that the MMC is a metamorphic core complex, the formation of which they linked to the collision of the Banggai-Sula microcontinent with eastern Sulawesi, which they assumed took place during the Middle Miocene. They suggested that the greenschist rocks represent a ductile shear zone (detachment fault).

We consider here two possible scenarios. In the first scenario a continental fragment rifted from the New Guinea-Australia margin during the Mesozoic and accreted to the Sundaland margin in the Cretaceous. It was subsequently exhumed as a metamorphic core complex in an extensional setting during the early mid-Miocene. The MMC has several features in common with a number of metamorphic core complexes in other parts of the world, including the North American Cordillera (e.g. Coney, 1980; Davis, 1980; Spear, 1995) and the Aegean region (e.g. Lister et al., 1984; Lister, 1990): 1. an elongated dome-like structure forming a local topographic high; 2. relatively small dimensions; 3. Barrovian-type metamorphism; 4. limited petrographic evidence for ductile deformation; 5. heterogeneous K/Ar and <sup>40</sup>Ar/<sup>39</sup>Ar ages (cf. Lister and Baldwin, 1993); 6. sharp contacts between slightly to unmetamorphosed cover rocks and relatively high grade metamorphics; 7. presence of a greenschist carapace; 8. thrusting and isoclinal folding in upper plate rocks; and 9. lack of thick syn-to late-orogenic sedimentary deposits, suggesting denudation was predominantly tectonic rather than erosional. Although the tectonic

setting of the Sulawesi region during the Neogene was overall contractional, there was also significant localized extension at various times (Hall, 2002), which led, for example, to formation of Early Miocene grabens in SE Sulawesi (Suruno, 1994), Middle Miocene block faulting accompanied by high-K magmatism in Western Sulawesi (MacPherson and Hall, 1999; van Leeuwen and Muhandjo, 2005), and during the Late Miocene formation of the Banda Sea (Hall and Wilson, 2000) and extension in the Lariang-Karama region in NW Sulawesi (Calvert and Hall, 2003). In the Malino region, a similar localized extensional event may have taken place, which facilitated the development of the MMC as a core complex.

The second scenario envisages that a fragment of continental crust was sliced from the Bird's Head during the late Oligocene and moved rapidly westwards to collide with the Sundaland margin in the earliest Miocene followed by rapid exhumation. The crust was subducted in a regime characterized by low pressure between the overriding and subduction plates as described by Chemenda et al. (1996). According to these authors such a regime is efficient in rapidly exhuming subducted material without requiring erosion. This model is primarily based on the occurrence of an Early Miocene tectonic event in the western and central parts of the North Arm, which caused strong deformation in and uplift of the Paleogene formations (see above). There is wide agreement that Sulawesi is a region of collision between continental, ophiolitic and island arc fragments, that began during the Late Oligocene–Early Miocene (e.g., Audley-Charles et al., 1977; Katili, 1978; Hamilton, 1979; Milsom, 2000; Hall and Wilson, 2000). However, opinions differ as to whether the earliest collision events were restricted to eastern Sulawesi (Parkinson, 1991, 1998; Hall, 2002) or involved subduction of a microcontinent beneath Western Sulawesi (and the North Arm) as proposed by Bergman et al. (1996), Villeneuve et al. (1998, 2002) and Kadarusman et al. (2004, 2005). The lack of collisional features in the Oligocene–Miocene stratigraphy of Western Sulawesi (Calvert, 2000; Hall and Wilson, 2000; Hall, 2002), except for the most northern part, suggests that only a relatively small fragment was involved in the collision, the impact of which was restricted to the North Arm.

## 11. Conclusions

The results of our study of the Malino Metamorphic Complex provide the first direct evidence for the existence in Northwest Sulawesi of a continental fragment that was derived from the New Guinea–Australian margin of Gondwanaland. Furthermore we propose that this tie to the Australian continent (isotope signatures and zircon ages) is one step removed, in that its Australian signature derives from deposition during the Paleozoic of a sediment pile at the continental margin in the New Guinea region, with the sediments probably being supplied from Western Australia. Devonian (~417 Ma) and Carboniferous (~321 Ma) granitoids intruded this sediment pile and were

probably derived from it, forming a continental margin magmatic arc. This piece of New Guinea crust was then transported to and incorporated in Sulawesi.

We can be specific about the part of New Guinea from which the MMC was probably derived. The MMC shares several characteristics with those from the basement rocks of the Kemum Terrane in the Bird's Head, including age range, nature of the protholiths (quartz-rich proximal turbidites), a Precambrian age of (some of the) provenances sources, and age of magmatic activity.

Two possible scenarios are proposed in this paper for the tectonic history of the MMC, i.e. 1. accretion of the continental fragment to the Sundaland margin in the Cretaceous and exhumation as a metamorphic core complex in the Miocene; and 2. the fragment was sliced off the Bird's Head, subducted beneath the North Arm and rapidly exhumed during the Late Oligocene–Middle Miocene. These scenarios are based on our current knowledge of the MMC and the tectonic evolution of the Sulawesi region, which in either case is still limited. More detailed studies of the MMC, including its structural features, and a better understanding of the tectonic history of the region are required to test the two scenarios. This should result in the formulation of a more detailed and reliable model.

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## Appendix A. Analytical methods

A limited number MMC rock samples were analyzed by Actlabs (Canada) for major oxides and a wide range of trace elements by fusion-inductively coupled plasma mass spectrometry methods, respectively, using a lithium metaborate fusion rather than acid digestion to place the sample into solution. This fusion technique ensures that all major elements can be determined with an accuracy that is better than 0.01% absolute. Accuracy for the trace elements varies between <20% and <5% for different elements.

Sr, Nd, and Pb isotope ratios of seven samples were analysed at the Department of Geology and Geophysics at the University of Adelaide on a Finnigan MAT 262 Thermal Ionisation Mass Spectrometer in static mode, following procedures described by Elburg et al. (2003b). The long-term average for the in house Nd standard (J&M specpure Nd<sub>2</sub>O<sub>3</sub>) is 0.511603 ± 9 (1 sigma of total population,

$n=105$ ). The LaJolla standard gave  $0.511828 \pm 11$  ( $n=9$ ) and BCR-1 was  $0.512593 \pm 16$  ( $n=12$ ). Replicate analyses of the in-house standard TasBas indicated that reproducibility of the  $^{147}\text{Sm}/^{144}\text{Nd}$  ratio was better than 0.8%. The average for the NBS987 Sr standard is  $0.710258 \pm 18$  ( $n=56$ ). Typical Sr blanks are better than 1.5 ng, which is negligible compared to a typical sample size of 10–100 µg of Sr.

Ten representative samples from the MMC were selected for chemical analyses of mineral compositions. Analyses were performed at the Department of Earth and Planetary Sciences of Tokyo Institute of Technology using a JEOL 8800 EPMA (electronprobe microanalyzer) with an accelerating voltage of 15 kV and beam current of 12 nA, and a beam spot of 1–5 µm in diameter. X-ray intensities were converted to deuterated concentrates using a ZAF matrix correction scheme.

K/Ar age determinations were made on muscovite from two muscovite schist samples. Dating of sample NWS-38 was performed at the Institute of Geological and Nuclear Sciences Ltd on a modified Ms-1a mass spectrometer, following in-house methods developed by C. Adams. Errors of age measurement are 2SD and combine individual errors of potassium (+0.5%),  $^{38}\text{Ar}$  spike calibration (+0.3%), Ar isotope measurement (<0.1%) and instrument mass discrimination variation (for  $^{40}\text{Ar}/^{36}\text{Ar} < 0.3\%$ ). The other sample, NWS-61, was dated at AMDEL analytical laboratories on a modified MS-10 mass spectrometer, using techniques described by Webb et al. (1986). Duplicate K analyses agree within 0.5% and duplicate Ar analyses within 1%. The errors given are for the analytical uncertainty at 2SD.

A total of 29 single crystals of muscovite, biotite and hornblende from six samples were analyzed by the laser incremental heating  $^{40}\text{Ar}/^{39}\text{Ar}$  dating method. A minimum of two mineral grains were analyzed for each sample; most samples, however, had 3–6 grains analyzed. In most cases reproducible and well-constrained results were obtained. The hand specimens were crushed and the crushed fraction was washed with distilled water and absolute ethanol. Ten to twenty pure grains from each sample were hand picked and loaded into irradiation disks along with Fish Canyon (28.02 Ma) and GA1550 biotite (98.8 Ma) standards. These disks were wrapped in Al-foil, vacuum sealed in silica glass tubes and irradiated for 14 h at the B-1 CLICIT facility at the Radiation Center, Oregon State University, USA. After a two-month cooling period, each sample was step-heated under a continuous Ar-iron laser with defocused beam. The fraction of gas released was cleaned through a cryocooled cold-trap ( $T=130^\circ\text{C}$ ) and two C-50 SAESTM Zr–V–Fe getters and analyzed for Ar isotopes in a MAP-215-50 mass spectrometer. Automation and analytical procedures followed are described in Deino and Potts (1990), Deino et al. (1990) and Vasconcelos (1999). Data corrected for mass discrimination, nucleogenic interferences, and atmospheric contamination were used to calculate apparent ages for each degassing step. All analyses were carried out at the UQ-AGES (University of Queensland Argon Geochronol-

ogy in Earth Sciences) laboratory. The grains analyzed in this study were small single crystals (0.3–2 mm) extracted from each sample. The grains obtained were large enough and suitable for analysis by the incremental heating  $^{40}\text{Ar}/^{39}\text{Ar}$  method. However, some hornblende crystals showed some signs of recrystallization or alteration.

Four samples (NWS 55; 941, 942, and 943) were selected for U–Th–Pb zircon dating. NWS-941, a diorite, did not yield zircon. The three zircon-bearing samples were run in different 8 h analytical sessions at the Research School of Earth Sciences, ANU. A Lambda Physik LPX 120I ArF excimer laser with a 193 nm wavelength and 29 µm beam diameter operating at 5 Hz was used to ablate holes in zircons, standard zircons, and NIST 610 glass mounted in epoxy and polished to expose zircon cores. The mineral mount was put into a sample cell connected to an Agilent 7500S quadrupole ICP-MS. The method is that reported in Bryan et al. (2004). Zircons ages are interpreted as robust if the analysis is concordant (see below) and if the measured uncertainty in the isotope ratio used for age determination is less than twice that expected from counting statistics (see Bryan et al., 2004).

Cathodoluminescence images were acquired on a Hitachi S-2250N SEM using a 15KV filament current and carbon-coated mineral mounts.

A zircon can be discordant for three reasons: 1. the presence of common Pb; 2. loss of Pb at some substantial time after zircon crystallization; or 3. the result of simultaneous ablation of zircon domains of different age (age mixing). We employed a noise filter on the main dating isotope ratios ( $^{207}\text{Pb}/^{206}\text{Pb}$  or  $^{206}\text{Pb}/^{238}\text{U}$ ) that screens from consideration such mixed zircon ages unless the age-domain contact is exactly parallel to the laser beam, a statistically uncommon circumstance (Bryan et al. accepted). Considering the other cases, as we cannot measure common Pb directly because counts on systemic Hg including  $^{204}\text{Hg}$  are many times that of  $^{204}\text{Pb}$ , we must assume that a grain is discordant for reason 1. or reason 2. We cannot correct for both common Pb and Pb-loss. Our strategy is as follows. First we determine which analyses (employing no common Pb correction) are concordant at the 5% level in samples where discordant grains are numerous, the case for all three samples presented here. We define a grain as concordant if  $(^{207}\text{Pb}/^{206}\text{Pb} \text{ age} \pm 1)/(^{206}\text{Pb}/^{238}\text{U} \text{ age} \pm 1) = 0.95$  to 1.05 for grains older than 900 Ma and we quote the former as the age, and  $(^{207}\text{Pb}/^{235}\text{U} \text{ age} \pm 1)/(^{206}\text{Pb}/^{238}\text{U} \text{ age} \pm 1) = 0.95$  to 1.05 for younger grains and we quote the latter as the age. The concordant results are shown in age histograms as solid bars, whereas discordant ones are open bars (Fig. 11). Because linear arrays, called ‘discordia’, are prominent on concordia plots, and since these discordia have low slopes on Terra–Wasserburg plots (ensuring that Pb-loss and not common Pb dominates), then these discordant analyses and appropriate concordant ones are modelled for upper and lower age intercepts using Isoplot (Ludwig, 2003). The differences among the ages we quote, and  $^{208}\text{Pb}$ -based common Pb-corrected and  $^{207}\text{Pb}$ -based common Pb-corrected

can be inspected in the complete data table, which has been lodged with the journal database.

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