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Mesoproterozoic subduction under the eastern edge of the Kalahari-Grunehogna Craton preceding Rodinia assembly: the Ritscherflya detrital zircon record, Ahlmannryggen (Dronning Maud Land, Antarctica)

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Abstract

The ~ 2000 m thick clastic and volcaniclastic sedimentary rock pile of the Mesoproterozoic Ritscherflya Supergroup is located near the eastern margin of the Archaean Grunehogna Craton of Dronning Maud Land (East Antarctica). The sedimentary rocks were deposited proximal to an active volcanic arc formed during subduction prior to the assembly of the Rodinia supercontinent. In this study, we investigated internal zonation and U-Pb ages of detrital zircon grains from all formations of the Ahlmannryggen and Jutulstraumen groups of the Ritscherflya Supergroup. Our results show an age distribution with a dominant age peak at ~ 1130 Ma, close to the sedimentation age of the sedimentary rocks ($\sim 1130 - 1107$ Ma), which strongly supports the model of deposition of the sediments in a convergent margin setting. Older peaks in the Ritscherflya sedimentary rock zircon spectrum with ages up to 3445 ± 7 Ma that were also identified in samples from the Grunehogna Craton basement reflect tectono-magmatic events in the Kalahari Craton. This provides further evidence for the Archaean and Proterozoic connection of the Grunehogna province to the African Kalahari Craton.

Parts of the Mesoproterozoic volcanic arc were located on Archaean cratonic basement ($\sim 2800 - 3450$ Ma), whereas other parts tapped late Palaeoproterozoic crust (~ 1750 Ma). This is evident from a number of inherited Archaean and Proterozoic cores in zircons with Stenian rims. The Ritscherflya zircon record, therefore, supports models of the eastern margin of the Kalahari-Grunehogna Craton that include inward subduction with an active continental margin prior to collision in Dronning Maud Land. The intercalation of the clastic sedimentary rocks with volcaniclastic materials strongly support the interpretation of a very proximal volcanic source.

The sedimentary rocks were affected by regional low-grade metamorphism during the collisional orogeny related to Rodinia assembly and during the Pan-African orogeny related to the assembly of Gondwana. This is evident from metamorphic recrystallisation of zircon at 1086 ± 4 Ma and from discordancy of many grains pointing to late Neoproterozoic to early Phanerozoic lead loss.

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Keywords: detrital zircon; geochronology; Dronning Maud Land; Rodinia; Mesoproterozoic

Introduction

The reconstruction of the palaeogeography and supercontinent amalgamation processes in the Precambrian is generally guided by the age of magmatic and metamorphic rocks in orogenic belts, which formed along the sutures of colliding continents or smaller terranes (e.g., Wareham *et al.*, 1998; Dalziel *et al.*, 2000). Yet, the investigation and interpretation of these belts becomes increasingly difficult with increasing age of the orogenic cycles, due to metamorphic overprint, fragmentation by subsequent rifting processes, erosive loss, and covering by younger deposits or by ice. An alternative and important recorder of geodynamic processes are clastic sediments that are fed from the eroding orogenic belts and are deposited on stable cratonic platforms, where they may escape erosion and high-grade metamorphism for billions of years. These clastic sediment deposits generally contain abundant detrital zircon, which provides an age record of the eroded orogenic belts, reflecting a large number of rock types. The age spectra of detrital zircon recovered from sedimentary basins can be used to distinguish between different tectonic settings in which the sediments were deposited, such as convergent margins, collisional orogens or extensional settings (von Eynatten & Dunkl, 2012; Cawood *et al.*, 2012).

The supercontinent Rodinia formed by convergence and collision of all the major landmasses between $\sim$ 1200 and $\sim$ 950 Ma, i.e. in the late Mesoproterozoic (Hoffman, 1991; Li *et al.*, 2008). Key evidence for the collisions is found in the late Mesoproterozoic orogenic belts, which span thousands of kilometres through North and South America, southern Africa, Australia, Asia and East Antarctica. Yet, the paleogeographic reconstruction of Rodinia is still uncertain, and at least three different configurations have been discussed (Li *et al.*, 2008). Issues arise in part from the uncertainties in terrane boundaries within East Antarctica and possible connections to the African Kalahari Craton. At least one major late Mesoproterozoic (Stenian) suture must be located in Dronning Maud Land (DML; East Antarctica), but its location and the extent of possible crustal blocks are still enigmatic (Jacobs *et al.*, 2008*a*).

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Parts of western DML were part of the Kalahari-Grunehogna Craton (KGC; Groenewald *et al.*, 1995; Jacobs *et al.*, 2008*b*; Marschall *et al.*, 2010), but the eastern limit of the KGC immediately prior to the assembly of Rodinia (at $\sim$ 1200 Ma) is unknown. Hence, the Mesoproterozoic suture(s) between Kalahari and the terrane or continent it collided with have not yet been located. The south-western margin of the KGC (the Namaqua-Natal sector of South Africa) was affected by the accretion of juvenile Palaeo- and Mesoproterozoic island arcs (Jacobs *et al.*, 2008*b*), but it is unknown whether or not similar accretion processes occurred at its eastern margin, i.e., the area exposed in DML today. Contrasting models have been put forward for the tectonic regime of the eastern margin of the KGC. Jacobs *et al.* (1996, 2008*b*) and Bauer *et al.* (2003) suggested that the KGC had a passive margin with subduction away from the craton and subsequent accretion of Proterozoic island arcs onto the southern and eastern margin. Basson *et al.* (2004) advocated a similar model based on clastic sedimentary rock analyses, yet introduced two subduction zones and a more complex subduction-collision history. In contrast, Grosch *et al.* (2007) argued, based on the geochemistry of mafic dykes, that the eastern margin of the KGC formed an active continental margin with subduction of oceanic crust underneath the KGC. Westward subduction underneath the eastern margin of the KGC was also advocated by Grantham *et al.* (2011), based on Nd and Sr isotope signatures of granitic gneisses from the Mozambique and Maud belts.

In this paper, we present U-Pb dates of detrital zircon from the Mesoproterozoic Ritscherflya Supergroup, a sequence of clastic and volcaniclastic sedimentary rocks deposited on the eastern margin of the KGC. The zircons provide strong evidence for the deposition of the sediments at an active continental margin in a convergent setting, supporting the models of Grosch *et al.* (2007) in which subduction beneath the eastern margin of the KGC created a continental volcanic arc. The sedimentary rocks were overprinted by (sub-)greenschist facies metamorphism, while the continental margin hosting the dominant portion of the arc itself was affected by high-grade metamorphism during subsequent Stenian and Pan-African collision events.

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Geological background

The Ritscherflya Supergroup is a Mesoproterozoic clastic to volcaniclastic sedimentary rock sequence interbedded with volcanic sequences and intruded by syn-sedimentary to syn-diagenetic mafic sills (Roots, 1953; Allsopp & Neethling, 1970; Wolmarans & Kent, 1982; Moyes *et al.*, 1995; Basson *et al.*, 2004). It is located on the eastern margin of the Archaean Grunehogna craton in western DML of East Antarctica (Fig. 1).

The Grunehogna Craton (GC) is thought to extend from the Weddell sea at $\sim 15°$W for 350 km to the Pencksøkket-Jutulstraumen glaciers at $\sim 2°$W (Fig. 1). Palaeogeographical reconstructions show that it forms a fragment of the Archaean to Palaeoproterozoic Kalahari Craton of southern Africa that was left attached to Antarctica during the Jurassic breakup of Gondwana (e.g., Smith & Hallam, 1970; Dietz & Sproll, 1970; Martin & Hartnady, 1986; Groenewald *et al.*, 1991, 1995; Moyes *et al.*, 1993; Jacobs *et al.*, 1998, 2008*b*; Fitzsimons, 2000; Marschall *et al.*, 2010). Evidence for this reconstruction comes from a wealth of geochronologic, palaeomagnetic, structural, petrologic and geochemical data. Marschall *et al.* (2010) have argued, based on zircon age populations and Hf isotopic signatures that the Grunehogna craton formed the eastern extension of the Swaziland Block of the Kaapvaal Craton since at least 3.10 Ga and possibly as early as 3.75 Ga. Basement outcrops of the GC are almost absent and are limited to a small exposure of granite at Annandagstoppane (Roots, 1953; Barton *et al.*, 1987; Marschall *et al.*, 2010). The crystallisation age of the granite is $3,067 \pm 8$ Ma, with inherited grains with ages of up to 3.43 Ga and Eoarchaean hafnium model ages (Marschall *et al.*, 2010).

The GC borders the high-grade metamorphic Maud belt to the east and south (Fig. 1), which comprises meta-igneous and meta-sedimentary rocks metamorphosed at amphibolite to granulite-facies grade during the Mesoproterozoic ($1,090 - 1,030$ Ma) orogeny related to the assembly of the Rodinia supercontinent (Arndt *et al.*, 1991; Jacobs *et al.*, 2003*a*, 2008*b*; Board *et al.*, 2005; Bisnath *et al.*, 2006). Parts of the high-grade Maud belt were reactivated in a second orogenic event in the late Neoproterozoic/early Phanerozoic

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(600 – 480 Ma) "Pan-African" orogeny leading to the assembly of Gondwana (Groenewald *et al.*, 1995; Jacobs *et al.*, 2003*a*,*b*, 2008*a*; Board *et al.*, 2005; Bisnath *et al.*, 2006). Palaeogeographic reconstructions correlate the Maud Belt with the Namaqua-Natal Belt at the southern margin of the African Kalahari Craton and with the Mozambique Belt on the Kalahari Craton's eastern margin (e.g., Groenewald *et al.*, 1991; Arndt *et al.*, 1991; Wareham *et al.*, 1998; Jacobs *et al.*, 2003*a*, 2008*b*).

Importantly, some sections of the Maud Belt closest to the GC comprise bimodal metavolcanic rocks with geochemical signatures of subduction-related volcanic-arc magmatic rocks (Groenewald *et al.*, 1995; Grantham *et al.*, 2011). Bimodal volcanism and felsic magmas are generally more abundancant in continental arcs than in island arcs, especially where subduction is less steep. Examples include the Cascades in Oregon in the Tertiary (e.g., Priest, 1990) and the Central Andes in Argentina (e.g., Petrinovic *et al.*, 2006). The bimodal Jutulrøra metavolcanic gneisses in the western H.U. Sverdrupfjella[1] occur on the eastern edge of the Jutulstraumen glacier (Fig. 1) and were interpreted as the remnants of a Mesoproterozoic continental arc (Groenewald *et al.*, 1995; Grantham *et al.*, 2011). The arc magmatism in the H.U. Sverdrupfjella, Kirwanveggan and Heimefrontfjella has been dated to 1170 – 1120 Ma (Wareham *et al.*, 1998; Board *et al.*, 2005; Jacobs, 2009; Grantham *et al.*, 2011). The most concordant, oscillatory zoned zircon grains in the metavolcanic Jutulrøra gneiss unit revealed an age of 1134 ± 4 Ma that is interpreted as the age of arc volcanism (Grantham *et al.*, 2011).

The Ritscherflya Supergroup is exposed in the Ahlmannryggen and Borgmassivet nunataks and comprises clastic and volcaniclastic sedimentary rocks with a total estimated thickness of ~ 2000 m (Wolmarans & Kent, 1982). These were deposited between 1130 Ma and 1107 Ma in a shallow marine to braided river system (Wolmarans & Kent, 1982; Perritt, 2001; Frimmel, 2004) and subsequently intruded by large (up to 400 m thick) mafic sills prior to diagenesis (Krynauw *et al.*, 1988; Curtis & Riley, 2003). The sills have an intrusion age of $1,107 \pm 2$ Ma and have been correlated with the coeval mafic sills in the Umkondo

[1] The H.U. Sverdrupfjella mountain range was named after Norwegian oceanographer and meteorologist Harald Ulrik Sverdrup (1888–1957), who was the chairman of the first expedition to DML, the 1949–1952 Norwegian-British-Swedish joint expedition.

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region (Zimbabwe and Mozambique) and several other large mafic sills in the Kalahari craton based on geochemistry, palaeomagnetism and intrusion age (e.g., Smith & Hallam, 1970; Martin & Hartnady, 1986; Powell *et al.*, 2001; Jones *et al.*, 2003; Frimmel, 2004; Hanson *et al.*, 2004, 2006; Grosch *et al.*, 2007).

The Ritscherflya Supergroup has been subdivided into groups, formations and members by several workers (e.g., Neethling, 1970; Roots, 1970; Wolmarans & Kent, 1982; Bredell, 1982; Perritt, 2001). The difficulty of establishing a consistent sedimentation sequence for the entire supergroup are related to the discontinuous outcrop conditions (Fig. 2). The nunataks expose several hundreds of metres in vertical outcrop in many places, and they provide several kilometres of continuous horizontal exposure, but they are also separated from each other by typically $5-10\,\mathrm{km}$ wide stretches of ice (Fig. 2, 3a). In addition, faulting with uplift and tilting of blocks throughout the Ahlmannryggen further hampers correlation of units and the establishment of a stratigraphic column (Wolmarans & Kent, 1982; Peters, 1989; Perritt, 2001). Certain members and formations have, therefore, been regrouped or renamed by different authors depending on their working area and evaluation criteria. In this study, we mostly follow the scheme of Wolmarans & Kent (1982) with some alterations introduced by Perritt (2001), as detailed below and in Fig. 4.

The contact between the Archaean basement and the Ritscherflya sedimentary rocks is not exposed, and no basement is exposed anywhere in the Ahlmannryggen or Borgmassivet. The lower part of the Ritscherflya Supergroup is formed by the Ahlmannryggen Group, dominated by shallow marine to braided-river clastic sedimentary rocks with an estimated total thickness of $\geq 1200\,\mathrm{m}$ (Wolmarans & Kent, 1982; Perritt, 2001). In the Ahlmannryggen, this group is further subdivided into three formations, the Pyramiden, Schumacherfjellet and Grunehogna Formations (Fig. 4; Perritt, 2001). Note that Wolmarans & Kent (1982) used a slightly different nomenclature that defines the Grunehogna Formation as the Grunehogna Member and includes it into the Høgfonna Formation. The Høgfonna Formation is otherwise only exposed in the Borgmassivet. The Ahlmannryggen Group is succeeded by the Jutulstraumen Group, which is formed by the Tyndeklypa and Istind Formations (Fig. 4). These consist of volcaniclastic beds and lava flows interbedded with clastic sedimentary rocks. The youngest group in the Ritscherflya is the Straumsnutane Group exposed

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145 further to the north. It consists mainly of andesitic lava flows and pyroclastic beds with a total thickness of
146 $>$ 850 m (Wolmarans & Kent, 1982; Perritt, 2001). This group was not investigated in this study.

147 Regional metamorphism of the Ritscherflya sedimentary rocks is evident from the formation of sub-
148 greenschist- to greenschist-facies mineral assemblages including chlorite, muscovite (sericite), epidote, sil-
149 ica and calcite (Wolmarans & Kent, 1982). Locally, metamorphic biotite and actinolite were described, doc-
150 umenting a slightly higher grade of metamorphism (reported in Wolmarans & Kent, 1982). Block faulting
151 and tilting of blocks was documented for parts of the Ahlmannryggen (Peters, 1989; Perritt, 2001). Defor-
152 mation in mylonite zones near Straumsnutane has been dated to $\sim$ 525 Ma (Peters, 1989; Peters *et al.*, 1991)
153 and was probably caused by Pan-African collision and escape tectonics in DML related to the mergence of
154 parts of East and West Gondwana (Jacobs & Thomas, 2004).

155 West of the Ritscherflya at Annandagstoppane, metamorphic temperatures remained below the closure
156 temperature of the Rb-Sr isotope system in muscovite ($\sim$ 500 °C) since the Mesoarchaean (Barton *et al.*,
157 1987). However, Rb-Sr in biotite ($\sim$ 300 °C) was reset in the late Mesoproterozoic and some hydrothermal
158 activity dates to $\sim$ 460 Ma, demonstrating the influence of the two major orogenic events on the margin of
159 the craton beyond the area occupied by the Ritscherflya sedimentary rocks.

160 Investigated samples

161 The Ahlmannryggen was visited in January 2008 by a two-man party (HRM with one field assistant) forming
162 the British Antarctic Survey 2007–08 field campaign in DML. The sample localities on the nunataks expos-
163 ing the Ritscherflya sedimentary rocks in the study area are shown in the detailed map (Fig. 2). Sixteen sed-
164 imentary rock samples from the Ahlmannryggen were taken, including sandstones, siltstones, mudstones,
165 conglomerates and greywackes. Zircons from twelve samples were separated and investigated, including ten
166 clastic sedimentary rocks, one sandstone partially melted during contact metamorphism, and one volcani-
167 clastic rock (Table 1; Fig. 4). The sedimentary rocks show angular to sub-angular quartz grains ($15-70$ %
168 modal abundance) that are dissolved at their margins and are internally deformed in some samples, but

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otherwise appear fresh. Heavy minerals, such as zircon, rutile and apatite are also preserved. In contrast, the matrix of the rocks and almost all feldspar is replaced by sub-greenschist to lower greenschist-facies metamorphic phases, such as chlorite, epidote, prehnite and calcite (Table 1).

The lowermost sedimentary rock formation, the Pyramiden Formation, was sampled at its type locality, the Pyramiden nunatak (Fig. 3b) at the south-western limit of the Ahlmannryggen (locality Z7-28; Fig. 2). The formation mainly consists of greenish-grey feldspathic greywacke alternating with grey siltstone and black mudstone (Wolmarans & Kent, 1982). Two samples were investigated: Z7-28-1, a medium-grey greywacke with a relatively low proportion of detrital quartz ($\sim 15\,\%$) with grain sizes of $100-300\,\mu m$. Z7-28-2 is a pale greenish-grey greywacke with similar grain size, but slightly higher proportion of quartz ($\sim 25\,\%$) and a more intense replacement of the matrix by epidote.

The Schumacherfjellet Formation was sampled at two different places in the Grunehogna nunataks. Locality Z7-36 represents Peak 1285, while locality Z7-37 represents Peak 1390 (Fig. 2; see also Fig.21 of Wolmarans & Kent, 1982). Peak 1285 is the type locality of the Schumacherfjellet Formation. The formation comprises an alternating sequence of light-coloured sandstone (arkose, greywacke) and dark-coloured mudstone. Sample Z7-36-6 was sampled from the contact-metamorphic domain above a mafic sill that caused partial melting of the sediment (Krynauw *et al.*, 1988; Curtis & Riley, 2003). The rocks were termed 'granosediments' (Krynauw *et al.*, 1988) and comprise rounded fragments of relic sedimentary rock floating in a matrix of crystallised magma generated by in-situ melting. Sample Z7-36-8 is a finely-laminated siltstone with $\sim 25\,\%$ quartz (grain size $20-80\,\mu m$). Sample Z7-37-6 was sampled from a light-coloured sandstone layer of the Schumacherfjellet Formation at Peak 1390. It is relatively rich in quartz ($\sim 60\,\%$) with a coarse grain size ($200-500\,\mu m$).

The Grunehogna Formation was also sampled in two different localities. Locality Z7-36 represents Peak 1285 of the Grunehogna nunataks, the type locality of the formation. Locality Z7-39 is the Viddalskollen nunatak in the south-eastern corner of the Ahlmannryggen (Fig. 2). The Viddalskollen was initially arranged with the Pyramiden Formation (Wolmarans & Kent, 1982). However, based on sedimentological

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characteristics, Perritt (2001) argued that these exposures should instead be included into the Grunehogna Formation.

Samples Z7-36-2 and Z7-36-3 were taken from a sequence of interbedded conglomerates and coarse-grained, cross-bedded sandstones (Fig. 3e). The conglomerate (Z7-36-2) contains sub-angular to well-rounded pebbles of chert, quartzite and other sedimentary rocks in a reddish matrix composed of quartz ($\sim 50\,\%$) and alteration phases (epidote, chlorite and white mica). The sandstone (Z7-36-3) is dark grey in hand specimen and relatively poorly sorted with quartz ($\sim 70\,\%$) ranging from 20 to 500 µm in diameter. Altered plagioclase and metamorphic white mica and chlorite compose the matrix. The rock shows fine layers enriched in heavy minerals, such as Fe(-Ti) oxides, rutile and zircon.

The beds exposed at Viddalskollen show coarse-grained, grey sandstones interbedded with conglomerates (or breccias) with pebbles of chert, mudstone and quartzite that are typically $0.5-2\,$cm in diameter. Yet, some angular mudstone fragments reach up to 10 cm in size (Fig. 3f). Some beds show a strong greenish-yellow colour from high modes of metamorphic epidote in the rock matrix. Sample Z7-39-1 comprises a coarse-grained sandstone ($1-2\,$mm) with a layer of rounded and sub-angular mudstone and chert pebbles. The rock matrix shows abundant epidote and chlorite. Sample Z7-39-2 shows a bimodal grain size distribution with course-grained, well-rounded quartz grains ($1-2\,$mm grain diameter) in a fine-grained ($\sim 100\,\mu$m) matrix composed of quartz ($\sim 60\,\%$), fresh plagioclase and metamorphic epidote and chlorite. Larger pebbles ($0.5-2\,$cm diameter) of chert and mudstone are found in layers in the rock. Sample Z7-39-3 is a medium grey, homogeneous greywacke with $\sim 30\,\%$ fine-grained quartz ($50-100\,\mu$m) embedded in a strongly altered, fine-grained matrix composed of epidote and other, unidentified phases. The quartz shows strong resorption by the rock matrix and boundaries of detrital grains can no longer be identified.

The Tyndeklypa Formation was sampled on nunatak 1320 near Istind (Fig. 2; see also Fig.35 of Wolmarans & Kent, 1982). The sequence is composed of volcaniclastic deposits. Agglomerates and tuffs with angular clasts and blocks ranging in size from microscopic to several metres in size were described by Wolmarans & Kent (1982). Most of the clasts are of sedimentary origin, representing all the formations of

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the Ahlmannryggen Group (Wolmarans & Kent, 1982). Other clasts include fragments of volcanic rocks and rare sub-volcanic xenoliths (Fig. 3g). Sample Z7-38-1 represents a homogenous sample of the dark grey agglomerate with abundant fragments of sedimentary rocks ($1-10$ mm), volcanic rock fragments and abundant pumice and fiamme that are devitrified and replaced by chlorite, calcite and oxides with former vesicles filled by calcite, chlorite and pumpellyite. Some of the fiamme contain fresh sanidine. The matrix of the rock is composed of mineral detritus, dominated by quartz and smaller rock fragments, as well as minor plagioclase, alkali feldspar (sanidine as well as minor perthite) and rare white mica. Calcite may be detrital or an alteration phase.

The Istind Formation follows directly on top of the Tyndeklypa Formation and the contact is exposed at nunatak 1320. The Istind Formation consists of clastic sedimentary rocks (mostly quartzites) interbedded with tuffs and agglomerates. Sample Z7-38-3 represents a brownish-red quartzite that was sampled at the same nunatak (Peak 1320) as sample Z7-38-1. The rock is very homogenous and well sorted with quartz ($\sim 70\,\%$) with a grain size of $50-100\,\mu$m. The grains show resorbed edges and the matrix is composed of former feldspar (?) grains that are completely transformed to prehnite and calcite. Detrital white mica is present, but rare.

Zircon

Zircon grains in the mineral separates range from well rounded with pitted surfaces to euhedral bi-pyramidal, prismatic shapes. Grain sizes typically range from 50 to 300 µm for the long axis of the grains. Clear, colourless varieties were observed as much as yellowish and reddish-brownish grains. Many grains contain mineral inclusions, such as apatite, quartz, alkali-feldspar, albite-rich plagioclase, biotite, Fe-Ti oxides and titanite, which were likely included during magmatic crystallisation of the zircon from the host magma. Secondary mineral inclusions with ragged outlines also occur, consisting of chlorite, epidote, albite, quartz and Fe oxides, or a combination of these. These secondary inclusions probably formed during the post-depositional low-grade metamorphism that affected the sedimentary rocks. Most grains are oscillatory zoned

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from core to rim or show distinct cores overgrown by oscillatory-zoned rims.

Methods

Samples were fragmented with a jaw crusher to a $< 500\,\mu m$ grain size. Heavy minerals were enriched by employing a Wilfley table, a magnetic separator and LST heavy liquid. $1-2\,kg$ of crushed rock yielded between 5 and 200 mg of heavy minerals. Between 60 and 100 individual zircon grains per sample were then hand picked and mounted in epoxy and polished together with grains of zircon reference materials 91500 and Temora-2 (Wiedenbeck *et al.*, 1995; Black *et al.*, 2004).

Cathodo-luminescence (CL) was used as an imaging technique for the characterisation of all grains for internal zoning and the degree of metamictisation to select grains or domains of grains that were suitable for isotope analyses. The CL detector was attached to a Hitachi® scanning-electron microscope at the Department of Earth Sciences, University of Bristol.

U-Pb dating of zircons was carried out at the Edinburgh Materials and Micro-Analysis Centre (EMMAC) using the Cameca® IMS1270 ion microprobe. Analytical procedures are well established at EMMAC and were similar to those described by Schuhmacher *et al.* (1994). A 5 nA, 12.5 kV mass filtered $^{16}O_2^-$ primary beam was focused to a $30\,\mu m$ (long axis) elliptical spot. U/Pb ratios were calibrated against measurements of the 91500 proposed reference zircon, which has a $^{206}Pb/^{238}U = 0.17917$ and a $^{207}Pb/^{206}Pb$ age of 1065.4 ± 0.3 Ma (Wiedenbeck *et al.*, 1995). Sequences of unknowns were bracketed by analyses of 91500 and Temora-2. Measurements over single sessions gave a standard deviation for the $^{207}Pb/^{206}Pb$ ratio of 91500 of 0.9 % (95 % confidence limit). Analyses of the secondary, external reference standard (Temora-2) during the analytical sessions yielded a mean $^{206}Pb/^{238}U$ age of 417.6 ± 3.5 Ma (95 % confidence limit). Correction for in situ common Pb has been made using measured ^{204}Pb counts and using modern day composition of common Pb. Uncertainty on this correction is included in the calculation of errors on the U/Pb and Pb/Pb ratios. Corrections for minor changes in beam density or energy were made based on the comparison of U/Pb to UO_2/UO ratios.

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Two different measurement modes were applied (see also Ustaömer *et al.*, 2012): (1) in the regular mode 20 analytical cycles were acquired after 120 s of pre-sputtering with the magnet cycling from the masses of HfO^+ to UO_2^+, analysing HfO^+, the four Pb isotopes, $Zr_2O_2^+$, $^{238}U^+$, ThO^+, UO^+ and UO_2^+ (see also Marschall *et al.*, 2010). The total analyses time was ~ 30 min per spot. (2) The fast mode has a reduced pre-sputter time (60 s), only 8 analytical cycles, and includes fewer masses, i.e. Zr_2O^+, the four Pb isotopes, ThO_2^+ and UO_2^+. Total analysis time for one spot was ~ 7 min. The fast mode was applied in order to increase the number of analysis in the analytical session, which is necessary in a study on detrital zircon.

Data were processed offline by R.W. Hinton (Edinburgh) using an in-house data reduction spreadsheet. Plots and age calculations were made using the ISOPLOT program (Ludwig, 2003).

Results

Detrital zircon age data

A total of 586 zircon grains from twelve samples in five different formations were analysed for their U and Pb isotopic compositions by SIMS. Additionally, in 44 grains more than one zone was analysed. Out of the total of 630 analyses (43 in regular and 587 in fast mode), 113 were more than ± 10 % discordant, a further 18 showed a discrepancy of more than ± 15 % between Th/U ratios and $^{208}Pb/^{206}Pb$ for the calculated ages, and 4 more analyses required a common Pb correction of more than 5 %. The remaining 495 analyses on 471 different grains passed all of these quality tests and are considered good quality, allowing for the calculation of precise and meaningful isotope ages (see supplementary data).

The dating results show an age distribution with a dominant age peak at 1130 Ma, i.e., close to the sedimentation age (Fig. 6). In this paper, we refer to zircon grains in the age range $1100 - 1200$ Ma as "Stenian" grains, referring to the youngest period of the Mesoproterozoic era. They comprise approximately two thirds of all analysed grains or zones of grains (Fig. 7). Older age peaks include those at approximately 1370 Ma, 1725 Ma, 1880 Ma, 2050 Ma, 2115 Ma and 2700 Ma (Fig. 6). Zircon grains with near-concordant

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Palaeo- and Mesoarchaean ages ($3445-2800$ Ma) were also discovered, corresponding to the age of the KGC basement. These comprise 2.3 % ($n = 11$) of the concordant detrital population (Fig. 7). Archaean grains (> 2500 Ma) that past our quality test altogether amount to 6.8 % ($n = 32$) of the population (Fig. 7).

Most significantly, we discovered zircon grains with inherited Palaeoproterozoic and Archaean cores and Stenian, oscillatory-zoned rims. Two such cores show ages of 2804 ± 8 Ma and 3419 ± 8 Ma, respectively (1σ errors), with oscillatory-zoned rims that are slightly older (1190 ± 23 Ma) or indistinguishable (1120 ± 29 Ma) from the sedimentation age (Fig. 8). Late Palaeoproterozoic cores ($1600-1800$ Ma) with Stenian rims (Fig. 8) were found to be relatively abundant and were dated in 12 grains.

Young apparent zircon ages

A number of analyses that past the quality filter show $^{207}Pb/^{206}Pb$ ages younger than 1107 Ma outside their 1σ errors. Hence, these 31 grains appear to be younger than the age of sedimentation. Of these, 24 are between 2.5 and 9.6 % reverse concordant with $^{206}Pb/^{238}U$ ages above 1107 Ma and a relatively high proportion of common Pb that required a correction of the analyses of ≥ 1 %. These analyses are considered less accurate and cannot be taken as evidence for zircon growth after sedimentation. However, one zircon grain from the Grunehogna Formation (sample Z7-39-2, grain 25; locality Viddalskollen) shows a $^{206}Pb/^{238}U$-$^{207}Pb/^{206}Pb$ concordia age of 1086 ± 4 Ma (1σ, probability of concordance = 0.78). A fast-mode and a full analysis were completed on this grain. Both analyses were less than 1 % discordant and resulted in an identical age within error. The CL image shows a homogenous dull grey grain without oscillatory zonation (Fig. 8f). The Th/U ratio is 0.07, and the U concentration is very high ($\sim 3100\,\mu g/g$). This grain is interpreted to have recrystallised after sedimentation during (sub-)greenschist-facies metamorphism that is evident from metamorphic chlorite and epidote in this sample (Table 1). Low-temperature overgrowth and solution-precipitation of zircon in sediments under diagenetic and sub-greenschist facies conditions has recently been documented in a number of studies (e.g., Rasmussen, 2005; Hay & Dempster, 2009).

All discordant grains with $^{207}Pb/^{206}Pb$ and $^{206}Pb/^{238}U$ ages below 1200 Ma (n = 44) form a broad discor-

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dant array stretching from the age of the dominant detrital age group (1130 Ma) to the age of Pan-African metamorphism in the Maud Belt at $600-480$ Ma (Fig. 5b). Most of these discordant grains show high U contents ($\sim 1500-4000\,\mu g/g$), and some grains show low Th/U (≤ 0.1) typical for metamorphic zircon.

Age patterns of individual sedimentary rock formations

All five investigated formations show the dominant zircon population peak close to the sedimentation age, with insignificant differences in the age of the peaks between 1144 and 1125 Ma (Fig. 9). The older part of the age spectra, however, show significant differences among the different formations. The Pyramiden Formation displays a significant peak at 1355 Ma, which is absent from the Schumacherfjellet and Grunehogna Formations, and much less significant in the Tyndeklypa and Istind Formations (Fig. 9). Similarly, the late Mesoproterozoic group of zircons between $\sim$ 1850 and 1650 Ma of the Pyramiden Formation is present in the Tyndeklypa and Istind Formations, but very small or absent in the Schumacherfjellet and Grunehogna Formations (Fig. 9). The Schumacherfjellet, Tyndeklypa and Istind Formations show a distinct peak at 1880 Ma, which is absent from the other two formations (Fig. 9). All formations show peaks between $\sim$ 2050 and 2150 Ma and peaks at $\sim$ 2700 Ma (Fig. 9). Grains older that 2.7 Ga are too infrequent to allow for a statistically robust distinction between the different formations. Nonetheless, older grains were found in all formations except for the Istind Formation (Fig. 9).

Discussion

Deposition of the Ritscherflya sediments in a continental-arc setting

The age spectrum of the Ritscherflya zircons show a dominant peak that is very close to or indistinguishable from the deposition age of the sedimentary rocks (Fig. 6). This bears strong evidence for the derivation of the entire Ritscherflya sediment sequence from an active volcanic zone, or at least a tectonically highly active magmatic zone with very rapid exhumation of intrusive rocks that would have intruded shortly be-

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fore exhumation. The close proximity to an active volcanic zone is also obvious from the occurrence of volcaniclastic deposits with pumice and fiamme in the Tyndeklypa Formation (e.g., sample Z7-38-1). However, there is still disagreement on the type of tectonic regime that generated this volcanic zone and on the location of this zone with respect to the present geography (Basson *et al.*, 2004; Grosch *et al.*, 2007; Jacobs *et al.*, 2008*b*; Grantham *et al.*, 2011).

Cawood *et al.* (2012) demonstrated recently that the age spectra of detrital zircon can be used to distinguish between tectonic settings of sediment deposition. They showed that convergent settings are characterised by a large population of grains with crystallisation ages close to the deposition age of the sediment, whereas collisional and extensional settings show larger temporal gaps between those events. The interpretation of the Ritscherflya sedimentary rocks as deposits formed at a convergent margin is consistent with this scheme. Cawood *et al.* (2012) plotted the cumulative proportion of the difference between the crystallisation ages of detrital zircon and the deposition age of the sediment in which it was found, and distinguished different fields for different tectonic settings. All investigated Ritscherflya sedimentary rock samples taken together, as well as all formations taken separately fall clearly into the field of sediments deposited at convergent margin settings (Fig. 10). This demonstrates that the depositional regime of the entire Ahlmannryggen and Jutulstraumen groups of the Ritscherflya Supergroup was most likely set in a convergent margin and is inconsistent with collisional and extensional settings (Fig. 10).

Accretion of Palaeoproterozoic microcontinents or island arcs was demonstrated for the southern margin of the KGC, i.e. the Namaqua-Natal sector in Africa and the Heimefrontfjella in DML (Fig. 11; Jacobs *et al.*, 2008*b*). The southern margin, therefore, formed a passive margin with subduction away from the craton in the Palaeo- and Mesoproterozoic prior to the accretion of Proterozoic arcs or microcontinents (Fig. 11; Jacobs *et al.*, 2008*b*). A similar scenario was tentatively suggested for the eastern margin of the KGC, with the Maud Belt in DML forming a Mesoproterozoic addition to the KGC following outward subduction (Jacobs *et al.*, 2008*b*).

Juvenile Mesoproterozoic crustal segments (based on Nd model ages) are exposed in central and eastern

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DML, stretching several hundred kilometres to the east of the Jutulstraumen (see Fig. 1), which are also interpreted as accreted Proterozoic island arcs (Fig. 11; Jacobs *et al.*, 2008*b*). Yet, metavolcanic gneisses and metamorphosed mafic dykes in the western H.U. Sverdrupfjella are characterised by Palaeoproterozoic and Archaean Nd and Pb model ages (Wareham *et al.*, 1998; Grosch *et al.*, 2007; Grantham *et al.*, 2011). These gneisses and amphibolites are interpreted to have formed in a continental volcanic arc located on the eastern margin of the KGC with subduction to the west underneath the continent (Grosch *et al.*, 2007). In this model, the metavolcanic Jutulrøra gneisses of western H.U. Sverdrupfjella are the remnants of the active continental margin that was metamorphosed under amphibolite- to granulite-facies conditions during late Mesoproterozoic collision (Fig. 12; Grosch *et al.*, 2007).

The Archaean cores enclosed in Stenian rims of zircon grains discovered in the Ahlmannryggen (Fig. 8a,b) are strong evidence in support of this model. They demonstrate that the Stenian volcanic arc was indeed located on Archaean continental crust, rather than in Mesoproterozoic, intra-oceanic island arcs (Fig. 11, 12). The age spectrum of the Ritscherflya detrital zircons contain $\sim 20\,\%$ Palaeoproterozoic and Archaean grains (Fig. 7), which is evidence for a significant cratonic component in the provenance of the sediments, and the dominant Stenian peaks in the spectra demonstrate the proximal volcanic provenance. However, the Archaean cores with Stenian rims are much more significant than those separate pieces of evidence, because they show that the Stenian volcanic arc was located on an active margin of an Archaean craton (Fig. 11).

The combined evidence from the model ages in the western H.U Sverdrupfjella (Wareham *et al.*, 1998; Grosch *et al.*, 2007; Grantham *et al.*, 2011) and the zircon record presented here demonstrates that the eastern limit of the Mesoproterozoic KGC cannot be located beneath the Pencksøkket-Jutulstraumen glaciers. Instead, the craton must have extended further to the east, as has been suggested by Grantham *et al.* (2011), and parts of the high-grade Maud Belt most likely represent an overprinted section of Archaean craton (Fig. 12). The location of the suture between the KGC and the accreted arcs or microcontinents of central DML has not been identified yet. Interestingly though, a signifiant NE-striking magnetic anomaly exists between western and eastern H.U. Sverdrupfjella, and was previously interpreted as the western front of

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Pan-African metamorphism (Riedel *et al.*, 2013) or as a rift flank of the Jutulstraumen rift that formed or was reactivated during Jurassic break-up of Gondwana (Ferraccioli *et al.*, 2005). Alternatively, this large magnetic anomaly may represent the Mesoproterozoic suture between the KGC and the Proterozoic arcs or continent that collided with the KGC during Rodinia assembly. Riedel *et al.* (2013) emphasise that the anomaly separates blocks with fundamentally different magnetic signatures to the NW and SE respectively, which would be expected from a suture between an Archaean craton and accreted Proterozoic arcs.

The Kaapvaal-Grunehogna connection

The age spectrum and Hf isotopic composition of inherited zircon grains in the GC basement granite of Annandagstoppane was demonstrated to reflect well-known tectono-magmatic events in the Kaapvaal Craton (Marschall *et al.*, 2010). This forms important evidence for the connection of the GC to the Kaapvaal Craton for at least 2.5 billion years and probably longer (Marschall *et al.*, 2010). Further evidence in support of this reconstruction comes from the pre-Stenian age peaks identified in the Ritscherflya sedimentary rocks. The significant proportion of Palaeoproterozoic and Archaean grains and especially Palaeoarchaean grains as old as 3.45 Ga demonstrate that cratonic basement is more widespread in DML and not restricted to the small outcrop of granite at Annandagstoppane. Ritscherflya zircon ages older than > 3 Ga coincide well with the inherited ages found in the Annandagstoppane granite (Fig. 6), which in turn overlap with tectono-magmatic events in the Swaziland Block of the Kaapvaal Craton (Marschall *et al.*, 2010).

Younger peaks in the Ritscherflya zircon age spectrum also demonstrate the Kalahari affinity. Zircon population peaks coinciding with major tectono-thermal events in the Kalahari Craton, such as magmatism and metamorphism in the Limpopo Belt at ~ 2700 and ~ 2050 Ma, magmatism in the Bushveld Complex, and Mesoproterozoic orogenies in the Kibaran, Rehoboth and Kheis provinces (Fig. 6).

Models that propose an independent Archaean-Proterozoic history of Grunehogna and Kaapvaal Cratons and a complex Mesoproterozoic convergence history (e.g., Basson *et al.*, 2004) seem unlikely in the light of this zircon provenance study. The latest aeromagnetic studies in DML also provide strong support for the

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model of one coherent KGC in the Proterozoic (Riedel *et al.*, 2013).

The significant population of Palaeo- and Mesoproterozoic zircon grains, evident from peaks in the age histogram at approximately 1880, 1725 and 1370 Ma (Fig. 6) may not be derived from the Proterozoic orogens in Africa. Instead, they may be taken as evidence for sediment contributions from the accreted Proterozoic microcontinents and arcs in the Heimefrontfjella (Jacobs *et al.*, 2008*b*). The relatively common occurrence of Palaeoproterozoic cores in Stenian zircons in the sedimentary rocks (Fig. 8c-e) demonstrates that part of the active continental arc at $\sim$ 1130 Ma was located on crust consisting of rocks with Palaeoproterozoic crystallisation ages ($\sim$ 1750 Ma).

Provenance variability in the Ritscherflya Supergroup

The zircon age histograms shows a certain variability throughout the investigated section of the Ritscherflya Supergroup (Fig. 9). The Proterozoic age peaks related to the erosion of Proterozoic crust (possibly in the Heimefrontfjella) are very large in the lowermost formation (Pyramiden Formation), but are small or absent from the stratigraphically higher Schumacherfjellet and Grunehogna Formations (Fig. 9). This may simply be due to a geographical change of the dominant flow direction of the palaeoriver, or it may show that the Proterozoic crust was effectively eroded early during the depositional cycle. The zircon population of the Ahlmannryggen Group is increasingly dominated by Stenian zircons with increasing stratigraphic height (Fig. 10).

The Proterozoic peaks reappear in the Tyndeklypa and Istind Formations (Fig. 9). The volcanic successions of the Tyndeklypa Formation are characterised by blocks and smaller detritus from all formations of the Ahlmannryggen Group (Fig. 4). The reappearance of the Proterozoic peaks may, therefore, be explained by recycling of material from the underlying Pyramiden Formation. The sandstones of the Istind Formation may in turn contain material redeposited from the Tyndeklypa volcanic rocks.

The spatial variation of metamorphic overprint and the spatial variations in the detrital zircon record within individual formations cannot be fully evaluated from the restricted number of sample sites selected

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in this study. However, metamorphism at the Pyramiden nunatak located at the western margin of the Ahlmannryggen produced as much coarse-grained epidote and chlorite as in the sedimentary rocks of the Viddalskollen nunatak at the Ahlmannryggen's eastern margin, ~ 50 km east of Pyramiden. The detrital zircon-age pattern of the Grunehogna formation sedimentary rocks exposed at Viddalskollen is very similar to that of the Grunehogna formation rocks at their type locality 40 km to the NW.

Metamorphic overprint of the craton margin

The regional metamorphic overprint of the sedimentary rocks reached conditions of the sub-greenschist to greenschist facies (Table 1). Metamorphic recrystallisation of zircon occurred at 1086 ± 4 Ma and was most likely caused by the same collisional event that produced the contemporaneous amphibolite- and granulite-facies metamorphic overprint in the nearby Maud Belt (e.g., Jacobs *et al.*, 2003*a*; Bisnath *et al.*, 2006).

In addition, Pb loss in Ritscherflya zircon produced a discordant array during the Pan-African collisional orogeny that is also recorded in the Maud Belt as a high-grade metamorphic-magmatic event in the late Neoproterozoic to early Phanerozoic (e.g., Jacobs *et al.*, 2003*b*; Board *et al.*, 2005). This event is also evident from hydrothermal activity at Annandagstoppane (Barton *et al.*, 1987), and from mylonitic shear zones in the Ahlmannryggen itself (Peters, 1989; Peters *et al.*, 1991).

The metamorphic record of the zircon shows that thermal overprint effected the margin of the craton at a distance of at least 100 km and probably more, depending on the exact eastern boundary of the pre-Rodinian KGC (Fig. 12). The younger metamorphic event was accompanied by crustal-scale shearing and fracturing during Pan-African collision and escape tectonics in the early Phanerozoic and followed by fragmentation of the ancient craton during Gondwana break-up in the Jurassic (e.g., Jacobs & Thomas, 2004).

Conclusions

The detrital zircon record in the Ritscherflya Supergroup demonstrates that subduction under the eastern margin of the KGC produced a continental volcanic arc located on the edge of the Archaean craton (Fig. 11).

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This supports models for the Mesoproterozoic setting of DML based on the radiogenic isotope geochemistry of mafic dykes and metamorphic rocks (Wareham *et al.*, 1998; Grosch *et al.*, 2007; Grantham *et al.*, 2011) and rejects models that suggest a passive eastern margin of the KGC (Jacobs *et al.*, 2008*b*) or separate Grunehogna and Kaapvaal cratonic blocks (Basson *et al.*, 2004).

The Ritscherflya Supergroup was deposited close to the active continental arc. The zircon record in all formations is strongly dominated by grains with crystallisation ages very close to or indistinguishable from the deposition age of the sediments, as is typical for convergent margin settings. The zircon population also shows a record of Archaean and of Palaeoproterozoic crust hosting parts of the Stenian volcanic arc, as well as a significant portion of Archaean and Proterozoic grains that reflect well-known tectono-magmatic events in the KGC.

The Jutulrøra metavolcanic gneisses in the eastern H.U. Sverdrupfjella are likely the metamorphosed remains of the Stenian continental arc formed in the run-up to Rodinia assembly, while the Ritscherflya sedimentary rocks were deposited further inland and escaped intense deformation and metamorphism (Fig. 12). The pre-Rodinian margin of the KGC and the Rodinia suture consequently cannot be hidden under the Pencksøkket-Jutulstrumen glaciers, but must be located further to the east in the metamorphic Maud Belt, possibly between the western and eastern H.U. Sverdrupfjella (Fig. 12).

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Fig. 1 Simplified map of Dronning Maud Land (DML) (modified after Board *et al.*, 2005). The major geologic units are the high-grade metamorphic Maud Belt (Meso- and Neoproterozoic), the Mesoproterozoic sedimentary rocks and sills in the Ritscherflya Supergroup (including Ahlmannryggen and Borgmassivet), and the Archaean basement of the Grunehogna craton (exclusively exposed at Annandagstoppane). The location of the boundary between craton and Maud Belt is discussed in the text. The inset shows the Antarctic continent with the location of DML at the edge of the Weddell sea.

Fig. 2 Satellite image of the southern Ahlmannryggen with sample stations marked by numbers and locality names. Rock outcrop is visible as dark spots; everything else is occupied by snow-covered ice and glaciers (Viddalen, Frostlendet, Jutulstraumen). Latitudes, longitudes and distances to the other nunataks are given for orientation. BAS 2007/08 camp sites are marked with blue squares. Google Earth image based on USGS image.

Fig. 3 Field photographs of the studied Ritscherflya sedimentary rocks in the Ahlmannryggen. **(a)** View of the Borgmassivet and Ahlmannryggen from Straumsvola (western H.U. Sverdrupfjella, Maud belt) in the east across the Jutulstraumen glacier. **(b)** View of the Pyramiden nunatak from the north-west (locality Z7-28). Hight of outcrop ~ 150 m. **(c)** View of the Grunehogna nunatak, Peak 1390 (locality Z7-37) exposing reddish sandstones of the Grunehogna Formation intruded by mafic sills. Hight of outcrop ~ 350 m. **(d)** Wave ripples in Pyramiden Formation sandstone at Pyramiden (locality Z7-28). **(e)** Conglomerate interlayered with cross-bedded sandstone of the Grunehogna Formation. Samples Z7-36-2 and -3 were taken from this rock. **(f)** Conglomerate, sandstone and greywacke with fragments of mudstone of the Grunehogna Formation at Viddalskollen (locality Z7-39). The yellow colouration is caused by metamorphic chlorite and epidote. **(g)** Volcaniclastic deposit (agglomerate) of the Tyndeklypa Formation at the Istind nunatak (locality Z7-38). Abundant blocks of sandstone and other sedimentary rocks, as well as less common fragments of older volcanic and sub-volcanic rocks are embedded in a fine-grained matrix that contains devitrified flattened pumice and fiamme (3 kg hammer with 75 cm handle for scale in e, f and g).

Fig. 4 Schematic profile through the Ahlmannryggen sedimentary rock sequence of the Ritscherflya Supergroup, possibly deposited on top of the Archean Grunehogna craton (adopted from Wolmarans & Krynauw, 1981; Wolmarans & Kent, 1982; Groenewald *et al.*, 1995; Perritt, 2001). Note, that basement granite is only exposed at Annandagstoppane ~ 80 km west of the westernmost sedimentary rock outcrop (see Fig. 1). Not shown are the large mafic sills (*Borgmassivet intrusives*) that dominate most of the outcrops today. Sample numbers and localities are given on the right hand side. The nomenclature and allocation of samples to formations follows Perritt (2001). *after Wolmarans & Kent (1982), the Viddalskollen nunataks expose rocks of the Pyramiden Formation, the Grunehogna Formation is the Grunehogna Member and as such part of the Høgfonna Formation, and the Straumsnutane Group was the Straumsnutane Formation and as such part of the Jutulstraumen Group. However, here we follow the conclusions and nomenclature of Perritt (2001).

Fig. 5 Concordia diagrams of zircon from all five sampled formations of the Ahlmannryggen. **(a)** The 495 analyses that passed the quality test ($< \pm 10$ % discordance, < 15 % disturbance of the $^{208}Pb/^{206}Pb$ system, < 5 % common-Pb correction). **(b)** All analyses (concordant and discordant) in the Stenian age group (< 1200 Ma) that required a common-Pb corrections of < 5 %. Note the broad discordant array reaching from the age of the dominant zircon population (~ 1130 Ma) to the age of Pan-African metamorphism in DML, and probably minor components of Jurassic or recent Pb loss.

Fig. 6 Histograms and probability-density plot for zircon $^{207}Pb/^{206}Pb$ ages in all investigated sedimentary rock samples from Ritscherflya Supergroup. Only zircon analyses that are less than 10 % discordant are considered ($n = 473$). The vertical blue bar marks the sedimentation age of $1130 - 1107$ Ma of the supergroup. The wide yellow bar marks the age of magmatism and metamorphism in the Maud belt (e.g., Jacobs *et al.*, 2003*a*; Board *et al.*, 2005). The narrow yellow bars mark the age of crystallisation and of inherited zircon grains of the Annandagstoppane (ADT) granite (Marschall *et al.*, 2010). Note the break in scale to accommodate the large peak of Stenian grains. Important southern African orogenic events and provinces are marked for comparison with the peaks in the detrital record (see Hanson, 2003; Dorland *et al.*, 2006; Gerdes & Zeh, 2009, and references therein).

Fig. 7 Pie chart displaying the relative proportions of zircon age populations in the investigated samples. Only zircon analyses that are less than 10 % discordant are considered ($n = 471$). Approximately two thirds of the grains (or rims of grains) crystallised close to the sedimentation age of the Ritscherflya Supergroup and are labeled "Stenian" (see dominant age peak in Fig. 6). Proterozoic grains (or cores) older than 1200 Ma comprise approximately one quarter of the population, while 6.8 % of the grains (or cores of grains) are Archaean. *1.5 % of the grains show slightly discordant ages around 1200 Ma that make their classification ambiguous.

Fig. 8 Cathodo-luminescence images of detrital zircon grains form the Pyramiden and Grunehogna Formations with inherited Palaeoproterozoic and Archaean cores and Stenian, oscillatory zoned rims. All ages shown are $^{207}Pb/^{206}Pb$ ages with 1σ precision. Values in parentheses are $10 - 15$ % discordant, while all other ages are < 10 % discordant. Zircon grains with Archaean cores and Stenian rims indistinguishable from the sedimentation age of the sedimentary rocks are strong evidence for a sediment source with volcanism in an active continental arc located on the edge of an Archaean craton.

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Fig. 9 Histograms and probability-density plots for zircon $^{207}Pb/^{206}Pb$ ages in sedimentary rock samples from the five investigated formations of the Ritscherflya Supergroup. Only zircon analyses that are less than 10 % discordant are considered. The vertical blue bar marks the sedimentation age of $1130 - 1104$ Ma of the supergroup.

Fig. 10 Plot showing the cumulative proportions of the time difference between zircon crystallisation ages and the depositional age of the Ritscherflya sedimentary rocks (after Cawood *et al.*, 2012). The coloured lines display the five different formations. The thick black line shows the complete set of samples. All lines are clearly within the field of sediments deposited at convergent margins as identified by Cawood *et al.* (2012). For the fields of collisional and extensional settings only the most discriminative lower parts ≤ 40 % are displayed.

Fig. 11 Reconstruction of the Kalahari-Grunehogna Craton (KGC) in the late Mesoproterozoic between $\sim$ 1200 and 1100 Ma (modified from Jacobs *et al.*, 2008*b*). Outward subduction followed by accretion of Proterozoic arcs or microcontinents characterised the southern margin of the KGC. Its eastern margin, in contrast, was the site of a continental arc and deposition of the Ritscherflya Supergroup which contains the detrital record of that arc. Large parts of the continental arc were overprinted by high-grade metamorphism and form now part of the metamorphic Maud Belt.

Fig. 12 Schematic east-west cross-section through western Dronning Maud Land. The Grunehogna Craton, exposed at Annandagstoppane shows Palaeoarchaean zircons and Eoarchaean model ages (Marschall *et al.*, 2010). In contrast, eastern H.U. Sverdrupfjella and central DML show Palaeoproterozoic model ages and detrital zircon ages (Board *et al.*, 2005; Grosch *et al.*, 2007). The western H.U. Sverdrupfjella likely represents the margin of the KGC overprinted by high-grade metamorphism (Wareham *et al.*, 1998; Grosch *et al.*, 2007) and exposes subduction-related metavolcanic gneisses that represent remnants of the Mesoproterozoic active volcanic margin of the KGC. The exact location of the suture between the KGC and the younger crustal blocks of central DML is not known, but it may be represented by the 'L5' or Forster magnetic anomaly, which runs in a NE direction between eastern and western H.U. Sverdrupfjella (Ferraccioli *et al.*, 2005; Riedel *et al.*, 2013).

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Table 1 Investigated sedimentary rock samples from the Ritscherflya Supergroup

Formation	Sample	Locality	Latitude S	Longitude W	Elevation (a.m.s.l.)	Rock type	Detrital components	Metamorphic minerals
Pyramiden	Z7-28-1	Pyramiden	72°16.826′	003°47.910′	1400 m	greywacke	Qtz, Pl†	Chl, Prh
	Z7-28-2	Pyramiden	72°16.826′	003°47.910′	1400 m	sandstone	Qtz, Pl†	Ep, Chl
Schumacher-fjellet	Z7-36-6	Grunehogna Peak 1285	72°01.759′	002°48.401′	974 m	granosediment**	Qtz	Pl, Chl, Ep
	Z7-36-8	Grunehogna Peak 1285	72°01.759′	002°48.401′	974 m	siltstone	Qtz	WM, Prh
	Z7-37-6	Grunehogna Peak 1390	72°02.898′	002°46.394′	1263 m	sandstone	Qtz	Chl, Cal, Ep, Ttn
Grunehogna*	Z7-36-2	Grunehogna Peak 1285	72°02.222′	002°47.924′	1172 m	conglomerate	Qtz, chert, quartzite	Prh, Ep, Chl
	Z7-36-3	Grunehogna Peak 1285	72°02.222′	002°47.924′	1172 m	sandstone	Qtz, Pl†, Rt	Prh, WM, Chl
	Z7-39-1	Viddalskollen*	72°25.424′	002°18.533′	1389 m	conglomerate	Qtz, mudstone, chert	Ep, Chl
	Z7-39-2	Viddalskollen*	72°25.424′	002°18.533′	1395 m	conglomerate	Qtz, Pl, chert	Ep, Chl
	Z7-39-3	Viddalskollen*	72°25.440′	002°18.905′	1321 m	greywacke	Qtz	Ep
Tyndeklypa	Z7-38-1	Istind Peak 1320	72°08.531′	002°15.683′	1017 m	agglomerate	sandstone, pumice, basalt Qtz, Pl, Kfs, WM	Chl, Cal, Pmp
Istind	Z7-38-3	Istind Peak 1320	72°08.441′	002°16.260′	1079 m	sandstone	Qtz, WM	Prh, Cal

*after Wolmarans & Kent (1982), the Viddalskollen nunataks expose rocks of the Pyramiden Formation, the Grunehogna Formation is the Grunehogna Member and as such part of the Høgfonna Formation. However, here we follow the conclusions and nomenclature of Perritt (2001). **granosediment is a sediment that was partially melted by contact metamorphism due to the intrusion of a mafic sill (Krynauw *et al.*, 1988; Curtis & Riley, 2003). Mineral abbreviations are: Qtz = quartz; Pl = plagioclase; Rt = rutile; Kfs = K-feldspar; Chl = chlorite; Ep = epidote; WM = white mica; Cal = calcite; Ttn = titanite; Prh = prehnite; Pmp = pumpellyite. †altered.

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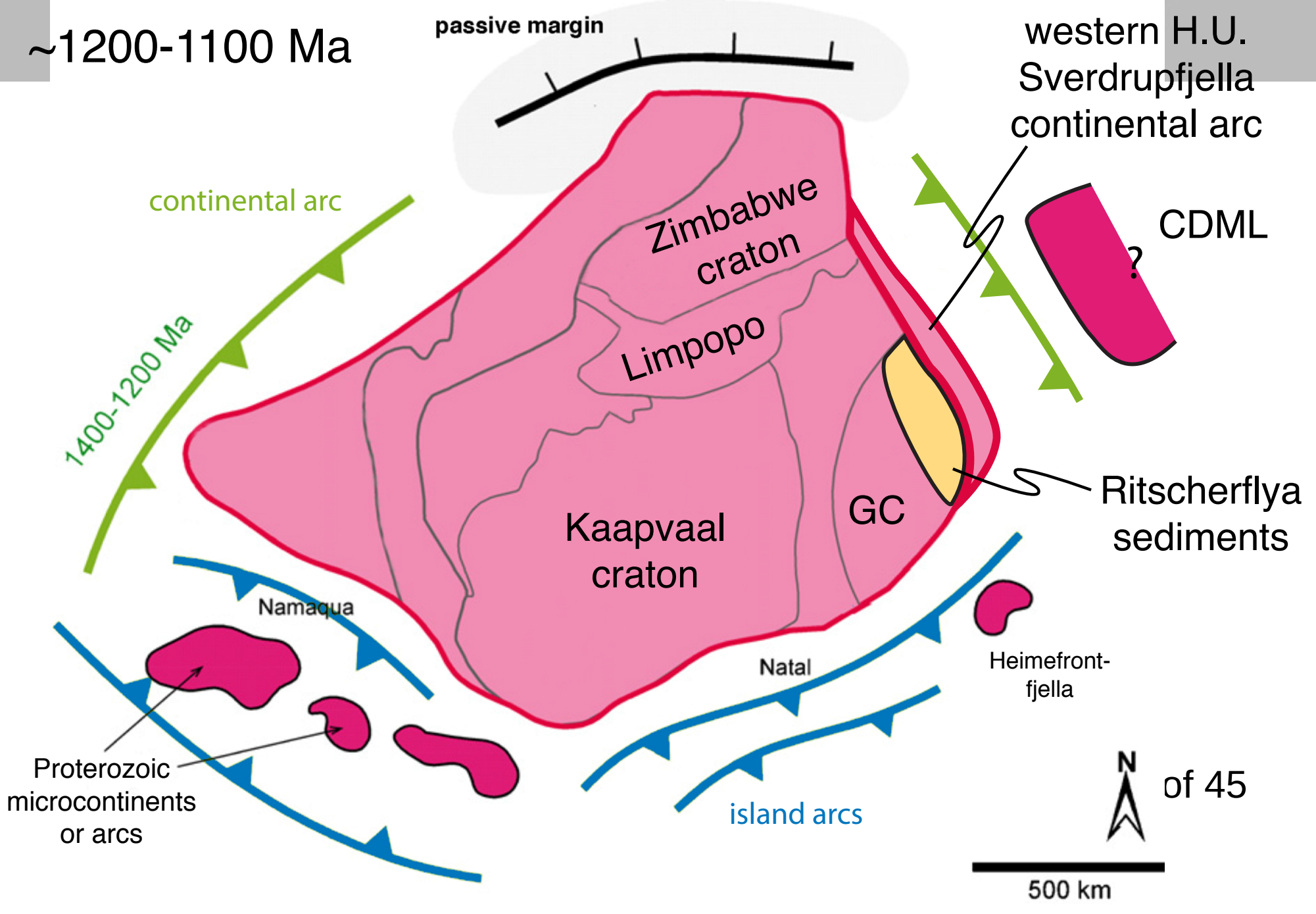
~1200-1100 Ma
passive margin
western H.U.
Sverdrupfjella
continental arc
continental arc
1400-1200 Ma
Zimbabwe
craton
Limpopo
CDML
?
Kaapvaal
craton
GC
Ritscherflya
sediments
Namaqua
Natal
Heimefront-
fjella
Proterozoic
microcontinents
or arcs
island arcs
N
500 km

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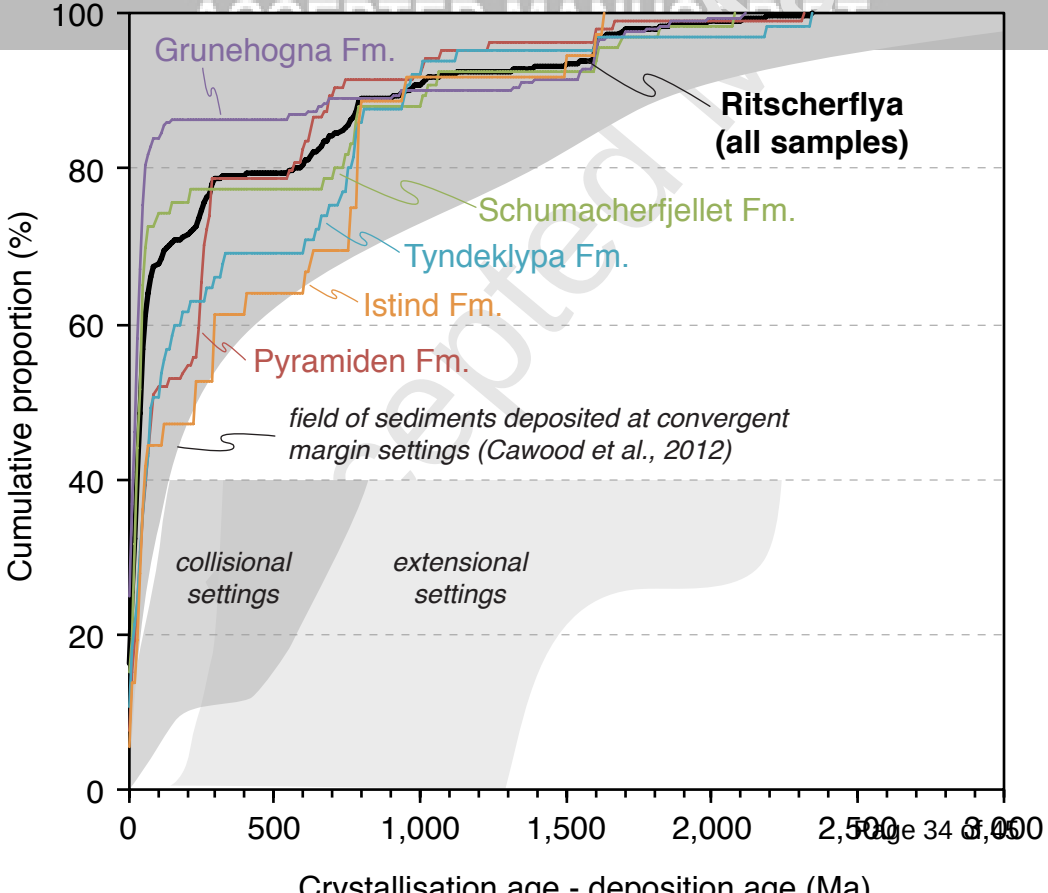

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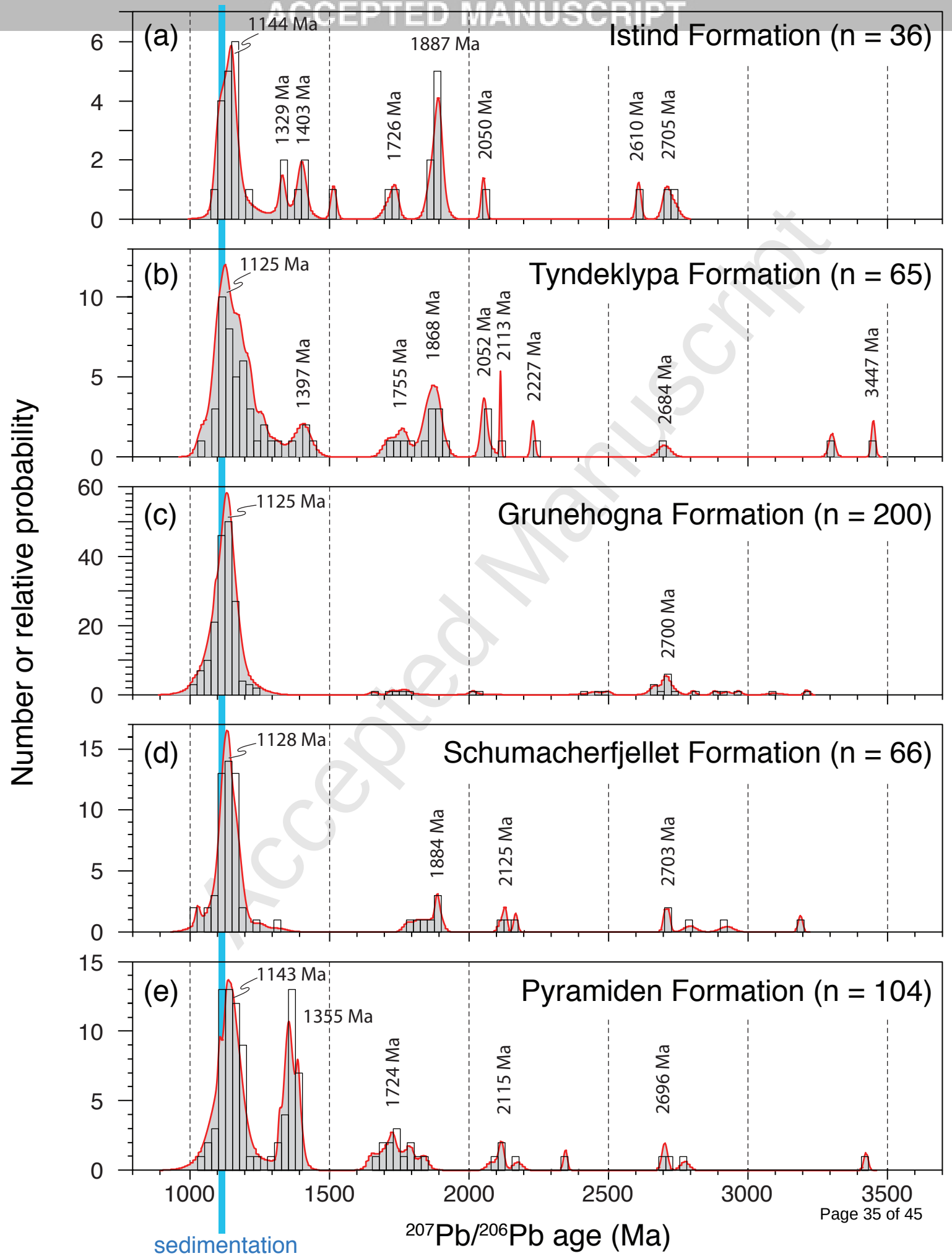

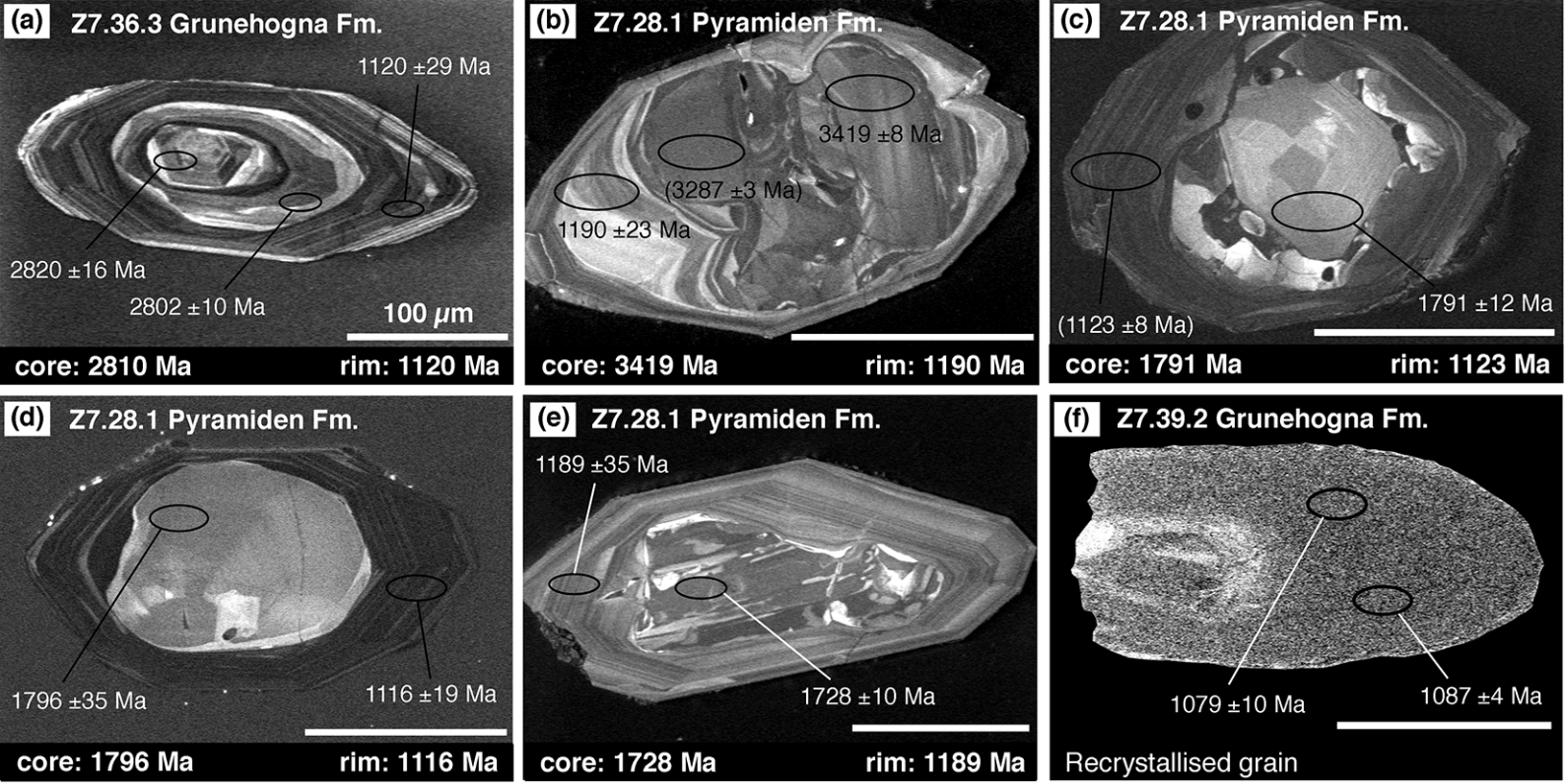
(a) Z7.36.3 Grunehogna Fm.
1120 ±29 Ma
2820 ±16 Ma
2802 ±10 Ma
100 μm
core: 2810 Ma
rim: 1120 Ma
(b) Z7.28.1 Pyramiden Fm.
3419 ±8 Ma
(3287 ±3 Ma)
1190 ±23 Ma
core: 3419 Ma
rim: 1190 Ma
(c) Z7.28.1 Pyramiden Fm.
(1123 ±8 Ma)
1791 ±12 Ma
core: 1791 Ma
rim: 1123 Ma
(d) Z7.28.1 Pyramiden Fm.
1796 ±35 Ma
1116 ±19 Ma
core: 1796 Ma
rim: 1116 Ma
(e) Z7.28.1 Pyramiden Fm.
1189 ±35 Ma
1728 ±10 Ma
core: 1728 Ma
rim: 1189 Ma
(f) Z7.39.2 Grunehogna Fm.
1079 ±10 Ma
1087 ±4 Ma
Recrystallised grain

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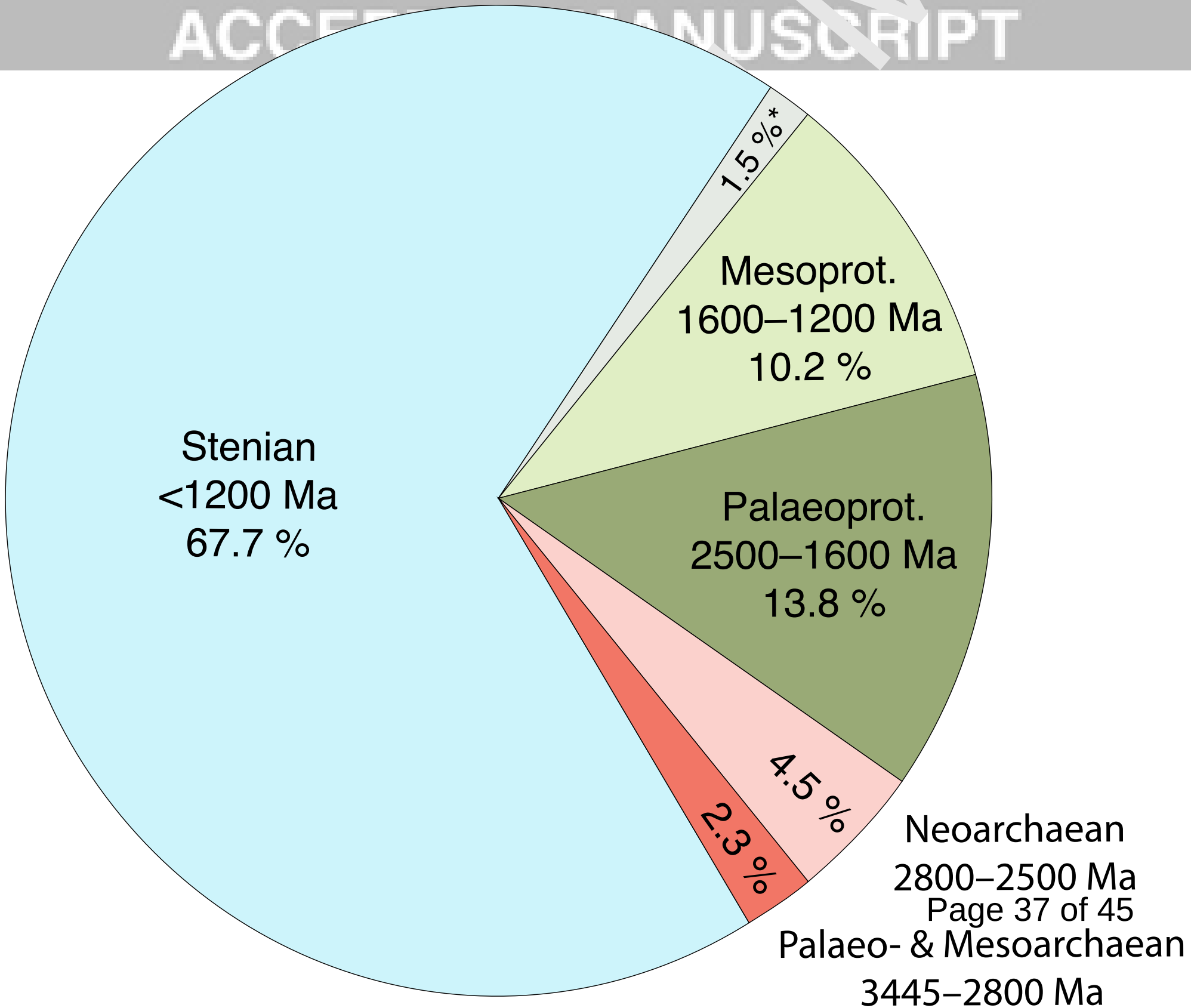

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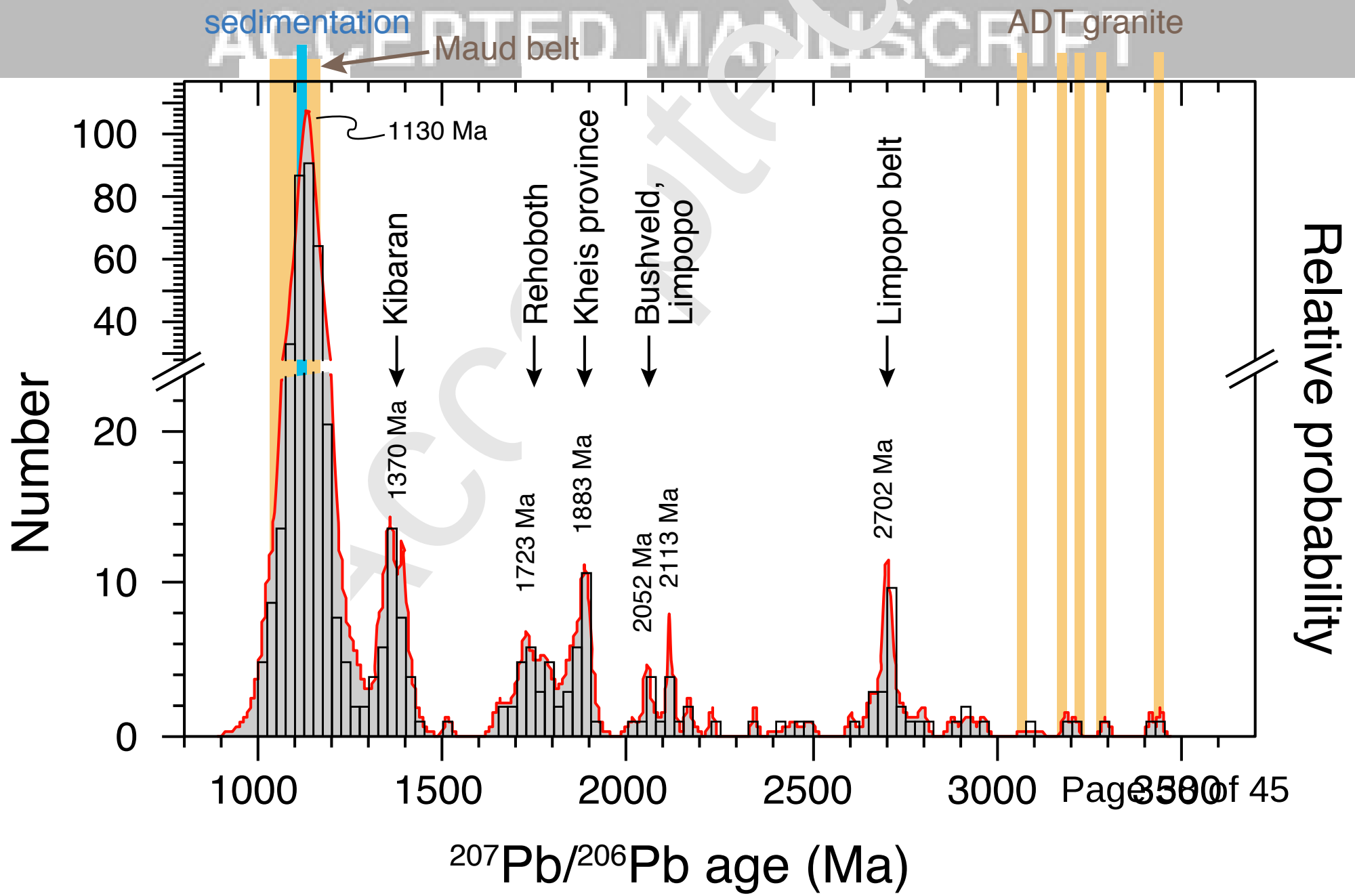

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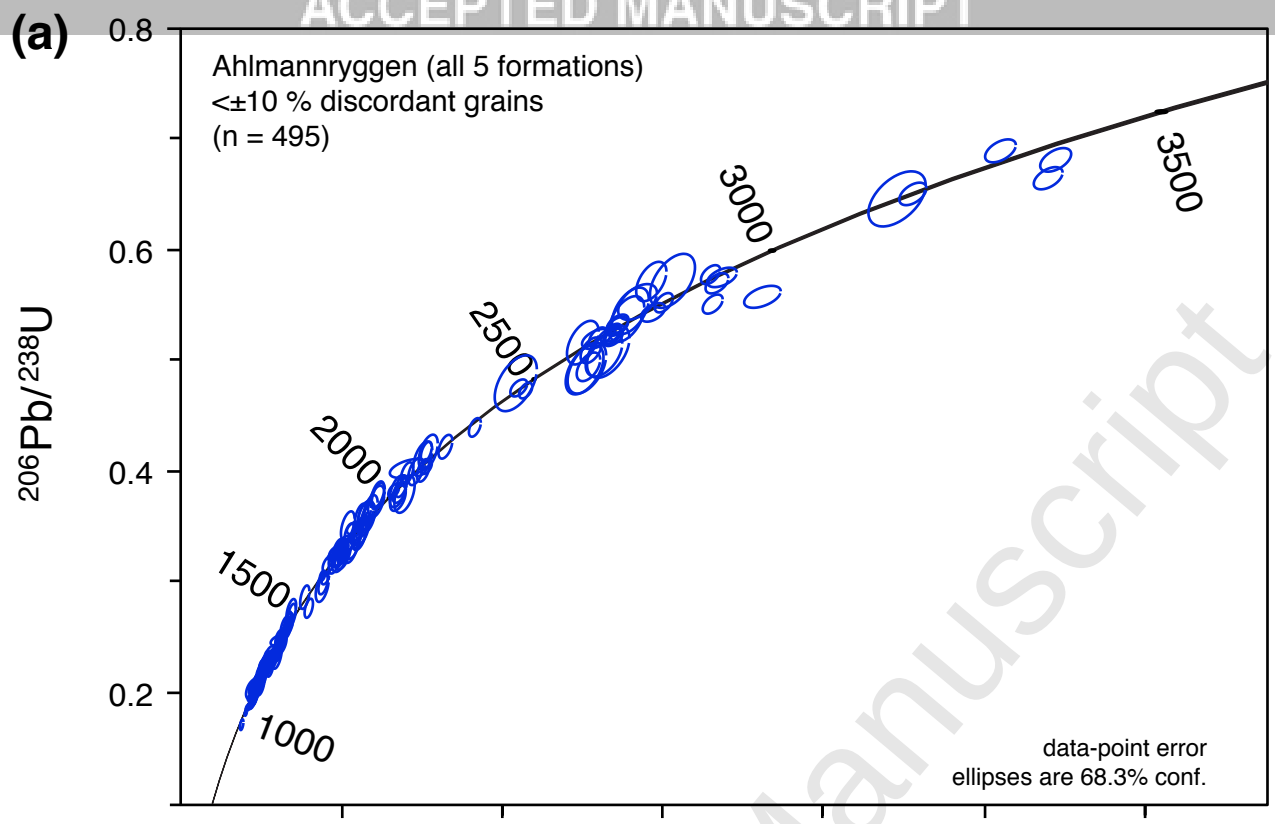

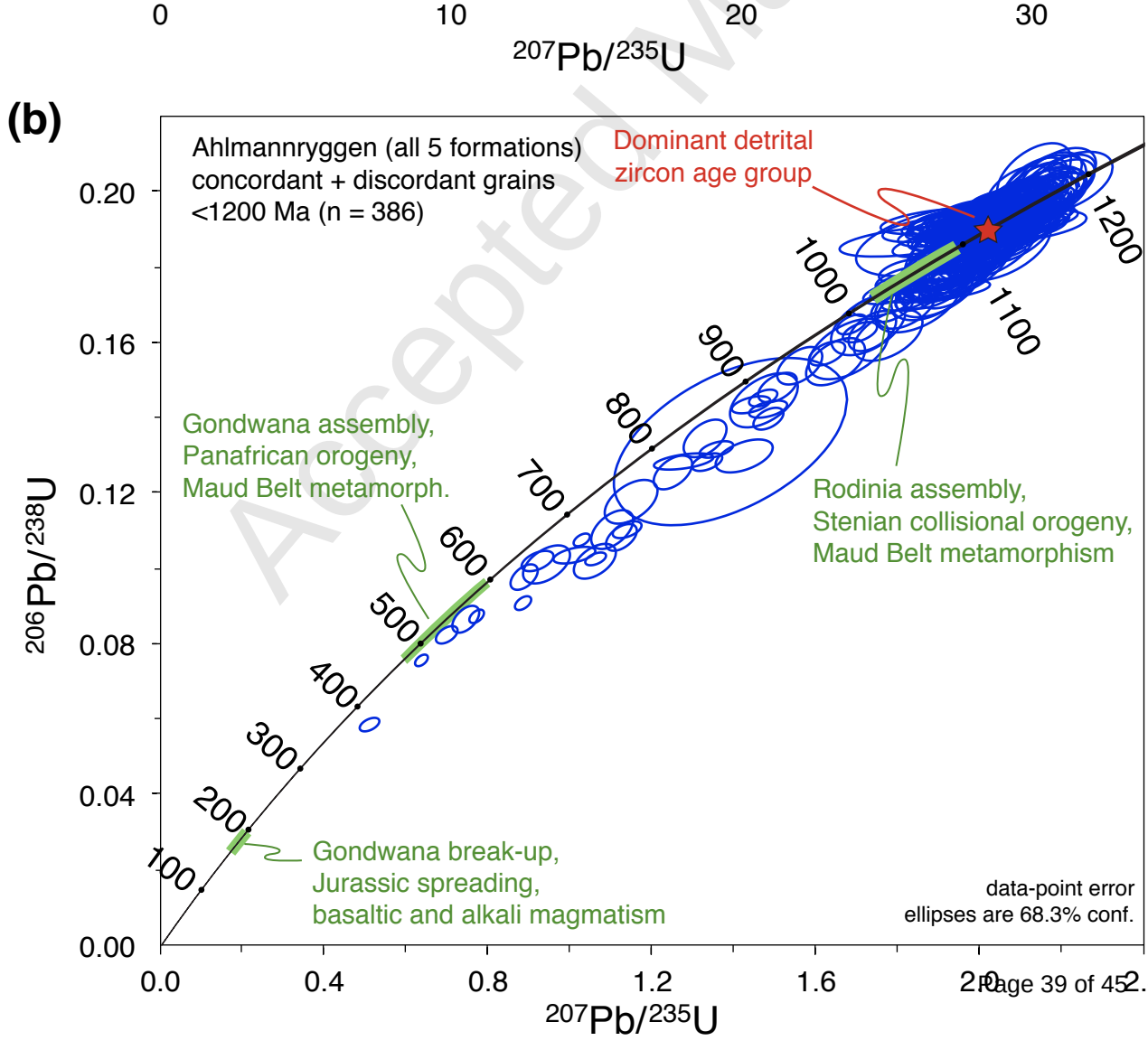

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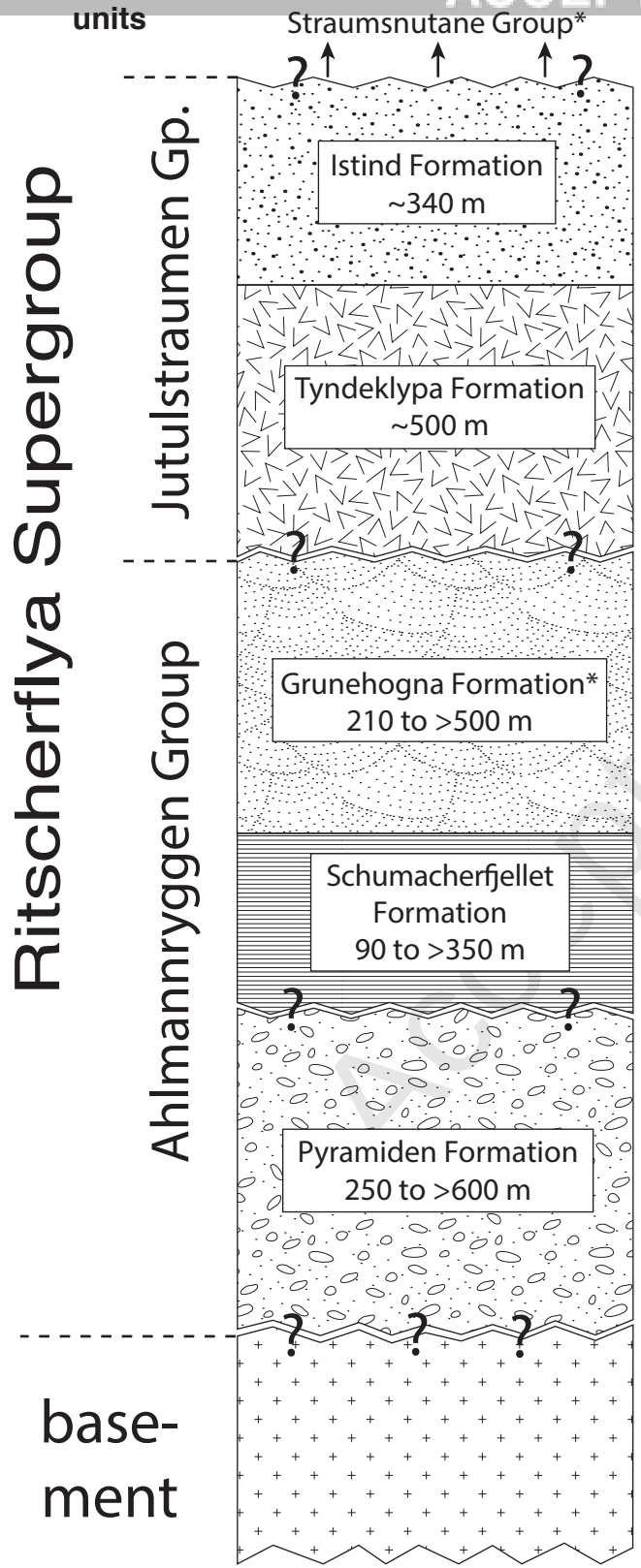

Dominant rock types	Investigated samples	Sample localities
quartzite interbedded with agglomerates, tuff and lava flows, sills	Z7.38.3	Istind
agglomerates (with blocks from all units of Ahlmannryggen fm. + blocks of plutonic rocks), tuff, tuffite, arenite	Z7.38.1	Istind
contact not exposed		
coarse-grained sandstone, quartzite, shale conglomerates (pebbles of red jasper, quartzite, shale)	Z7.39.1 Z7.39.2 Z7.39.3	Viddalskollen*
	Z7.36.2 Z7.36.3	Grunehogna
light-coloured ortho-quartzite interbedded with dark-coloured mudstone	Z7.36.6 Z7.36.8	Grunehogna
	Z7.37.6	Grunehogna
contact not exposed		
feldspathic greywacke, siltstone, mudstone, conglomerates (mud pebbles)	Z7.28.1 Z7.28.2	Pyramiden
contact not exposed		
basement only exposed in granite outcrops of Annandagstoppane; 3067 ±8 Ma (Marschall *et al.*, 2010)		

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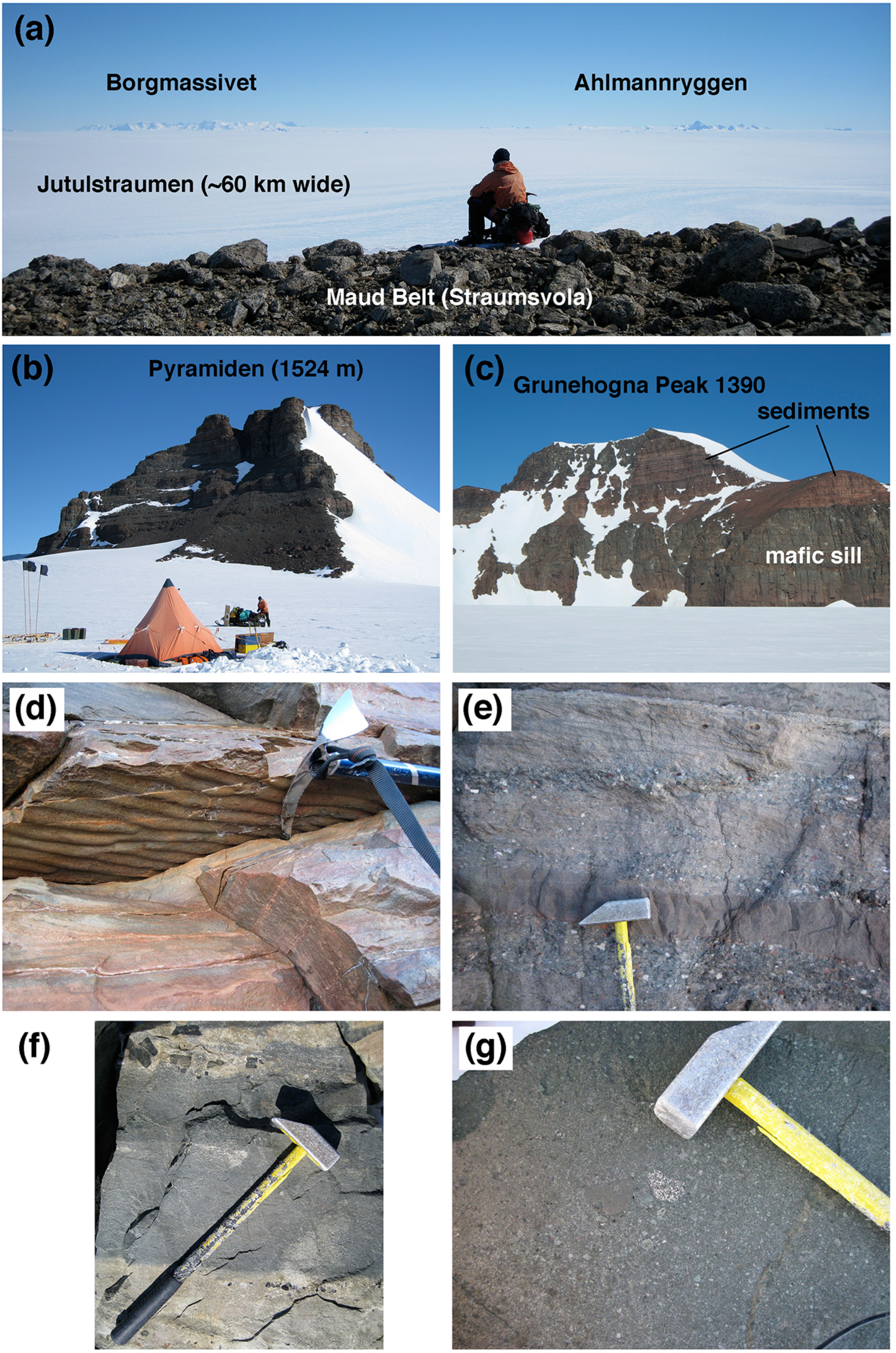
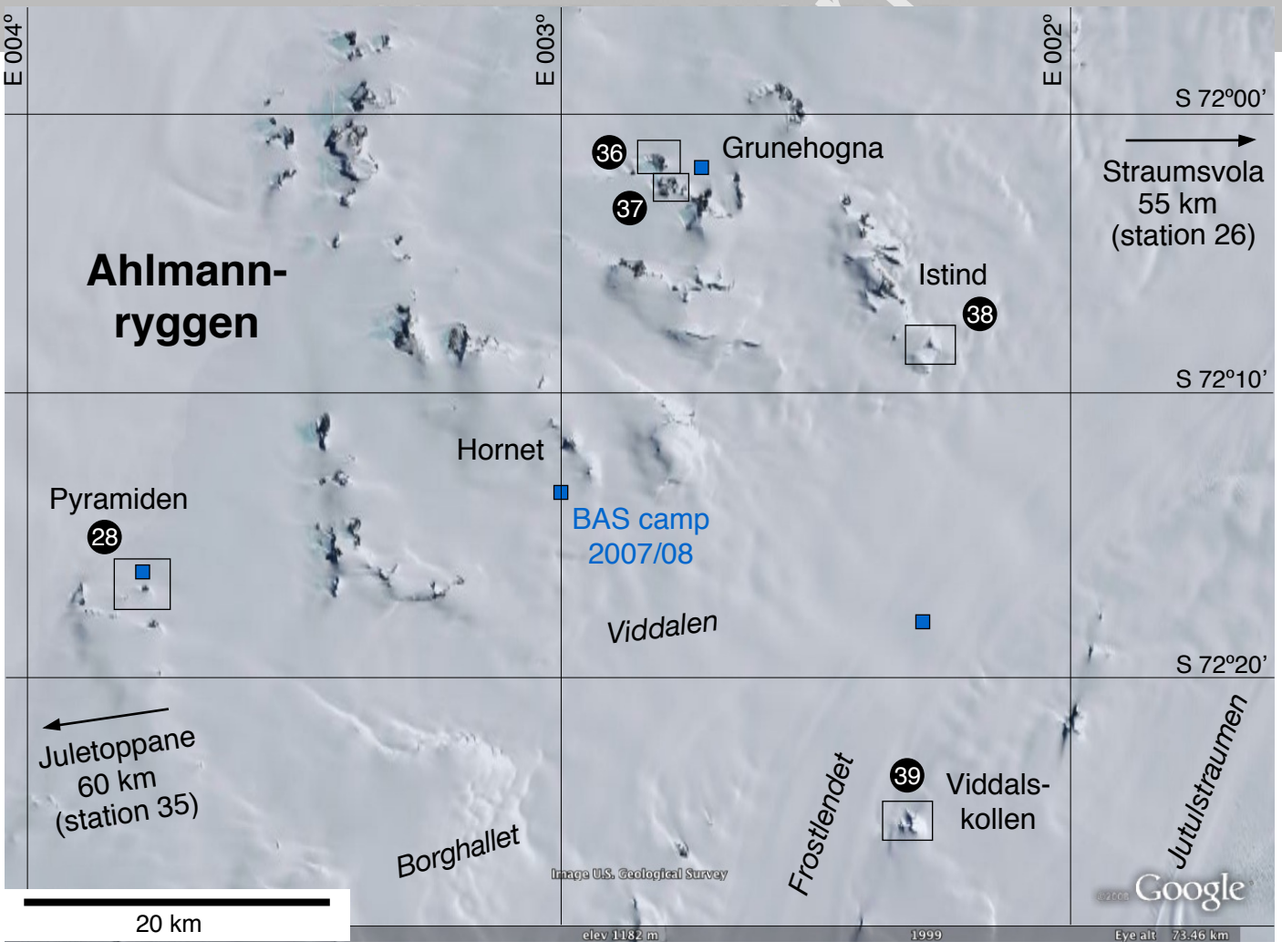
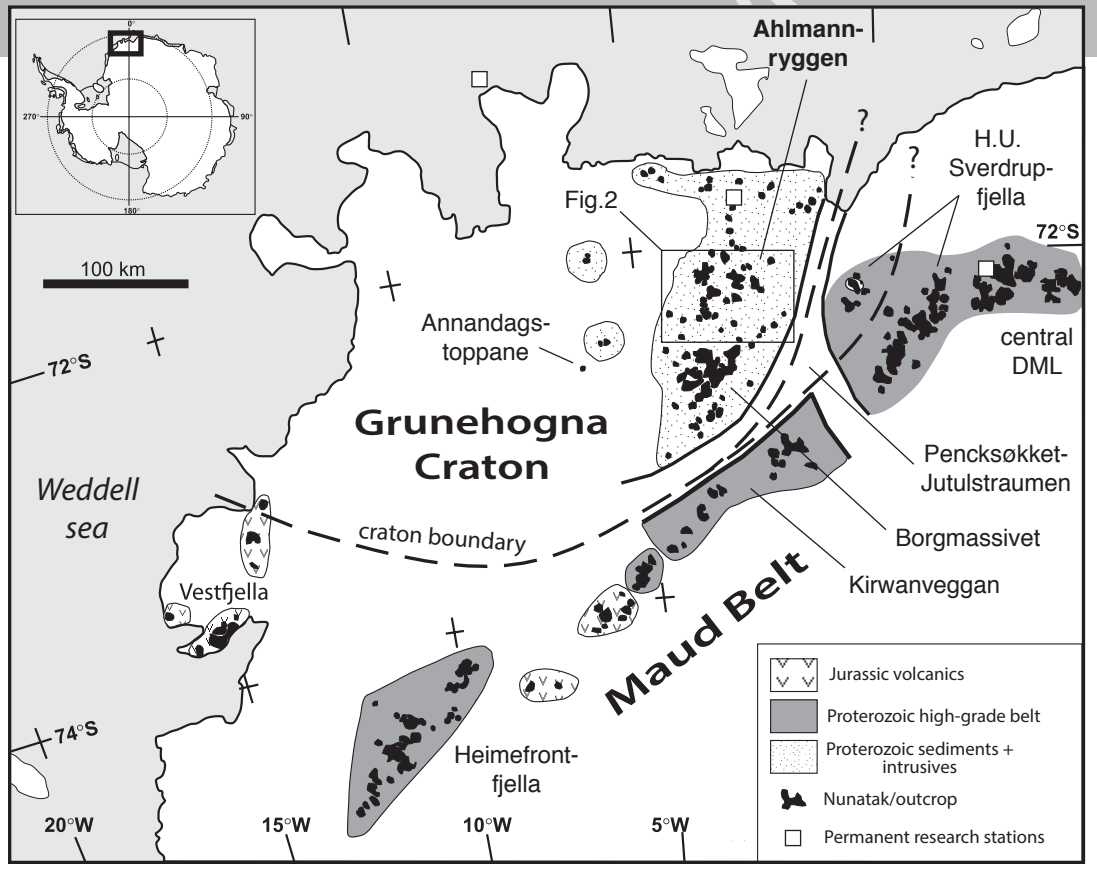

(a)
Borgmassivet
Ahlmannryggen
Jutulstraumen (~60 km wide)
Maud Belt (Straumsvola)
(b)
Pyramiden (1524 m)
(c)
Grunehogna Peak 1390
sediments
mafic sill
(d)
(e)
(f)
(g)


E 004°
E 003°
E 002°
S 72°00'
S 72°10'
S 72°20'
Ahlmann-
ryggen
36
Grunehogna
37
Straumsvola
55 km
(station 26)
Istind
38
Hornet
Pyramiden
28
BAS camp
2007/08
Viddalen
Juletoppane
60 km
(station 35)
Borghallet
Frostlendet
39
Viddals-
kollen
Jutulstraumen
Image U.S. Geological Survey
Google
20 km
elev 1182 m
1999
Eye alt 73.46 km


Ahlmann-
ryggen
?
?
H.U.
Sverdrup-
fjella
Fig.2
72°S
100 km
Annandags-
toppane
central
DML
72°S
Grunehogna
Craton
Weddell
sea
craton boundary
Pencksøkket-
Jutulstraumen
Borgmassivet
Kirwanveggan
Vestfjella
Maud Belt
Heimefront-
fjella
74°S
20°W
15°W
10°W
5°W
Jurassic volcanics
Proterozoic high-grade belt
Proterozoic sediments +
intrusives
Nunatak/outcrop
Permanent research stations

ACCEPTED MANUSCRIPT

WEST

EAST

GRUNEHOGNA CRATON

METAMORPHOSED MARGIN OF GC

HIGH-GRADE MAUD BELT

Weddell Sea + shelf

Annandagstoppane

Ahlmannryggen

Jutulstraumen

H.U. Sverdrupfjella

western

eastern

Central Dronning Maud Land

S-type granite
3068 Ma

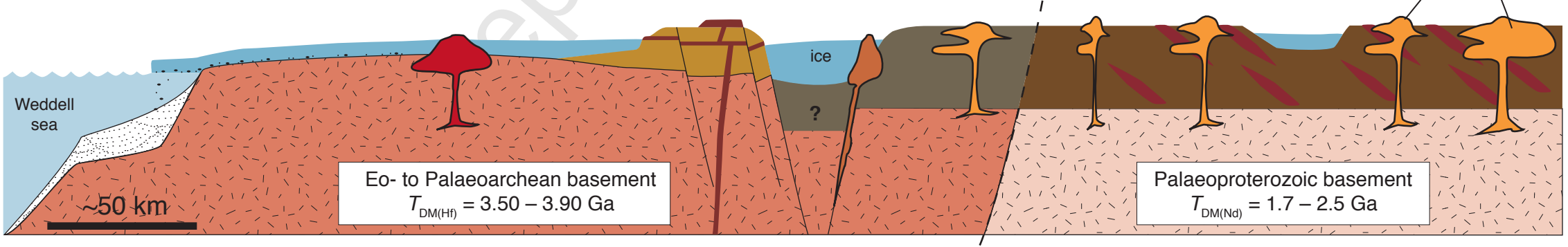
Ritscherflya supergroup sediments + Borg mafic sills, 1130 – 1107 Ma

Calc-alkaline metavolcanics ~1130 Ma volcanism

paragneiss + metagranite (from ~1070 Ma intrusion)

Panafrican post-orogenic A-type intrusions (~500 Ma)

oldest zircon: *3433 Ma (inherited)* *3445 Ma (detrital)* *1765 Ma (detrital)*



STENIAN METAMORPHISM

(sub-)greenschist facies Mesoproterozoic metamorphism

Stenian suture? (~1080 Ma?)

amphibolite- to granulite-facies Mesoproterozoic metamorphism (~1080 – 1030 Ma)

EDIACARAN-CAMBRIAN METAMORPHISM

Neoproterozoic hydrothermal activity, deformation low-grade metamorphism(?)

? ?

amphibolite- to granulite-facies Neoproterozoic metamorphism (~580 – 540 Ma)

ACCEPTED MANUSCRIPT

- U-Pb dates of detrital zircon form Mesoproterozoic Ritscherflya sediment
- dominant age peak indistinguishable from deposition age
- significant Archaean popoulation as old as 3445 Ma
- zircon with Mesoarchaean cores and Stenian rims evidence for continental arc
- low-grade metamorphism during Rodinia and Gondwana formation

Accepted Manuscript