Geological evolution of the Antongil Craton, NE Madagascar


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1. Introduction

The crystalline basement on the island of Madagascar is made up of a collage of accreted terranes (cratons) ranging in age from Archaean to Neoproterozoic, which lie within the core of the Palaeozoic supercontinent of Gondwana. Most of the terranes were juxtaposed during the late-Neoproterozoic to early Palaeozoic collision of East- and West-Gondwana during the “Pan-African” or “East African” orogeny (EAO; Stern, 1994). The oldest rocks in Madagascar are found in three cratonic blocks known as the Antananarivo, Antongil and Masora Cratons (Fig. 1). The Antananarivo Craton, located in central and northwest Madagascar, is composed of Neoarchean rocks considered to lie within the core of the Palaeozoic supercontinent of Gondwana. The Antongil Craton, along with the Masora and Antananarivo cratons, make up the fundamental Archaean building blocks of the island of Madagascar. They were juxtaposed during the late-Neoproterozoic to early Palaeozoic assembly of Gondwana. In this paper we give a synthesis of the geology of the Antongil Craton and present previously published and new geochemical and U–Pb zircon analyses to provide an event history for its evolution.

The oldest rocks in the Antongil Craton form a nucleus of tonalitic gneiss, characteristic of Palaeo-Mesoarchaean cratons globally, including phases dated between 3320 ± 14 Ma to 3231 ± 6 Ma and 3187 ± 2 Ma to 3154 ± 5 Ma. A series of mafic dykes was intruded into the Mesoarchaean tonalites and a sedimentary succession was deposited on the craton prior to pervasive deformation and migmatisation of the region. The age of deposition of the metasediments has been constrained from a volcanic horizon to around 3178 ± 2 Ma and subject to migmatisation at around 2597 ± 49 Ma. A subsequent magmatic episode generated voluminous, weakly foliated granitic rocks, that also included additions from both reworked older crustal material and younger source components. An earlier granodiorite-dominated assemblage, dated between 2570 ± 18 Ma and 2542 ± 5 Ma, is largely exposed in xenoliths and more continuously in the northern part of the craton, while a later monzogranite-dominated phase, dated between 2531 ± 13 Ma and 2513 ± 0.4 Ma is more widely developed. Together these record the stabilisation of the craton, attested to by the intrusion of a younger dyke swarm, the age of which is constrained by a sample of metagabbro dated at 2147 ± 6 Ma, providing the first evidence for Palaeoproterozoic rocks from the Antongil Craton.

The youngest events recorded in the isotopic record of the Antongil Craton are reflected in metamorphism, neocrystallisation and Pb-loss at 792 ± 130 Ma to 763 ± 13 Ma and 553 ± 68 Ma. These events are interpreted as being the only manifestation of the Pan-African orogeny seen in the craton, which led to the assembly of the tectonic blocks that comprise the island.

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Fig. 1. Simplified geological map (A) of the Antongil Craton, showing the main lithodemic units and U–Pb dates. Units of the Ankavanana Suite are too small to be differentiated at this scale. Coordinates are in lat/long and laborde. Geochronological ages in italic represent results of previously published U–Pb dating of Tucker et al. (1999) and Paquette et al. (2003). Inset map (B) shows the major Precambrian crustal terranes of Madagascar, modified after Collins, 2006. APT = Andaparaty Thrust; AB–MO = Anaboriana Belt–Manampotsy Belt; AN = Antongil Craton; BE = Bemarivo Belt; MA = Masora Craton; NT = Antananarivo Craton; IT = southern mobile belts including the Itremo Group; VO = Vohibory Unit.
Table 1
Summary lithological and petrographic features of each lithodemic unit from the Antongil Craton. Mineral abbreviations after Sivula and Schmid (2007).

<table>
<thead>
<tr>
<th>Unit</th>
<th>Compositional range (major bold)</th>
<th>Mineralogy</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Supracrustal rocks</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mananara Group</td>
<td>Mafic schist</td>
<td>Act (30%) + Chl (20%) + P + (15%) + Ttn (10%) + Qtz (5%) + Op (5%) + Ep (2%) + (Px ± carbonate) (13%)</td>
<td>Occurs as xenolith pods and lenses. Preserves a strong foliation defined by oriented aggregates of brown, poikiloblastic amphibolite. Widely exposed in the south and west of the craton. Psammites are interlayered with pelitic and quartzitic units as well as amphibolite and ultramafic lenses.</td>
</tr>
<tr>
<td><strong>Ambodiriana Formation</strong></td>
<td>Psammitic schist, pelitic schist, quartzite, magnetite quartzite, amphibolite, ultramafic rock</td>
<td>Pl (5–35%) + Qtz (20–50%) + Am (5–5%) + Bt (15–20%) + Grt (10–5%) + Op (5–10%) + Px ± Ttn (5%) + Ms ± Kfs ± Chl ± (Ky, Stil and Crd in pelitic horizons).</td>
<td></td>
</tr>
<tr>
<td><strong>Granitic rocks</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Nosy Boraha Suite</td>
<td>Tonalite gneiss, granodiorite and granite gneiss, amphibolite</td>
<td>Pl + Qtz + Bt = Am + Cpx ± Kfs</td>
<td>Intercalated tonalite to granite orthogneiss. Preserves strong planar fabrics. Locally occurs as enclaves within the Masoala Suite.</td>
</tr>
<tr>
<td>Masoala Suite: Nosy Mangabe Granite</td>
<td>Monzogranite to granodiorite</td>
<td>Pl (30–40%) + Kfs (20%) + Qtz (15–20%) + Bt (15–20%) + Am (5–15%) + Op (5–10%)</td>
<td>Most extensive granitic facies of the suite. Heterogeneous and locally migmatitic. Contains diffuse tonalite to diorite facies as well as enclaves of ortho- and paragneiss. Homogeneous, pink, unfoliated medium to coarse granite. Exposed in northeast of the craton.</td>
</tr>
<tr>
<td>Masoala Suite: Onive Granite</td>
<td>Monzogranite to granodiorite</td>
<td>Kfs (15–30%) + Pl (5–10%) + Qtz (20–25%) + Bt (5–8%) + Am (1–5%) + Op (2–5%) + Px ± Chl ± Ap ± Zrn ± Aln Kfs + Pl + Qtz + Bt</td>
<td>Widespread migmatitic granite with nebulitic and schlieric textures. Pink, locally porphyritic granite.</td>
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<tr>
<td>Masoala Suite: Ampanabe Granite</td>
<td>Syenogranite</td>
<td>Pl (25–30%) + Qtz (20–35%) + Kfs (5%) + Bt (5–10%) + Qtz (10–20%) + Am (15–25%) + Op (1–5%) + Px = Zrn ± Aln ± Ms</td>
<td>Widespread migmatitic granite with nebulous and schlieric textures. Pink, locally porphyritic granite.</td>
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<td>Masoala Suite: Sarahandraniso Granodiorite</td>
<td>Granodiorite to tonalite</td>
<td>Qtz (25–30%) + Pl (20–35%) + Kfs (5%) + Bt (5–10%) + Ep (&lt;10%) + Chl (&lt;10%) + Am (1–5%) + Op (1–3%) + Ms ± Ttn (5%) ± Pt</td>
<td>White-weathering and homogeneous, widespread in the north of the craton. Locally interlayered with units of diorite.</td>
</tr>
<tr>
<td>Masoala Suite: Andranotfosty Foliated Granite</td>
<td>Monzogranite to granodiorite</td>
<td>Kfs ± Pl + Qtz + Bt + Am</td>
<td>Widespread migmatitic granite with nebulous and schlieric textures. Pink, locally porphyritic granite.</td>
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<td>Masoala Suite: Betavona Granite</td>
<td>Syenogranite to monzogranite</td>
<td>Kfs (45–50%) + Qtz (35–40%) + Pl (5–10%) + Bt (5–7%) + Op (1–3%)</td>
<td>Medium- to coarse-grained with relic igneous px and opx to sub-ophitic textures commonly preserved. Widespread alteration to chlorite and epidote.</td>
</tr>
<tr>
<td><strong>Metabasites</strong></td>
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<tr>
<td>Antanovana Suite</td>
<td>Diorite/Gabbro</td>
<td>Pl (25–30%) + Am (2–30%) + Chl/Ep (20–30%) + Op (5–10%) + Qtz (5%) ± Px ± Ttn (5%) ± carbonate</td>
<td>Widespread migmatitic granite with nebulous and schlieric textures. Pink, locally porphyritic granite.</td>
</tr>
</tbody>
</table>
matism, led Cox et al. (1998) to infer that the two cratons were distinct crustal blocks separated by a suture zone. This interpretation considers the Antongil Craton to be a displaced fragment of the Dhawar Craton of India and to have been juxtaposed against the Antananarivo Craton during the Pan-African EAO. The ‘Betsimisaraka Suture Zone’ proposed by Collins and Windley (2002) and Kröner et al. (2000) was interpreted as the suture zone between the Antongil and Antananarivo Cratons. However, an alternative interpretation was proposed by Tucker et al. (1999) who noted that depleted mantle ages of around 3.2 Ga for both the Antongil and the Antananarivo Cratons did not rule out derivation of both terranes from similar primary sources.

3. Geological description

3.1. Contacts of the Antongil Craton with adjacent crustal blocks

Along most of its northern boundary, the Antongil Craton is structurally overlain by medium- to high-grade metasedimentary and granitoid rocks of the Bemarivo Belt (Fig. 1). This preserves dissected remnants of accreted Neoproterozoic arc and Palaeoproterozoic marginal basin successions that were thrust over the Antongil Craton during the Cambrian EAO (Thomas et al., 2009). The basal contact of the thrust forms an imbricate fan, referred to as the Andaparaty Thrust, that generally dips northward at a low angle, and is locally marked by a thin mylonitic unit preserved in the metasedimentary rocks of the hanging wall. Along much of its trace, the thrust zone is a complex imbricate belt several hundreds of metres thick, with structural interleaving between units of the Antongil Craton and Bemarivo Belt (Thomas et al., 2009).

South of the contact with the Bemarivo Belt, along its northwestern margin, the Antongil Craton is in tectonic contact with the Antananarivo Craton (Fig. 1). The contact is poorly exposed, and trends across a high, deeply lateritised plateau. However, in some areas comprises a ca. 2 km wide zone of moderately NW-dipping mylonitic gneisses with asymmetric structures and steeply plunging mineral lineations giving a top-to-the-south sense of tectonic translation.

Further south, the Antongil Craton is in contact with the Manampotsy Belt (Fig. 1), a newly mapped belt that is contiguous with the Antananarivo Craton (Fig. 1). The contact is poorly exposed, and crosses a high, deeply lateritised plateau. However, in some areas comprises a ca. 2 km wide zone of moderately NW-dipping mylonitic gneisses with asymmetric structures and steeply plunging mineral lineations giving a top-to-the-south sense of tectonic translation.

3.2. Geological units of the Antongil Craton

3.2.1. Supracrustal rocks

3.2.1.1. Mananara Group. Volumetrically minor, disrupted metatexitic, mafic schist and paragneiss bodies, termed the Mananara Group (Fig. 1), occur as xenolithic pods, lenses and schlieren in the Masaola Suite granitoids and as a more extensive unit in the, poorly exposed, central part of the craton. In the literature (e.g. Besairie, 1964) they are included in the “Groupe d’Antongil”, but this name has been changed to avoid confusion. Field relationships show that the formation comprises supracrustal rocks that likely represent isolated remnants of the country rocks to the Neoarchaean Masaola Suite granitoids. In the southern part of the craton, a diverse assemblage of schist and gneiss, The Ambodiriana Formation, represents a subdivision of the group.

The Mananara Group includes fine- to medium-grained, yellowish-green quartz-chlorite schists and more mafic green chlorite-actinolite schists, interpreted as former mafic volcanic rocks. Exposures of this formation are largely covered by dense vegetation, hence the field relations of these rocks were rarely observed, but they appear to have been intruded by the Neoarchaean granitoids of the Masaola Suite and, at their most extensive, broadly define a lensoid unit, up to around 3 km in length.

3.2.1.2. Ambodiriana Formation. The southern and western parts of the craton are underlain by an extensive, but poorly exposed, supracrustal assemblage, termed the Ambodiriana Formation (Fig. 1) that may have originally unconformably overlain the craton, following exhumation of the granitoids. The formation is broadly equivalent to the “Série d’Ambodiriana” of Giraudon (1958). Previous studies have divided the formation into three lithofacies: the Vavetenina (ky-bearing), Ambodiriana (ky+sill-bearing) and Antenina (ky+sill+grt-bearing) facies (Hottin, 1969, 1976), although the present survey found these unsustainable as mappable units.

The formation is dominated by a variety of proportionally variable and locally interleaved migmattic, staurolite-, sillimanite- and kyanite-bearing pelitic and psammitic schists and gneiss (Fig. 2a–c). Subordinate units of foliated amphibolite occur throughout the formation and fuchsite- and magnetite-bearing quartzites locally occur in association with units of ultramafic rocks.

The Ambodiriana Formation, a protracted structural history with up to threelfold generations and multiple transposed tectonic fabrics preserved in any one locality. Kyanite porphyroblasts are aligned with tectonic fabrics that are overprinted by at least two subsequent fold episodes suggesting that medium- to high-grade metamorphism may have occurred in these rocks relatively early in their tectonothermal development. Generally, the overall structural style is defined by the latest generation and comprises a series of tight, upright pericline folds with an approximately 6–8 km wavelength. Earlier folds are generally preserved as tight to isoclinal intrafold structures. The relationship between the
Ambodiriana Formation and Mananara Group cannot be demonstrated as, during the present study, the latter were only observed as units enclosed entirely within younger granitoids.

3.2.2. Migmatitic orthogneiss

3.2.2.1. Nosy Boraha Suite. In the southern part of the craton, the Ambodiriana Formation is locally intercalated with an extensive unit of Mesozoic age rocks (>3 Ga) TTG orthogneiss, known as the Nosy Boraha Suite, first identified by Tucker et al. (1999). The suite crops out on the island of Nosy Boraha (Ile Sainte Marie) and on the adjacent mainland area in the south of the craton (Fig. 1). On Nosy Boraha, Tucker et al. (1999) described the lithology as comprising ‘migmatized tonalitic and granodioritic gneiss’. In the southern part of the craton on the mainland (Fig. 1), the Nosy Boraha Suite forms a more extensive exposure. Here, the main lithology is a coarse- to medium-grained granitic, granodioritic and tonalitic gneiss locally intercalated with sheets of garnet-bearing amphibolite ranging from centimetres to a few metres in thickness (Fig. 2d) and interpreted to be an intrusive facies that was deformed along with the host granitoids to form the observed gneiss.

3.2.3. Granitic rocks

3.2.3.1. Masoala Suite. The Masoala Suite, which is the dominant Neoarchean igneous complex of the Antongil Craton (Fig. 1), is included within a single igneous unit following Besairie (1970). The suite comprises a variety of local, texturally distinct phases whose inter-relationships are complex at all scales from the regional to individual outcrops. However, it can be subdivided into two assemblages. A northern sub-suite comprises a granodiorite-dominated, strongly heterogeneous facies largely exposed adjacent to the northern contact with the Bemarivo Belt, but also includes scattered enclaves exposed within a more extensive, uniform southern
sub-suite of dominantly monzogranitic composition. The lithological descriptions of the different phases are based on field characteristics and petrography; the limited number of samples analysed means that geochemical distinction of the different phases is not possible.

The Masoala Suite is characterised by a high degree of structural isotropy; a feature that sets the Antongil Craton apart from the other cratonic blocks of Madagascar. Although planar fabrics are widely preserved within the granitic rocks of the Masoala Suite, they are generally of variable intensity, orientation and inclination, ranging from locally developed gneissic layering, to homogeneous magmatic fabrics and alignments of cognate enclaves and xenoliths and vein networks with a preferred orientation. Proto-mylonitic sub-solidus fabrics are locally preserved in narrow ductile shear zones that represent weak, moderate grade (greenschist facies) deformation of the granitic rocks that post-dates magmatic assembly. Although no clear regional structural grain is preserved, open folding of the mapped units around a north-trending axis in the north of the craton and a west-trending axis toward the south, exert regional control to the distribution of the units (Fig. 1). On the basis of the overall mapped architecture, the west-trending folds appear to be equivalent to the latest phase of folding in the Ambodiriana Formation.

**Southern sub-suite of the Masoala Suite:** The greatest part of this assemblage comprises monzogranitic facies termed the Nosy Mangabe Granite. This unit underlies >50% of the craton and forms most of the easily accessible outcrops around Antongil Bay (Fig. 1).

The Nosy Mangabe Granite is typically grey to pinkish, coarse-grained, inequigranular, heterogeneous and composed of variable proportions of quartz, K-feldspar, plagioclase, biotite and hornblende, with macroscopic titanite and sulphide occasionally visible. This unit ranges in composition from granodiorite through to syenogranite and quartz-monzodiorite to quartz-syenite, but is dominated by monzogranite. In fresh coastal exposures, the rocks preserve fabrics including both magmatic and tectonic foliations, although these tend to be irregular in intensity and orientation at outcrop scale.

Textural variants range from medium to coarsely crystalline aphyric facies that make up the vast majority of the unit, to K-feldspar porphyritic or megacrystic facies (Fig. 2e). These facies occur as irregular sheet-like bodies with diffuse to sharp boundaries. Narrow biotite selvedges are observed transecting granitic facies that are indistinguishable in the field, but are interpreted to represent separate, possibly comagmatic, intrusive phases with weakly chilled margins. Irregular migmatitic facies are widely developed and commonly have an agmatic (blocky) or nebulitic (diffuse) appearance. In the latter, variable amounts and multiple generations of coarse leucosome are developed as irregular patches, blebs, veins, anastomosing vein arrays, small shear-zone infills, stockworks and diffuse masses (Fig. 2f). Vein arrays locally preserve a crude parallelism. K-feldspar-phric varieties are locally contiguous with coarse-grained aphyric granite forming narrow veins and apophyses.

Most outcrops of Nosy Mangabe Granite contain abundant mafic enclaves. These occur either in isolation, or more commonly as clusters or elongate trains. Individual enclaves quite commonly occur as large rounded to sub-angular and irregular blocks ranging up to 2 m in size, to small mafic blebs and wispy, attenuated schlieren in more foliated parts (Fig. 2g). The margins of the enclaves are commonly diffuse and they frequently preserve partly digested K-feldspar phenocrysts or quartz ocelli textures suggesting that they are themselves the products of magma mixing and may, in some cases, represent relic syn-plutonic dykes.

Diffuse to sharply bounded bodies of tonalite and diorite occur as enclaves and isolated masses within the granite, ranging from several metres in size, through to ovoid or folded relics up to around 10 km in length. The largest body covers about 60 km², and was originally mapped as the “Plagioclaseolite quartzite” unit of Besairie and Bertucat (1974). The tonalite is either massive and homogeneous or variably migmatised (Fig. 2h). It is distinguishable from the Nosy Boraha tonalitic orthogneisses by a more nebulous, less penetrative fabric in which hornblende or biotite-bearing leucosomes form irregular veins, blebs and networks.

In a few localities the Nosy Mangabe Granite contains mafic to ultramafic enclaves composed of amphibole, chlorite and epidote. In other areas retrogressed xenoliths of orthogneiss are preserved with prominent pale green plagioclase-rich rims.

The heterogeneous nature of the Nosy Mangabe Granite, including both gradational and sharp intrusive boundaries, as well as partially or thoroughly mixed syn-plutonic dykes, suggest that it results from the injection of multiple phases of monzogranitic and more mafic magmas where both widespread mixing and mingling played a significant part in the assembly of the unit.

In the northeast part of the craton, the dominant monzogranitic facies is termed the Onive Granite (Fig. 1). In contrast to the Nosy Mangabe Granite, this unit comprises equigranular, medium- to coarse-grained, homogeneous, massive, pinkish, biotite granite, largely devoid of enclaves, later veins or penetrative fabrics.

A subordinate facies of the southern sub-suite, the Anpanobe Granite forms a discrete ovoid pluton hosted by the Nosy Mangabe Granite (Fig. 1). This body comprises coarse-grained, locally K-feldspar-phric biotite granite. Although contacts with the adjacent Nosy Mangabe Granite were not observed, the periphery of the pluton is distinguished by a distinctive aphyric marginal facies, in which pink K-feldspar still forms the largest grains. The main facies of the pluton is typically megacrystic, with “matchbox” K-feldspar phenocrysts measuring up to 4 cm × 2 cm in size. These are commonly crudely aligned to form a shape fabric, interpreted as having formed by igneous flow. The main megacrystic facies is mostly homogeneous and contains only scattered rounded mafic enclaves.

**Northern sub-suite of the Masoala Suite:** Much of the northern sub-suite is made up of the Sarahandrano Granodiorite (Fig. 1). This unit forms a large body of light grey, white-weathering leucogranodiorite. The granodiorite varies from being equigranular and homogeneous to porphyritic (Fig. 2i) or migmatitic with a chaotic or nebulitic fabric locally developed. Locally the granodiorite contains alkali feldspar augen 1–2 cm across. Xenoliths are uncommon, though some thin biotite-rich schlieren locally define a foliation. Later intrusive veins are similarly uncommon. The granodiorite is typified by greenish plagioclase and small amounts of epidote, locally in lenses. In places, the granodiorite is intruded by coarse-grained biotite granite phases. A dioritic facies, locally interleaved with the Sarahandrano Granodiorite, tends to be heavily weathered and comprises fine- to medium-grained, granoblastic hornblende-plagioclase rock. A coarse-grained variety occurs as spotted biotite-hornblende-plagioclase ± epidote rocks.

The second most widespread unit in the northern part of the craton, the Andranofotsy Foliated Granite (Fig. 1), is characterised by a high proportion of migmatitic granite and is taken to include foliated anatectic granitoids crosscutting older gneiss units elsewhere in the craton. The granite was referred to as “Granite rose, migmatitique” by Bertucat (1965) and is characterised by variable amounts of coarse leucosome development as blebs, veins (often in anastomosing arrays), stockworks and diffuse masses. These rocks are not always migmatites sensu stricto as textural heterogeneity probably also reflects repeated phases of magma injection, or mingling and mixing of co-existing melts. The most intense stage of migmatisation is composed entirely of light grey, inequigranular, leucocratic, anatectic to nebulitic granite with mafic blebs, irregular masses and schlieren.
Immediately underlying the contact with the Bemarivo Belt, the Bezavona Granite (Fig. 1) crops out as extensive bodies of pink, equigranular to moderately porphyritic or glomeroporphyritic biotite granite (Fig. 2j), locally approaching syenite in composition. It is typically unfoliated and is rich in pink K-feldspar. Adjacent to the margins of the pluton a mylonitic foliation is strongly developed although contacts with the adjacent units of the craton were not observed.

### 3.2.4. Metabasic rocks

#### 3.2.4.1. Ankavanana Suite

Small bodies of unfoliated metabasite, termed the Ankavanana Suite, are preserved throughout the Antongil Craton. Although volumetrically small, their wide distribution suggests that they represent an important phase of magmatism intruded into the main granitoid phase. On the 1:200 000-scale geological map of Hottin et al. (1963), the suite is shown as individual intrusions of “dolerites filoniennes” and schematically as “nombreux filons de dolérites non relevés”, amplifying the wide geographical extent. The limited exposure of the suite has made it hard to determine the scale of these bodies (and negates their being differentiated in Fig. 1). However, where observed, they range from thin (1–3-m scale), sub-vertical dykes with a NE trend, to thick, over 100-m-wide pods (Fig. 2k). Larger bodies appear to be irregular and podiform, rather than straight-sided and dyke-like. At outcrop, the lithology is dark greyish-green metagabbro, with an average grain size of 1–5 mm. Typically, these metagabbro bodies are massive, unfoliated away from the contacts, with a faint, locally developed margin-parallel foliation. Intrusions are often transected by greenish chloritic veins. The coarser-grained metagabbros often contain diffuse, 10-cm-scale epidote pods. Larger intrusions tend to have finer grained contacts, representing original chilled margins.

### 4. Geochronology

Before this study, few published age determinations were available from the Antongil Craton. The only U–Pb zircon geochronological data are presented in two key publications.
Table 2

<table>
<thead>
<tr>
<th>Sample</th>
<th>XY</th>
<th>Lithology</th>
<th>Reference</th>
<th>Unit name</th>
<th>Age (Ma)</th>
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<td>U-Pb SHRIMP</td>
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<td>Tucker et al. (1999)</td>
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<td>Schofield et al. (2010)</td>
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<td>U-Pb SHRIMP</td>
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<td>Schofield et al. (2010)</td>
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<td>U-Pb SHRIMP</td>
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<td>Schofield et al. (2010)</td>
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<td>U-Pb SHRIMP</td>
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<td>800291 1224969</td>
<td>Granodiorite</td>
<td>Schofield et al. (2010)</td>
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Tucker et al. (1999) reported ages of 3187 ± 2 Ma and 2522 ± 4 Ma for units of tonalitic gneiss and biotite granite, respectively. Paquette et al. (2003) reported a three ages, 2532.1 ± 0.3 Ma, 2522.7 ± 0.3 Ma and 2512.7 ± 0.4 Ma, for various granitic rocks. In both cases, these ages were taken to represent the timing of the emplacement of the various protoliths, suggesting two major magmatic episodes in the development of the craton. More recently, a study by Tucker et al. (in press) has reported new ages of 4.1 Ma for paragneissic leucosome and 2502 ± 0.4 Ma, for various granitic rocks. As part of this study, a sample of the Antongil Craton was dated using the Sensitive High Resolution Ion Microprobe (SHRIMP) U–Pb technique on selected zircon grains at Curtin University of Technology, Perth, Western Australia. These include samples of the Masoala, Nosy Boraha and Ankavanana Suites, and the Ambodiriana Formation.

Zircon was separated from large, fresh rock samples using standard crushing, washing, heavy liquid separation (LST and MI liquids) and magnetic separation (Frantz Isodynamic Separator), followed by handpicking under a binocular microscope. The grains were mounted in epoxy, and polished mid-section to expose their center. Mounts were imaged using transmitted and reflected optical microscopy as well as cathodoluminescence (CL) on a scanning electron microscope. Examples of CL images of analysed zircon are shown in Fig. 3. Summary physical properties of zircons separated from the dated samples are shown in Table 3. Sample localities are also illustrated in Fig. 1. All coordinates are in the Laborde projection.

4.1. Supracrustal rocks – Ambodiriana Formation

None of the metasedimentary rocks were directly dated as part of this study. However, Sample DS30509 was collected from a narrow sheet of foliated granitoid concordant with the main layering/foliation preserved in host kyanite-bearing pelitic and migmatitic gneiss exposed on the foreshore at [730450 1005543] and interpreted as foliated leucosome derived from the host migmatitic paragneiss. Analysis of this sample is predicted to give an estimate of the age of metamorphism as well as upper age limit for the deposition of the metasediments.

43 analyses were conducted, including analyses on core and rim domains. The core domains range from irregular to subrounded, with variable preservation of original shape, and dominantly show medium response to CL. Many of the cores display remnant concentric zoning patterns consistent with an igneous protolith, but these patterns are often disturbed, possibly reflecting some radiogenic damage or solid-state recrystallisation, or a combination of both. $^{206}$Pb values for cores and rims range from 0 to 28.12%, with only three analyses containing more than 1% non-radiogenic Pb. U and Th contents are variable for both cores and rims, in the range 146–2590 ppm and 11–395 ppm overall respectively, but Th/U ratios are much lower for rims, between 0.01 and 0.20. The Th/U ratio for cores is in the range 0.03–1.18, but the lower values are invariably related to high-U and high-Pb analyses. Excluding those radiogenically damaged cores, the Th/U ratio ranges from about 0.3 to 1.2, consistent with a magmatic origin.

The data plot from near-concordant to extremely discordant, with a clearly defined cluster of data points near 3.2 Ga (Fig. 4), all obtained from zoned core domains, and interpreted to reflect the presence of a xenocrystic population of zircons. Many analyses plot significantly away from the concordia line and do not appear to define a meaningful regression line, interpreted to either reflect
Fig. 4. Wetherill (Concordia) plot from sample DS30509. All error crosses/ellipses are at 2σ confidence.

additional xenocrystic populations with slightly younger ages, or complex Pb-loss over time.

The most concordant data points allow the calculation of a Concordia age of 3176 ± 6 Ma (MSWD = 7.3). A second group of analyses mainly on rim domains but also containing some single-sector and concentrically zoned zircon, yields a set of poorly aligned discordant data, providing a poorly constrained upper intercept of 2597 ± 49 Ma (MSWD = 19). The age of the older population compares favourably with the age of 3178 ± 2 Ma yielded by a concordant metarhyolite within paragneisses north of Fenerive Est (Tucker et al., in press) and confirms the age of deposition of the sedimentary protolith. The regression of the younger population also concurs with results of Tucker et al. (in press) who interpreted 2550 ± 42 Ma analyses from pegmatitic leucosome sample 92 and 101% concordant. One analysis plots well away from the main linear array, and defines a 207Pb/206Pb age of 3352 ± 8 Ma, which we take the minimum crystallisation age of this xenocrystic grain. The remaining 14 data points define a linear set with an upper intercept at 3162 ± 18 Ma and a lower intercept at 955 ± 430 Ma (MSWD = 8.7; Fig. 5). The large MSWD value for this regression indicates significant scatter, possibly reflecting complex minor Pb-loss in the zircons. Four analyses define a Mesoarchaean concordia age of 3154 ± 5 Ma (MSWD = 0.46), which is taken as the best age estimate for crystallisation of the protolith to the gneiss.

A single concordant analysis on a large, medium-CL zircon that does not appear to match in appearance with other populations (Spot 1M; Fig. 3b) provides a Concordia age of 763 ± 13 Ma. In the light of the clear field relationships presented by Tucker et al. (in press), this age could be interpreted to represent neocrystallisation during a Neoproterozoic metamorphic event, but more data would be needed to fully confirm this.

4.2. Migmatitic orthogneiss – Nosy Boraha Suite

Sample MH73 is a medium- to coarse-grained, weakly foliated gneiss of granodioritic to tonalitic composition collected from a river section at [686698 0888561]. Fifteen analyses were conducted on fifteen separate zircon crystals, and reveal very low Pb contents. U and Th contents are in the ranges 129–837 and 35–655 ppm respectively, leading to Th/U ratios between 0.20 and 0.97, typical for magmatic zircon. The data plot between 92 and 101% concordant. One analysis plots well away from the main linear array, and defines a 207Pb/206Pb age of 3352 ± 8 Ma, which we take the minimum crystallisation age of this xenocrystic grain. The remaining 14 data points define a linear set with an upper intercept at 3162 ± 18 Ma and a lower intercept at 955 ± 430 Ma (MSWD = 8.7; Fig. 5). The large MSWD value for this regression indicates significant scatter, possibly reflecting complex minor Pb-loss in the zircons. Four analyses define a Mesoarchaean concordia age of 3154 ± 5 Ma (MSWD = 0.46), which is taken as the best age estimate for crystallisation of the protolith to the gneiss.

Sample DS30507 is a granitic schlieric migmatite gneiss collected from the foreshore at [729336 1002107]. Although contact relationships were not observed, the outcrop is hosted by relatively undeformed granitoids of the Masoala Suite, and is interpreted as a xenolith of older, previously deformed gneiss attributable to the Nosy Boraha Suite. In this area, the Masoala granites themselves are subordinate to metasediments of the Ambodiriana Formation (Fig. 1). 29 analyses were conducted on 25 zircon grains, including 4 core-rim pairs, 20 cores and 1 rim. 206Pb values are low, ranging from below background up to 0.27%. U and Th are in the range 120–964 ppm and 54–240 ppm respectively, with Th/U ratios between 0.07 and 0.63. The lowest Th/U ratios are recorded in the rim analyses and in two high-U cores (C26 and C27) which also show the lowest 207Pb/206Pb ratios (and thus lowest apparent ages).

Fig. 5. Plots of U–Pb data from magmatic zircon of the Nosy Boraha Suite: Tera Wasserburg plot from sample MH73; Wetherill (Concordia) plot from sample DS30507. All error crosses/ellipses are at 2σ confidence.
The data range from concordant to 53% discordant, and define a range of $^{207}\text{Pb}/^{206}\text{Pb}$ ages between 2371 Ma and 3217 Ma (Fig. 5). The oldest cluster of analyses of zircon cores defines a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of $3159 \pm 12$ Ma (MSWD = 30). Eight data points from this cluster give a concordia age of $3176 \pm 5$ Ma (MSWD = 2.7) which is interpreted as the best estimate for either the crystallisation of magmatic zircon in the protolith or the age of xenocrystic zircon. Analyses of two zircon rims provide $^{207}\text{Pb}/^{206}\text{Pb}$ ages of $2729 \pm 18$ Ma and $2749 \pm 12$ Ma, respectively (e.g. analysis 25 in Fig. 3d), interpreted to record a thermal pulse at that time. A younger cluster of core and rim analyses yields $^{207}\text{Pb}/^{206}\text{Pb}$ ratios corresponding to a weighted mean age of $2504 \pm 20$ Ma (MSWD = 10, see example of analysis 3 in Fig. 3e). The most concordant rim yields a concordia age of $2497 \pm 5$ Ma which is taken as the best estimate for a second thermal event and renewed zircon growth, including growth of relatively low Th/U zircon rims. One discordant zircon grain with a minimum $^{207}\text{Pb}/^{206}\text{Pb}$ age of $2371 \pm 14$ Ma does not lend itself to interpretation.

In summary, two new analyses of gneiss from the Nosy Boraha Suite yield Mesoarchaean crystallisation ages for granitic protoliths of $3176 \pm 5$ Ma and $3154 \pm 5$ Ma. This is comparable to an age of reported by Tucker et al. (1999) for tonalitic gneiss from the island of Nosy Boraha and is also of similar at to the metarhyolite of Tucker et al. (in press) dated at $3178 \pm 2$ Ma, apparently constraining the age of a more widespread magmatic event at this time. Complex reworking is indicated by the analysis of zircon rims which indicate thermal events at $\sim 2.73$ and $2.49$ Ga.

4.3. Granitic rocks

4.3.1. Masoala Suite – Southern sub-suite

Sample BT/06/2/14 (Nosy Mangabe Granite) comprises fairly homogeneous, coarse-grained granodiorite, with some patchy coarse K-feldspar segregations, collected from the northern part of Nosy Mangabe island [757326 1175639].

24 analyses were conducted on 24 zircons and indicate low $^{238}\text{U}$ and $^{232}\text{Th}$ in the narrow range 41–320 and 41–554 ppm, respectively, with the exception of analysis 17, which has $^{238}\text{U}$ and $^{232}\text{Th}$ contents of 835 and 1625 ppm, respectively. $^{207}\text{Pb}/^{206}\text{Pb}$ ratios are between 0.21 and 2.01, characteristic of magmatic zircon. Zircon analyses yield a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of $2528 \pm 3$ Ma (MSWD = 1.36, excluding the most discordant data point), which provides the best age estimate for the crystallisation of the granodiorite (Fig. 6).

Sample BT/06/2/70 (Onive Granite) is a homogeneous, equigranular, medium-grained, unfoliated biotite granite collected from a river section at [819488 1194465].
23 analyses were conducted on 23 zircons and indicate highly variable \( f_{206} \) values. \( f_{206} \) values appear to be positively correlated with U and Th content, indicating Pb-contamination effects from radiogenic damage. The same radiogenic damage to the crystal lattice is also likely to permit continuous and increasing radiogenic Pb-loss through diffusion. \( f_{206} \) values are in the range 0.01–19.88%, but the majority of analyses contained no more than 2% Pb. U and Th are in the ranges 109–13,824 and 36–91,464 ppm, respectively, with Th/U between 0.04 and 6.84. The two zircons with the highest U, Th and \( f_{206} \) values (5 and 14) were discarded from the dataset, as were analyses 17, 19 and 22 based on high \( f_{206} \) values.

Three zircons analyses provide a concordia age of 2532 ± 7 Ma (MSWD = 1.7) taken to be the best estimate for the age of crystallisation (Fig. 6). The remaining analyses define a poorly defined trend indicating Pb-loss associated with the Pan-African event at ~550 Ma.

Sample BT30192 (Tonalite hosted by Nosy Mangabe Granite): comprises foliated, coarse-grained, heterogeneous tonalite to granodiorite with disrupted mafic enclaves and schlieren that was collected from the coast south of Mananara Nord [72022 107311].

19 analyses were conducted, including 1 core–rim pair and 15 oscillatory zoned domains. \( f_{206} \) values are low, between below background and 0.65%. U and Th are in the range 83–750 and 36–434 ppm, respectively, with Th/U ratio defining a range 0.16–0.72 consistent with magmatic zircon. The data range from 3 to 28% discordant, with discordance directly related to U and Th content indicating Pb-loss due to radiogenic damage.

A single core analysis records a xenocrystic component with a minimum age of 3187 ± 15 Ma. Remaining zircon analyses yield two discordia trends one with an upper intercept at 2570 ± 18 Ma and lower intercept at 792 ± 130 Ma (MSWD = 0.56; Fig. 6), the other with an upper intercept of 2546 ± 11 Ma and lower intercept of 526 ± 39 Ma. We interpret the data to indicate that the tonalite crystallised at 2570 ± 18 Ma with a second phase of growth at recorded at 2546 ± 11 Ma. Xenocrysts formed at around 3187 ± 15 Ma reflect an older crustal source component within error of the 3178 ± 2 Ma metahylolite dated by Tucker et al. (in press). Poorly constrained lower intercepts of 792 ± 130 Ma and 526 ± 39 Ma suggest the sample was affected by Pb-loss during Late-Neoproterozoic to Cambrian times.

Sample BT30167 (Amanobo Granite) is a coarse-grained, K-feldspar–phyric augen gneiss collected from river outcrops at [687177 0966223].

23 spots were analysed, including 9 core–rim pairs, 2 cores and 3 rims. Proportions of common Pb are low for 22 analyses, between below background and 0.76%, but is extremely high (14.30%) for analysis on core 17. U and Th content are variable and are in the range 244–3865 and 69–5847 ppm, respectively, defining a narrow range of Th/U ratios between 0.06 and 0.67 (excluding one high Th analysis (C7), which has a Th/U ratio of 5.25).

The data are from 2% to 90% discordant (excluding the high \( f_{206} \) core 17), with discordance directly related to U and Th content indicating Pb. Regression of zircon core analyses yields an upper intercept age of 2531 ± 16 Ma (MSWD = 4.1; Fig. 6) interpreted to be the best estimate for the crystallisation of the zircon cores in the protolith. Analyses of zircon rims yield an upper intercept of 2514 ± 13 (MSWD = 3.0), probably due to the effects of a later thermal overprint. Scatter within the analyses is interpreted to reflect complex Pb-loss. Imprecise lower intercepts (246 ± 150 Ma for cores and 38 ± 150 Ma for rims) indicate more recent Pb-loss.

4.3.2. Masoala Suite – Northern sub-suite

Sample GW175 (Sarandranano Granodiorite) is a medium- to coarse-grained granodiorite with scattered K-feldspar megacrysts, collected at [800291 1224969].

Twenty analyses were conducted on 20 zircon grains, and indicate variable \( f_{206} \) values. Seven analyses recorded over 1% Pb, with values up to 21.34%. The remaining analyses have \( f_{206} \) values between 0.01 and 0.91%. U and Th are also quite variable, with higher U + Th values correlated with high Pb. Because of the variable U and Th contents, Th/U ratios have a wide range, 0.42–4.60, but the majority are between 0.5 and 1.5, consistent with magmatic zircon.

The data define a widely spread linear array, which allows a regression with upper intercept at 2542 ± 7 Ma (MSWD = 3.5; Fig. 7) with the lower intercept of the regression of all data points indicative of recent Pb-loss.

Sample GW126 (Bezavona Granite) comprises homogeneous, very coarse- to coarse-grained, foliated, pink granite collected from an outcrop at [786276 1238383] occupying a tectonic window through the Sahantaha Group of the Bemarivo Domain at the northern margin of the Antongil Craton.

28 analyses were conducted on oscillatory zoned domains of 28 zircon crystals. \( f_{206} \) values range from 0 to 26.08%, with higher proportions of common Pb related to higher U + Th content, and thus Pb-gain through radiogenic damage of the crystal lattice. The majority of analyses have less than 1% Pb, U and Th are in the ranges...
calculation of a concordia age of 2529 ± 10 Ma (MSWD = 2.5; Fig. 7) is taken to be the best estimate for crystallisation of the granite. A lower intercept of 553 ± 68 Ma, although imprecise, indicates Late-Neoproterozoic Pb-loss from the zircon.

In summary, new analyses from six samples record Neoarchaean ages for crystallisation of the various phases of the Masoala Suite, ranging between 2570 ± 18 Ma and 2528 ± 3 Ma with the older dates derived from the northern sub-suite (between 2570 ± 18 Ma and 2542 ± 5 Ma). Possible xenocrysts preserved in one sample indicate inheritance from Mesoarchaean source components of 3187 ± 15 Ma or incorporation of older xenoliths during the main magmatic episode. These are interpreted as having been derived from the older Archaean nucleus represented by the Nosy Boraha Suite. Analysis of zircon rims in one sample provides evidence for a later, Neoarchaean metamorphic or thermal event at 2514 ± 13 Ma, probably recording a local, late thermal perturbation associated with the waning effects of the emplacement of the suite. Some lower intercept ages provide evidence for poorly constrained Late-Neoproterozoic to Early Cambrian and Recent Pb-loss events.

5. Geochemistry of the granitoids

A total of 30 samples from the Antongil Craton were analysed for major and trace elements and REE. Fresh rock samples selected for geochemistry were crushed and milled in agate at the DMG Laboratory of the Ministry of Energy and Mines in Antananarivo, and at ACTLABS, Canada (by their Code 4 Lithoresearch package). Major oxides and some trace elements were analysed by Li-metaborate/tetraborate fusion with an ICP-OES analysis, and these sample solutions were further diluted and spiked for ICP-MS analysis. The samples were run for major oxides and selected trace elements on a combination simultaneous/sequential Thermo Jarrell-Ash ENVIRO II ICP or a Spectro Cirros ICP, and for other trace elements on a Perkin Elmer SClXI ELAN 6000 or 6100 ICP-MS. Tabulated analytical results are available as supplementary materials. The overall inaccessibility, dearth of outcrop and difficulty in collecting suitable material where deep tropical weathering is widespread, have resulted in an uneven geographic and lithotratigraphic distribution of analyses.

A plot of SiO2 versus total alkalis (Fig. 9a) shows that the geochemical classification of these samples broadly matches the names based on petrography. The northern sub-suite of the Masoala Suite comprises solely granitic and granodioritic compositions, but the southern sub-suite also includes a few lower silica, dioritic to tonalitic rocks (and may also reflect undifferentiated older enclaves of Nosy Boraha Suite lithologies). In common with many Archaean terranes there are few rocks of intermediate composition (SiO2 < 64%; Martin et al., 2005). The Masoala Suite granitoids from both sub-suites spread across the high-K and medium-K fields on a K2O–SiO2 plot (Fig. 9b).

Recent studies of the geochemical distinctions between Archaean granitoids and younger, late- to post-Archaean granitoids (e.g. Martin, 1999; Moyen et al., 2003) have shown that TTG gneisses have very fractionated REE patterns, with (La/Yb)n values typically >10, whereas post-Archaean granitoids typically have much less fractionated (La/Yb)n values (Martin, 1999). The dated orthogneiss sample from the Nosy Boraha Suite (MH73) is a typical TTG gneiss, with a very strongly fractionated REE pattern (La/Yb)n ~ 120 (Fig. 9c). Sample DS30507 does not have the same fractionated REE pattern, and preserves a composition that probably reflects anatectic partial melting that has affected this sample. The granitoids from the Masoala Suite show a wide range of REE patterns, with (La/Yb)n varying from 3 to 90.
Fig. 9. (a) Total alkalis versus silica plot for samples from the Masoala and Nosy Boraha suites. Classification fields from Gillespie and Styles (1999) after Le Bas et al. (1986). Alkalic/sub-alkalic division from Miyashiro (1974); (b) K2O versus SiO2 plot for samples from the Masoala and Nosy Boraha suites. Fields from Le Maitre (1989); (c) plot of chondrite-normalised La/Yb versus Yb for samples from the Masoala and Nosy Boraha suites. Normalising factors from McDonough and Sun (1995). Field of typical TTG from Martin (1999); (d) plot of chondrite-normalised Ce/Yb versus K/Na for samples from the Masoala and Nosy Boraha suites, showing how the samples can be split into three distinct groups. Normalising factors from McDonough and Sun (1995).

In general, the samples from the northern sub-suite have $(\text{La/Yb})_n > 20$, whereas those from the southern sub-suite show a wider range, being generally more enriched in the HREE. The less fractionated granitoids typically have low K/Na ratios, whereas the more fractionated, apparently TTG-like, samples are more potassic (Fig. 9d). This is the reverse of what might be expected for TTG gneisses, which are typically sodic in composition (Moyen et al., 2003).

The Masoala Suite intrusions can be broadly divided into three geochemical groups, based on their REE patterns and K/Na ratios: (A) A high-K group, dominated by samples from the southern sub-suite, has K/Na > 1, and also shows high levels of REE fractionation ($\text{La}_n/\text{Yb}_n$ most commonly > 50, though one sample is much less fractionated); (B) a main group has K/Na between 0.2 and 0.9, and moderately fractionated REEs ($\text{La}_n/\text{Yb}_n$ 5–50); (C) a subordinate group of very sodic composition (K/Na < 0.1) but have fairly flat REE patterns ($\text{La}_n/\text{Yb}_n < 5$). The main group includes samples from the slightly older northern sub-suite and the larger southern sub-suite. Samples from the high-K group (group A) are also enriched in the other LILE, including Rb, Ba and Th. They have pronounced negative Nb-Ta anomalies on a spider diagram, and smaller Sr and Ti anomalies (Fig. 10a). They are strongly enriched in the LREE over the MREE and HREE, but lack strong Eu anomalies (Fig. 10b).

In the Dharwar Craton of southern India four main types of Archaean granitoid have been identified, with which the rocks of the Masoala Suite can be compared: (1) TTGs, formed by melting of basaltic material; (2) Sanukitoids (high Mg granitoids), formed when TTG-type magmas interact with mantle peridotite; (3) Closepet-type granites, formed by mixing of magma derived from enriched mantle with magma from crustal sources; (4) K-rich biotite granites, formed by remelting of older TTGs (Moyen et al., 2001, 2003).

Samples from the high-K group (group A) of the Antongil Craton show many similarities to the K-rich biotite granites (4 above) of the Eastern Dharwar Craton (Moyen et al., 2003). Samples of the Dharwar Craton K-rich biotite granites have been dated at ca. 2540 Ma (Jayananda et al., 2000), within error of the age of 2542 ± 5 Ma for the dated sample from the Antongil high-K group (GW175). It should be noted that the high-K granitoids of the Antongil Craton have somewhat more highly fractionated REE patterns than the typical Dharwar biotite granites, but analogies for this are found in the low-Y monzogranite of the Arsikere pluton in the Western Dharwar Craton (Jayananda et al., 2006).

The main group of Masoala Suite granitoids (group B) includes many samples that show similarities to the sanukitoids and Closepet-type granites (2 above) in the Dharwar Craton (Moyen et
Fig. 10. Primitive mantle-normalised trace element, and chondrite-normalised rare earth element patterns for selected samples from the high-K group (a, b), main group (c, d) and high-Na group (e, f) of the Masoala Suite. Normalising factors from McDonough and Sun (1995).

6. Event history of the Antongil Craton

The oldest event recorded by the Antongil Craton is the crystallisation of a TTG suite during the Mesoarchaean, recorded by the Nosy Boraha Suite and by xenocrystic zircons preserved in samples from the Masoala Suite. Including the analyses of Tucker et al. (1999, in press), these document crustal growth between 3320 ± 14 Ma to 3231 ± 6 Ma and 3187 ± 2 Ma to 3154 ± 5 Ma. The distribution of Mesoarchaean ages is restricted to the southern part of the craton, suggesting the presence of a putative Mesoarchaean nucleus formed during a major early crust-forming event. Such older nuclei, surrounded by accreted Neoarchaean, granoididominated crust are common in Archaean terranes. This resembles the ca. 3000–3400 Ma Karnataka nucleus of the (West) Dharwar Craton, largely hosted by gneisses and granites of Neoarchaean age and together included in the Peninsular Gneisses; e.g. Beckinsale et al. (1983), Radhakrishna and Naqvi (1986), and Jayananda et al. (2000).

A subsequent episode is indicated by overgrowths on zircon from the Nosy Boraha Suite at 2749 ± 12 Ma and 2729 ± 18 Ma. While there is little evidence for the nature of this event in the Antongil Craton, an Sm–Nd isochron age of 2747 ± 15 Ma from the Ingadthal Metavolcanics of the Dharwar Craton (Kumar et al., 1996) provides some evidence for tectonothermal activity elsewhere in the region at that time.

Following the first main magmatic event, a succession of volcano-sedimentary rocks (Mananara Group and Ammodiriana Formation) was deposited and a suite of mafic dykes intruded into the earlier TTG suite. Together, these were then metamorphosed and widely intruded by granitoids during the subsequent plutonic phase. These volcano-metasedimentary rocks are now preserved as rafts and mega-xenoliths within the Masoala Suite, as well as a broader region of exposure in the western part of the craton. Recent findings of Tucker et al. (in press) indicate that these were deposited at around 3178 ± 2 Ma. These findings are supported by this study which indicates deposition between 3176 ± 6 Ma and before 2597 ± 49 Ma, given by the ages zircon yielded by a leucosome layer in migmitic paragneiss. Speculatively, we interpret these rocks as fragments of the host rocks to the younger components of the Neoarchaean phase of igneous intrusion. They may have been deposited unconformably on the Mesoarchaean intrusive assemblage, which is given credence by the presence of ca. 3176 Ma inherited components, and are of equivalent age to supracrustal rocks of the western Dharwar craton (Nutman et al., 1992, 1996). The sediments were clearly strongly deformed and suffered partial melting to form migmatite, recording a tectonothermal event that broadly overlaps the age of the earliest Neoarchaean granitic intrusions. Collins et al. (2003c) dated a sample of metasediment from Fenerive Est, from the Ammodiriana Formation as it is delimited by this study. They interpreted the youngest concordant zircon core analysed from the sample as providing an upper estimate for the age of deposition at 710 ± 10 Ma and went on to conclude that these rocks formed part of a suture
zone between the Antongil and Antananarivo cratons. Similarly a single concordant analysis from north of Fenerive Est (his study), dated at 763 ± 13 Ma could be taken to support this interpretation. However, the clear field relationships presented by Tucker et al. (in press), illustrated that the metasediments from this area were deformed and penetratively deformed before 2502 ± 8 Ma, the age of a crosscutting dyke. Clearly there is a considerable contrast between these two interpretations which is not fully resolved. Given the similarity between our new data and those of Tucker et al. (in press) we prefer to interpret the Neoarchaean ages as reflecting neocrystallisation during metamorphism at that time. However, in the light of these new data, a thorough re-examination of the localities studied by Collins et al. (2003c) is probably required to fully unravel this controversy.

A second major crust-forming and continental accretion event is recorded by the crystallisation of voluminous granitic magmas of the Masoala Suite dated between 2570 ± 18 Ma and 2512.7 ± 0.4 Ma. Tucker et al. (in press) attribute this event to assembly with the adjacent Antananarivo Domain in which magmatism and metamorphism of the same age is preserved. This widespread event is also recorded in the ages of xenocrysts preserved in Palaeoproterozoic mafic intrusions of the Ankavanana Suite and thermal events affecting the older Mesoarchaean gneisses of the Nosy Boraha Suite. The widespread and long-lived magmatism is characterised by heterogeneous igneous rocks, suggesting repeated batch additions, widespread magma mingling and mixing, localised partial anatexis and re-homogenisation by processes of migmatisation.

As in many cratonic areas (e.g. the Zimbabwe Craton, see Nisbet et al., 1981), this major Neoarchaean magmatic and thermal event appears to have been associated with the main phase of craton stabilisation, marking the terminus of an important Archaean orogeny, and subsequent events were limited to localised deformation and smaller volumes of magmatism.

Overall, the geochemistry of the Neoarchaean granitoids of the Masoala Suite suggests the craton evolved through both reworking of older TTG of the Mesoarchaean nucleus and addition from younger source components. This is consistent with the interpretation of Tucker et al. (1999) who reported eNd values ranging between +1.4 and −4.1 for dated samples of Neoarchaean rocks ranging in composition from gabbro to granite which they interpreted to reflect mixing between a Mesoarchaean mantle component of similar composition to the Nosy Boraha Suite with a Neoarchaean depleted mantle component. Together these features suggest that accretion of new crust was important at that time. Some of the Neoarchaean granitoids, particularly in the southern sub-suite, are analogous to K-rich biotite granites from the Dhawar Craton and represent reworking of older TTGs (Moyen et al., 2001, 2003). The recognition of some similar magma types provides additional credence to proposed links between the two terranes (e.g. Tucker et al., 1999; Kröner et al., 2000; Collins and Windley, 2002; Collins et al., 2003b; Collins, 2006).

In the Masoala Suite, the Neoarchaean intrusive episode appears to have been initiated by the intrusion of granodiorite–dominated facies preserved in the northern sub-suite, and as tonalitic enclaves within the southern sub-suite, intruded between 2570 ± 18 Ma and 2542 ± 5 Ma. This was followed by more widespread intrusion of the monzogranitic facies preserved in the southern sub-suite. Paquette et al. (2003), on the basis of three granitoid analyses from near Mananara Nord (Fig. 1), went on to suggest that this younger episode could be subdivided into sequential magmatic pulses emplaced at around 2532 Ma, 2523 Ma and 2513 Ma. Our data provide a wide range of dates suggesting that this interpretation has little statistical validity. Rather, the granitoids are the product of repeated intrusion during a protracted magmatic event. There is insufficient data to assess whether Neoarchaean magmatism in the Antongil Craton is characterised by punctuated intrusive events or by a continuum of batch additions.

A Palaeoproterozoic magmatic event, widespread in the northern part of the Antongil Craton, is recorded by a maﬁc dyke swarm and the associated gabbronitroic intrusions of the Ankavanana Suite, dated at 2147 ± 6 Ma. Although this event is of similar age to widespread dyke emplacement and formation of Palaeoproterozoic orogenic belts that took place between ca. 2.1–1.8 Ga (Ernst and Buchan, 2001), the NE-trending orientation of the Ankavanana Suite dykes is very approximately similar to ENE-trending basic dykes from the Dhawar Craton recently dated at 2367 ± 1 Ma (Halls et al., 2007). Although the dates appear to rule out intrusion during a single event, their orientation may point toward a related orientation with regard to crustal stress during that period. This is likely to be associated with an extended period of elevated magmatic activity in early Palaeoproterozoic times, variously interpreted as including widespread mantle plume development (Ernst and Buchan, 2001) or as continental collision culminating in the assembly of a postulated pre-Rodinia Palaeoproterozoic supercontinent (e.g. Nuna of Hoffman, 1997; Columbia of Zhao et al., 2002). The implications of the Ankavanana Suite for the evolution of the Antongil Craton are unclear, but it appears to represent a general period of crustal extension.

Both Cox et al. (1998) and Collins et al. (2003a) argued that the absence of ca. 820–730 Ma magmatic rocks in the Antongil Craton, compared to their widespread distribution in the adjacent Antananarivo Craton and intervening Neoproterozoic Manampotsy Belt to the west in Madagascar (Fig. 1), provided evidence that the cratons were widely separated at that time. A poorly constrained lower intercept of 792 ± 130 Ma from one sample of Neoarchaean granite as well as the possible metamorphic age of 763 ± 13 Ma from the Ambodiriana Formation raises the possibility of Late-Neoproterozoic disturbance in the Antongil Craton and consequently, that the distribution of Neoproterozoic magmatism may not be a secure criteria upon which to base palaeogeographic reconstruction of the Archaean terranes of Madagascar. Evidence for Late-Neoproterozoic to Cambrian age Pb-loss at 553 ± 68 Ma seen in one sample of Neoarchaean granite is interpreted as a manifestation of widespread tectonothermal activity during the East African (Pan-African) orogeny. The timing of this event compares favourably to the age of metamorphism, magmatism and deformation in both the Antananarivo Craton and Bemarivo Belt (e.g. Buchwald et al., 2003; Collins et al., 2003a,b; Thomas et al., 2009; Goodenough et al., 2014) and provides additional evidence for the shared events with the adjacent terranes during their assembly to form the East African Orogen.

Acknowledgements

The authors would like to thank the BGS, USGS and Malagasy colleagues who contributed to field mapping during the field seasons of 2005–2007. R. Tucker and F. Maldonado (USGS) are thanked for greatly improving an earlier draft of this manuscript and I. Fitzsimons and B. Moine are thanked for providing thoughtful reviews. The Projet de Gouvernance des Ressources Minérales (Madagascar) are acknowledged for providing permission to report data herein. BGS authors publish with the permission of the Executive Director BGS (Natural Environment Research Council). The Perth Consortium SHRIMP facilities at the John de Laeter Center for Mass Spectrometry (Curtin University of Technology) are funded by the Australian Research Council.

Appendix A. Supplementary data
Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.precamres.2010.07.006.