Anorthosites in the Eastern Granulites of Tanzania—New SIMS zircon U–Pb age data, petrography and geochemistry

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Abstract

Several occurrences of anorthosites are known in the Neoproterozoic Mozambique Belt in Tanzania. These are tectonically incorporated into a suite of enderbitic rocks and migmatitic orthogneisses of the Eastern Granulites. Two larger anorthosite bodies and associated rocks from the Pare Mountains and the Uluguru Mountains have been chosen for a comparative study regarding their formation age, age of subsequent metamorphic processes, mineral chemistry and rock chemistry. All the investigated magmatic bodies were overprinted by Neoproterozoic high-pressure granulite facies metamorphism and deformation, documented by similar petrography and mineralogy, metamorphic textures and deformational characteristics. However, mineral chemistry, geochemistry and geochronology reveal major differences between the two occurrences. The Pare anorthosite contains igneous zircons that yield Archean formation ages of ca. 2.64 Ga and a more calcic mineral and rock chemistry that is typical for Archean-type anorthosites. Extremely narrow metamorphic rims around zircons could not be dated but may have grown during the Neoproterozoic metaporphic overprint. In contrast, the Uluguru anorthosite contains igneous zircon cores with U/Pb ages of 880–820 Ma. A broad metamorphic rim was dated at ca. 640–650 Ma. The adjacent migmatitic basement gneisses yield formation ages of ca. 986 Ma followed by Pb-loss. Early Neoproterozoic events can also be inferred from both anorthosite suite and basement samples of the Mahenge Mountains. There U/Pb ages from zircon cores cluster around 730–800 Ma. Well-developed metamorphic rims yield Neoproterozoic ages of 650 Ma. The basement gneisses are related to a destructive plate margin setting. The anorthosites may have formed when parts of Central Madagascar rifted off East Africa at the time considered.

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1. Introduction

Anorthosites are magmatic rocks that consist of at least 90% plagioclase and represent an important rock type in many Precambrian terranes. Anorthosites occur in both Archean and Proterozoic with the different eons having distinctive anorthosite characteristics. About 70 occurrences of Archean anorthosites are known worldwide that are sill-like bodies characterized by calcic plagioclase megacrysts. More voluminous Proterozoic anorthosite massifs (therefore massif-type) are known from ca. 130 locations and exhibit different mineralogy.
Table 1
Overview of age data from recent literature for East African and some Gondwanan anorthosites

<table>
<thead>
<tr>
<th>Age domains</th>
<th>Reference</th>
<th>Age</th>
<th>M/X</th>
<th>Method</th>
</tr>
</thead>
<tbody>
<tr>
<td>Archean Ages</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sittampundi Complex (South India)</td>
<td>Bashkar et al. (1996)</td>
<td>2935 ± 60 Ma</td>
<td>X</td>
<td>Sm–NdWR</td>
</tr>
<tr>
<td>Bhuvanes Complex (South India)</td>
<td>Bashkar et al. (1996)</td>
<td>2899 ± 28 Ma</td>
<td>X</td>
<td>Sm–NdWR</td>
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<tr>
<td>Proterozoic Ages</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Tete Complex—Mozambique</td>
<td>Evans et al. (1999)</td>
<td>1025 ± 79 Ma</td>
<td>X</td>
<td>Sm–NdWR</td>
</tr>
<tr>
<td>Anorthosites—Madagascar</td>
<td>Ashwal et al. (1998)</td>
<td>600 ± 60 Ma</td>
<td>X</td>
<td>Sm–Nd</td>
</tr>
<tr>
<td>Anorthosites—Madagascar</td>
<td>Ashwal et al. (1998)</td>
<td>ca. 950 Ma</td>
<td>X</td>
<td>TDM model age</td>
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<tr>
<td>Antarctica</td>
<td>Jacobs et al. (1998)</td>
<td>ca. 680 Ma</td>
<td>X</td>
<td>U/Pb Shrimp</td>
</tr>
<tr>
<td>Uluguru Anorthosite</td>
<td>Maboko (1995)</td>
<td>1.36–1.65 Ga</td>
<td>X</td>
<td>TDM model age</td>
</tr>
<tr>
<td>Uluguru Anorthosite</td>
<td>Maboko and Lenoir (1994)</td>
<td>695 ± 4 Ma</td>
<td>X</td>
<td>U/Pb</td>
</tr>
<tr>
<td>Uluguru Anorthosite</td>
<td>Maboko and Nakamura (1995)</td>
<td>633 ± 7 Ma</td>
<td>M</td>
<td>Sm–Nd (Get)</td>
</tr>
<tr>
<td>Kanene Complex (Namibia)</td>
<td>Mayer et al. (2004)</td>
<td>1371 ± 2.5 Ma</td>
<td>X</td>
<td>U/Pb</td>
</tr>
<tr>
<td>Kanene Complex (Namibia)</td>
<td>Mayer et al. (2004)</td>
<td>1470 ± 25 Ma</td>
<td>X</td>
<td>Sm–Nd</td>
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<tr>
<td>Kanene Complex (Namibia)</td>
<td>Mayer et al. (2004)</td>
<td>1319 ± 28 Ma</td>
<td>X</td>
<td>Sm–Nd</td>
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</table>

Abbreviations in the row M/X are used for the subdivision of metamorphic (‘M’) vs. crystallization (‘X’) ages as interpreted in literature.

The characteristics of all known anorthosites have been nicely compiled by Ashwal (1993). From these abundant northern hemisphere anorthosites, which are termed ‘anorthosite’, are overprinted by metamorphism to various degrees. During the East African orogeny, the Eastern Granulites were thrust onto the basement of the Tanzania Craton and Usagaran Belt (Fritz et al., 2005). This led to a complex distribution of Archean and Proterozoic ages in the Mozambique Belt and opened the debate about how much juvenile crust exists in this orogen. Möller et al. (1998) found that only the Eastern Granulites contain juvenile Neoproterozoic crust and interpreted a fundamental crustal age domain boundary in the Uluguru Mountains. However, the database of in situ U/Pb crystallization ages of zircons in the Eastern Granulites is still scarce, as shown below. Amongst the anorthosites in Tanzania, only those of the Uluguru Mountains have been studied geochronologically in some detail (Table 1). Apart from descriptive studies (Sampson and Wright, 1964), few isotope data are available. Muhongo and Lenoir (1994) dated a zircon fraction that gave a U/Pb (concordia upper intercept) age of 695 ± 4 Ma, interpreted as the crystallization age of the body. A Neoproterozoic metamorphic age (633 ± 7 Ma) from a garnet fraction in the anorthosite was published by Maboko and Nakamura (1995) and TDM model ages from the Uluguru anorthosites range between 1.36 and 1.64 Ga (Maboko, 1995).

In this paper we present for the first time crystallization ages of the anorthosites from three mountain ranges that belong of the Eastern Granulites (Pare, Uluguru and Mahenge Mountains). We used the ion microprobe for the determination of U/Pb-isotopic ratios of individual zircon grains of anorthosites and associated rocks.
Fig. 1. Geological overview of the Mozambique belt and adjacent units in Tanzania. Sample localities plus GPS positions of the samples that were chosen for age dating are indicated.

Additionally, two country rocks from the Uluguru and the Mahenge Mountains have been dated. For descriptive and comparative reasons the mineral chemistry and geochemistry of the anorthosites and country rocks were investigated in detail.

1.1. Overview of the prevalent age domains in Tanzania

The crystallization ages of magmatic rocks from the Paleoproterozoic Usagaran Belt and the Neoproterozoic
### Table 2
Overview of age data from literature for significant age domains in Central and Northern Tanzania (and S. Kenya)

<table>
<thead>
<tr>
<th>Age domains</th>
<th>Reference</th>
<th>Age (Ma)</th>
<th>Error (Ma)</th>
<th>Method</th>
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<tr>
<td><strong>ARCHEAN AGES</strong></td>
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<tr>
<td>Dodoman gneisses</td>
<td>Wendt et al. (1972)</td>
<td>2529 ± 60</td>
<td>X</td>
<td>Rb–Sr WR</td>
</tr>
<tr>
<td>380 magmatic rocks of the Tanzanian Shield</td>
<td>Bell and Dodson (1981)</td>
<td>younger event: 2540</td>
<td>X</td>
<td>Rb–Sr WR</td>
</tr>
<tr>
<td>Uluguru Mountains (Jensen Quarry)</td>
<td>Muhongo et al. (2001)</td>
<td>2740 ± 3 (Zr)</td>
<td>X</td>
<td>SHRIMP II + U–Pb</td>
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<tr>
<td>Uluguru Mountains (Mikese Quarry)</td>
<td>Muhongo et al. (2001)</td>
<td>2705 ± 0.3 (Zr)</td>
<td>X</td>
<td>SHRIMP II + U–Pb</td>
</tr>
<tr>
<td>Wami River (Granodiorite WM2)</td>
<td>Muhongo et al. (2001)</td>
<td>2608 ± 0.2 (Zr)</td>
<td>X</td>
<td>U–Pb</td>
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<tr>
<td>Usambara (Amat Atra, AM3-Metapelite)</td>
<td>Muhongo et al. (2001)</td>
<td>1094.6–2023.8 (Zr)</td>
<td>X</td>
<td>U–Pb</td>
</tr>
<tr>
<td>Archean Granites (three samples)</td>
<td>Maboko and Nakamura (1996)</td>
<td>2611 ± 58 Ma isochron</td>
<td>X</td>
<td>Rb–Sr WR</td>
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<tr>
<td>Ruaha Orthogneiss (Lake River TZ00/06)</td>
<td>Sommer et al. (2003)</td>
<td>2969 ± 0.3 (Zr)</td>
<td>X</td>
<td>SHRIMP II</td>
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<tr>
<td>Kilosa Metaquartzite (TZ00/10)</td>
<td>Sommer et al. (2003)</td>
<td>2124 ± 5 (Zr)</td>
<td>X</td>
<td>U–Pb</td>
</tr>
<tr>
<td>Kidete granite</td>
<td>Reddy et al. (2003)</td>
<td>2698 ± 15 Ma discordia (Zircon)</td>
<td>X</td>
<td>SHRIMP II</td>
</tr>
<tr>
<td>Granite gneiss (Usambaran)</td>
<td>Reddy et al. (2003)</td>
<td>2705 ± 11 Ma discordia (Zr)</td>
<td>X</td>
<td>SHRIMP II</td>
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<td><strong>USAGARAN</strong></td>
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<td></td>
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<td>Usagaran granitoids</td>
<td>Wendt et al. (1972)</td>
<td>1870 ± 34</td>
<td>X</td>
<td>Rb–Sr WR</td>
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<td>Usagaran granitoids</td>
<td>Wendt et al. (1972)</td>
<td>1942 ± 82</td>
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<tr>
<td>Lukumburu-Wino Granites surrounding rock (5 granites, gneisses)</td>
<td>Prior et al. (1979)</td>
<td>1771 ± 145 (Zr)</td>
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<td>Rb–Sr WR</td>
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<td>Ilulagranite</td>
<td>Maboko and Nakamura (1996)</td>
<td>1910 ± 89 Ma isochron</td>
<td>X</td>
<td>Rb–Sr WR</td>
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<tr>
<td>Usagaran granitoids (10 samples near large)</td>
<td>Maboko and Nakamura (1996)</td>
<td>1872 ± 91 Ma isochron</td>
<td>X</td>
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<tr>
<td>Kidete granite (undeformed two Fsp granites)</td>
<td>Reddy et al. (2003)</td>
<td>1877 ± 7 Ma discordia (Zr)</td>
<td>X</td>
<td>SHRIMP II</td>
</tr>
<tr>
<td>Ruaha granodiorite gneiss (TZ00/02)</td>
<td>Sommer et al. (2003)</td>
<td>1870 ± 5 (Zr)</td>
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<td>SHRIMP II</td>
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<td>Ruaha trondhjemite gneiss (TZ00/07)</td>
<td>Sommer et al. (2003)</td>
<td>1898 ± 2 Ma (Zr)</td>
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<td>SHRIMP II</td>
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<tr>
<td>Eclogite</td>
<td>Möller et al. (1995)</td>
<td>2010 ± 12 Ma discordia (Ms, Bl, Sp)</td>
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<td>U–Pb</td>
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<tr>
<td>Eclogite</td>
<td>Collins et al. (2004)</td>
<td>1999 ± 1.1 Ma (Zr)</td>
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<td>U–Pb SHRIMP</td>
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<td>Eclogite</td>
<td>Muhongo et al. (2001)</td>
<td>1878 ± 1 ± 1 Ma (Zr)</td>
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<td>U–Pb</td>
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<tr>
<td>Taita Hills Unit 4K120 (migm. orthogranits)</td>
<td>Hausenberger et al. (2006)</td>
<td>1447 ± 33 Ma (Zr inherited core)</td>
<td>X</td>
<td>U–Pb SHRIMP</td>
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<tr>
<td>Pare Mountains and Loboa Serri (Granulites)</td>
<td>Sponor et al. (1970)</td>
<td>738 ± 8 Ma</td>
<td>X</td>
<td>Rb–Sr WR</td>
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<tr>
<td>Pare Mountains and Loboa Serri (Granulites)</td>
<td>Sponor et al. (1970)</td>
<td>736 ± 63 Ma</td>
<td>X</td>
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<tr>
<td>Wami River Anorthosite Ultrugan Mountains (Granulites)</td>
<td>Muhongo et al. (1985)</td>
<td>714 ± 49 Ma (Zr)</td>
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<td>U–Pb</td>
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<tr>
<td>Eastern Granulites–Konsoke Quarry (Chalcoser)</td>
<td>Muhongo et al. (2001)</td>
<td>841 ± 22 Ma (Zr)</td>
<td>X</td>
<td>SHRIMP II</td>
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### Table 2 (Continued)

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<tr>
<th>Age domains</th>
<th>Reference</th>
<th>Age</th>
<th>M/X Method</th>
<th>Method</th>
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<tbody>
<tr>
<td>Uluguru Mountains–Granulites</td>
<td>Muhongo et al. (2001)</td>
<td>725 ± 14 (Zrn)</td>
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<td>SHRIMP II</td>
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<td>Uluguru Mountains–anulites MVZ3</td>
<td>Muhongo et al. (2001)</td>
<td>745 ± 5 (Zrn)</td>
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<td>SHRIMP II</td>
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<tr>
<td>Usambara (Amani Area, AM1–Metapelite)</td>
<td>Muhongo et al. (2001)</td>
<td>737.5 ± 1.1</td>
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<td>U–Pb</td>
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<td>Usambara (Amani Area, AM3–Metapelite)</td>
<td>Muhongo et al. (2001)</td>
<td>753.6 ± 0.9 Ma (Zrn)</td>
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<td>U–Pb</td>
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<td>Usambara Complex (8 Orthogneisses)</td>
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<td>815.5 ± 58 Ma</td>
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<td>U–Pb SHRIMP</td>
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<td>Galana Shear Zone Unit 2KGB3</td>
<td>Hauzenberger et al. (2006)</td>
<td>976 ± 19 Ma (Zrn)</td>
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<td>U–Pb SHRIMP</td>
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<td>Sagala Hills Unit 3 Sample 142</td>
<td>Hauzenberger et al. (2006)</td>
<td>871 ± 7 Ma (Zrn)</td>
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<td>U–Pb SHRIMP</td>
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<td>Sagala Hills Unit 3 K209</td>
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<td>959 ± 43 Ma (Zrn)</td>
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<td>Taits Hills Unit 4 K120 (migm. orthogneiss)</td>
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<td>864 ± 33 Ma (Zrn)</td>
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<td>U–Pb SHRIMP</td>
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<td>PAN-AFRICAN</td>
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<td>Usagaran domain (5 banded gneisses)</td>
<td>Priem et al. (1979)</td>
<td>589 ± 70 Ma</td>
<td>M</td>
<td>Rb–Sr WR</td>
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<td>Muscovite and Biotite</td>
<td>Priem et al. (1979)</td>
<td>526±465 Ma</td>
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<td>Rb–Sr and K–Ar</td>
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<tr>
<td>North-western Tanzania</td>
<td>Malvoso (2000)</td>
<td>648 ± 14 Ma(Bfl)</td>
<td>M</td>
<td>Rb–Sr</td>
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<td>Eclogite</td>
<td>Möller et al. (1995)</td>
<td>501 ± 26 Ma (Mnx, Ri, Spb)</td>
<td>M</td>
<td>U–Pb</td>
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<tr>
<td>Raudha Metapelite (TZ00/04)</td>
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<td>641 ± 3.8 Ma (Zrn)</td>
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<td>SHRIMP II</td>
</tr>
<tr>
<td>Uluguru Minda Quarry (granite gneiss)</td>
<td>Sommer et al. (2003)</td>
<td>633 ± 6.6 Ma (Zrncore)</td>
<td>M</td>
<td>SHRIMP II</td>
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<tr>
<td>Usambara Complex (Orthogneisses)</td>
<td>Muhongo et al. (2001)</td>
<td>640 Ma (Zrn)</td>
<td>M</td>
<td>U–Pb</td>
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<tr>
<td>Usambara (Amani Area, AM1–metapelite)</td>
<td>Muhongo et al. (2001)</td>
<td>641.4 ± 0.9 Ma (Zrn)</td>
<td>M</td>
<td>U–Pb</td>
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<td>Usambara (Amani Area, AM2–granodiorite)</td>
<td>Muhongo et al. (2001)</td>
<td>641.1 ± 0.9 Ma (Zrn)</td>
<td>M</td>
<td>U–Pb</td>
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<tr>
<td>Usambara (Amani Area, AM4–magmatic rock)</td>
<td>Muhongo et al. (2001)</td>
<td>656.4 ± 0.9 Ma (Zrn)</td>
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<td>U–Pb</td>
</tr>
<tr>
<td>Usambara granulites</td>
<td>Möller et al. (2000)</td>
<td>~625 Ma (Mnx)</td>
<td>M</td>
<td>U–Pb</td>
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<tr>
<td>Usambara granulites</td>
<td>Malvoso (2001)</td>
<td>~605 Ma (Grt)</td>
<td>M</td>
<td>U–Pb</td>
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<tr>
<td>Pare Mountains (Pyroxen gneiss)</td>
<td>Malvoso and Lenoir (1994)</td>
<td>645 ± 10 Ma (Zrn)</td>
<td>M</td>
<td>U–Pb</td>
</tr>
<tr>
<td>Wami River (Metapelite WMI)</td>
<td>Muhongo et al. (2001)</td>
<td>641 ± 0.9 Ma (Zrn)</td>
<td>M</td>
<td>U–Pb</td>
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<tr>
<td>Uluguru Complex (Pegmatuates)</td>
<td>Cahen and Snelling (1966)</td>
<td>597 ± 10 Ma (Uraninite)</td>
<td>M</td>
<td>U–Pb</td>
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<tr>
<td>*</td>
<td>Cahen and Snelling (1966)</td>
<td>611 ± 10 Ma (Uraninite)</td>
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<td>U–Pb</td>
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<td>Uluguru</td>
<td>Malvoso et al. (1989)</td>
<td>626 ± 3 Ma (Hbl)</td>
<td>M</td>
<td>Ar-Ar</td>
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<td>Uluguru Complex (Anorthosite)</td>
<td>Malvoso and Nakamura (1995)</td>
<td>633 ± 7 Ma (Grt)</td>
<td>M</td>
<td>Sm–Nd</td>
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<td>&quot;(Granulite)</td>
<td>Malvoso and Nakamura (1995)</td>
<td>618 ± 16 Ma (Grt)</td>
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<td>Sm–Nd</td>
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<td>Uluguru Granulites</td>
<td>Möller et al. (2000)</td>
<td>~625 Ma (Sph)</td>
<td>M</td>
<td>U–Pb</td>
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<td>Uluguru Mountains–Granulites KIR1</td>
<td>Muhongo et al. (2001)</td>
<td>639 ± 13 Ma (Zrn)</td>
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<td>SHRIMP II</td>
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<td>642 ± 5 (Zrn)</td>
<td>M</td>
<td>SHRIMP II</td>
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<td>Furua Complex (Granulites)</td>
<td>Coolen et al. (1982)</td>
<td>652 ± 10 Ma (Zrn)</td>
<td>X</td>
<td>4 fractions U–Pb system</td>
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<td>Furua Complex (Granulites) *</td>
<td>Andriessen et al. (1985)</td>
<td>M1 Hornblende: 630 Ma</td>
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<td>Andriessen et al. (1985)</td>
<td>M2 Hornblende: 630 Ma</td>
<td>M</td>
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<th>M/X</th>
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<td>Galana Shear Zone Unit 1 K13 (migm. gneiss)</td>
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<td>541 ± 17 Ma (Zrn)</td>
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<td>Hauzenberger et al. (2006)</td>
<td>548 ± 6 Ma</td>
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<td>Galana Shear Zone Unit 2 KGB35 (migm. tonalite)</td>
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<td>651 ± 20 Ma (Zrn)</td>
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<td>Galana Shear Zone Unit 2 LF1 (conc. pegmatite)</td>
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<td>586 ± 10 Ma (Zrn)</td>
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<td>U-Pb SHRIMP</td>
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<td>U-Pb SHRIMP</td>
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Abbreviations in the row M/X are used for the subdivision of metamorphic (‘M’) vs. crystallization (‘X’) ages as interpreted in literature.

Mozambique Belt in Tanzania are summarized in Table 2. Three domains of formation ages can be distinguished. (1) Late Archean ages ranging between 2700 and 2500 Ma are interpreted as igneous emplacement ages and related to the formation of the Tanzanian Craton (e.g. Wendt et al., 1972; Bell and Dodson, 1981). Archean ages can be mainly found in the Craton, but also in the Usagaran Belt and in the Mozambique Belt (Sommer et al., 2003; Muhongo et al., 2001). (2) The second age domain is represented by the Usagaran–Ubendian orogeny between 2000 and 1800 Ma (Wendt et al., 1972; Reddy et al., 2003; Collins et al., 2004). These ages mainly occur in the Usagaran orogen, but subordinate scattered ages also can be found within the eastwards adjacent units of the Mozambique Belt (Sommer et al., 2003; Hauzenberger et al., 2006). These ages mainly occur in the Usagaran orogen, but subordinate scattered ages also can be found within the eastwards adjacent units of the Mozambique Belt (Sommer et al., 2003; Hauzenberger et al., 2006). (3) Early Neo-protorozoic crystallization ages (700–1000 Ma) are rarely documented only from the Eastern Granulites (Maboko et al., 1985; Muhongo et al., 2001; Hauzenberger et al., 2006) (Table 2).

Neo-protorozoic metamorphism in the Mozambique Belt is dated at ca. 640 Ma (Möller et al., 2000; Muhongo et al., 2001; Sommer et al., 2003) and reached granulite facies in the Eastern Granulites (Table 2). Lower grade coeval metamorphism, however, can be traced through the Mozambique Belt and the Usagaran Belt to the eastern margin of the Tanzania Craton (Reddy et al., 2003; Fritz et al., 2005).

Most age data shown in Table 2 are related to collision events. However, little is known about the rifting stage of the Gondwana cycle that must have occurred prior to 640 Ma. In contrast to the Mozambique Belt in Kenya and Sudan, where ophiolites are reported (Vearncombe, 1983; Kröner et al., 1992; Ries et al., 1992; Berhe, 1990), no remnants of ophiolitic material in Tanzania have been identified. The fact that Muhongo and Lenoir (1994) found ages of this time span in the Uluguru anorthosites, and the observation that anorthosites might have been generated in extensional plate boundaries, is another impulse for this study.

1.2. Geological setting of the anorthosite complexes in the Eastern Granulites

The Eastern Granulites are a roughly NE–SW trending chain of mountain ranges (up to 2500 m altitude)
forming the eastern margin of the Mozambique Belt in Tanzania (Fig. 1). The anorthosites are characteristic constituents of the Eastern Granulites and were tectonically incorporated into a sequence of metamagmatic deep crustal rocks (enderbites and migmatitic gneisses). Structurally overlying portions of the granulites are composed of meta-sedimentary rocks (marbles, mica-schists). The whole sequence, including the anorthosites, was intensely deformed and metamorphosed at high-pressure granulite facies at about 640 Ma.

From North to South the anorthosite bodies discussed here are (Fig. 1):

(1) The 'Ikongwe' anorthosite complex in the North Pare Mountains (close to the village 'Same') covers an area of ca. 100 km². Some smaller lenses are also found further south in the high-grade gneisses of the Pare-Usambara Mountains. Central portions of the 'Ikongwe' body are coarse grained and undeformed. A marginal zone up to 3 km in width, however, is strongly foliated together with the host gneisses (Pohl and Prochaska, 1983). This deformation led to intensely sheared mafic enclaves of garnet around clinopyroxene in a feldspathic matrix (Fig. 2a). The anorthosite complex contains mafic segregations, meta-dolerite sills and concordant dykes.

(2) The Uluguru anorthosite suite is a pear-shaped body in the central Uluguru Mountains with a north-south extension of ca. 25 km and a west-east extension of ca. 12 km and reaches three times the size of the Ikongwe anorthosite complex in the Pare Mountains (Sampson and Wright, 1964). The body is bordered by high-grade NE–SW striking sinistral strike-slip zones that juxtaposed it against the surrounding basement gneisses (mostly enderbites and migmatitic orthogneisses) (Fig. 2b). Smaller occurrences of anorthosites are also found in the western Uluguru Mountains but are not discussed here. All rocks of this anorthosite massif are intensely sheared with elongated lenses and bands of mafic material in the anorthositic matrix. The anorthosite suite comprises different rock types ranging from true anorthosite to anorthositic gabbros (Fig. 2c) as well as spatially related felsic and mafic magmatic rocks (Fig. 2d).

(3) In the Mahenge Mountains, minor occurrences of garnet-rich anorthositic gabbros are mapped (Birch

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Fig. 2. Photographs from outcrops where geochronology samples were taken (sample numbers are added). (a) The Pare Mountains anorthosite. (b) The Uluguru basement rock—a highly foliated migmatitic gneiss with flat-lying isoclinal folds. (c) Strongly sheared anorthosite with mafic bands. (d) Outcrop of the spatially related granitoid in the central part of the Uluguru anorthosite suite from which TV02/46a (felsic part) and TV02/46c (mafic part) are taken.
and Stephens, 1962). The largest body is ca. 5 km² in size in the northwest of the Mahenge Mountains (location ‘Iromue Hill’). The studied samples from this body are only marginal meta-tonalitic and monzogranitic rocks, spatially related to the anorthositic gabbros (that could not be reached in the field). The sample numbers and locations of the dated samples are shown in Fig. 1.

2. Petrography and mineral chemistry

Precambrian and Archean anorthosites can generally be distinguished by their igneous mineral chemistry and magmatic textures (e.g. Ashwal, 1993). Archean anorthosites are typically emplaced as sill-like bodies and magmatic textures (e.g. Ashwal, 1993). Archean anorthosites are typically emplaced as sill-like bodies and magmatic textures (e.g. Ashwal, 1993). Archean anorthosites are typically emplaced as sill-like bodies and magmatic textures (e.g. Ashwal, 1993).

2.1. Pare Mountains

2.1.1. The anorthosite

The anorthosite complex of the Pare Mountains comprises two rock types: (1) the metamorphosed anorthosite with transitions to leucogabbro (samples MB81, A157, A158a, A165) and (2) mafic dykes (samples A161, A162).

The anorthosite shows a highly deformed dynamically recrystallized fabric consisting largely of plagioclase (Pl) including deformed mafic lenses (Fig. 2a) with clinopyroxene (Cpx), paragastic amphibole (Prg) and a coronitic garnet (Grt) (Figs. 2a and 3a). The Pl grains show a variation in An content that can be grouped in three clusters (Fig. 4a) between XAn = 0.53–0.61, 0.68–0.72 and 0.80–0.86 (Table 3). The highest An contents are found in the Pl inclusions in Cpxs (up to 0.99) or in cores of concentrically zoned Pl grains in the matrix.

The composition of Prg is XScP = 0.74 and Si = 6.20 a.p.f.u. Small idiomorphic grains of the Grt coronae show a continuous growth profile with a slight increase of XAlm (0.36–0.40) and XPrp (0.37–0.43) from core to rim and a slight decrease of XGrt (0.25–0.17) from core to rim. The Mn content (XMn = 0.01) and the XGrt (around 0.50) remains constant throughout a ca. 400 mm long profile. The Cpx is diopside with XFe = 0.17–0.18. The co-existing assemblage Grt–Cpx is shown in Fig. 4b. Minor constituents are abundant Rt crystals, Fe–Ti oxides as well as Ap. Zircon is extremely rare.

2.1.2. Mafic dykes within the anorthosite

Mafic dykes (Fig. 3b) contain Pl grains with lower XAn = 0.36–0.46 than those in the anorthositic samples. The metamorphic Grt is Alm-rich but shows a range of compositions (XAlm between 0.35 and 0.73). Clinopyroxene has a low XScP (0.05 and 0.18) and Cpx only occurs in the mafic dykes with XScP between 0.20 and 0.30.

2.2. Uluguru Mountains anorthosite suite

2.2.1. The anorthosite

The anorthosite suite of the Uluguru Mountains comprises three rock types: (1) anorthosites and leucogabbros (TV02/05, TV02/09, TV02/10); (2) granitoids with mafic segregations, that are spatially related and exposed in the central portions of the anorthosite suite (TV02/46a, TV02/46c); and (3) various mafic rocks that are spatially associated with the anorthosite intrusion (TV02/18, TV02/22b). A sample of granulite facies migmatite orthogneiss from the country rocks was also analysed (TV02/45).

Anorthosites close to the tectonic boundary to the granulites have the mineral assemblage Pl–Grt–Cpx–Prg ± Scp ± Kfs (e.g. TV02/05). The feldspathic matrix contains mm to cm sized elongated lenses of mafic minerals (Figs. 2c and 3c and d). The dynamically recrystallized Pl grains (XAn = 0.43–0.48) show deformation twinning and subgrain rotation deformation along the grain boundaries. Small (0.1 mm) idiomorphic and inclusion-free Grt crystals are distributed in the matrix or around co-existing elongated Cpx aggregates. Both Grt types have the same mineral chemistry, typically: XAlm = 0.48–0.50, XScP = 0.11–0.19, XMn = 0.30–0.37 and XMg = 0.02; however, the Grt chemistry can vary between individual samples as seen in Fig. 4b. The
Fig. 3. Microphotographs from the dated rocks. (a) Anorthosite from the Pare Mountains with mafic lense and Grt corona. (b) Mafic dyke from the Pare Mountains anorthosite. (c) Assemblages of the mafic lenses, that are sheared into the Pl fabric of sample from the Uluguru Mountains anorthosite. (d) Anorthosite from the Uluguru Mountains with mafic lense. (e) Zircon-rich granitoid rock from the central part of the Uluguru anorthosite suite. (f) Migmatitic basement gneiss adjacent to the Uluguru anorthosite suite. (g) Meta-tonalite that is associated with the anorthositic rocks from the Mahenge Mountains. (h) Basement gneiss from the Mahenge Mountains.
Table 3
Representative mineral analyses of the investigated feldspars

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a Mineral formulae normalized to eight oxygens. \( X_{\text{Ab}} = \frac{\text{Na}}{(\text{Na} + \text{Ca} + \text{K})} \); \( X_{\text{An}} = \frac{\text{Ca}}{(\text{Na} + \text{Ca} + \text{K})} \); \( X_{\text{Ksp}} = \frac{\text{K}}{(\text{Na} + \text{Ca} + \text{K})} \).
Cpx composition is more homogeneous in different samples (e.g. $X_{Fe} = 0.21–0.29$). Along the rim of the Cpx aggregates retrograde greenschist to amphibolite crystals occur with $X_{Fe} = 0.6$ (decreasing to 0.45 with increasing retrogression along rims). The Si content of amphiboles ranges between 5.90 and 6.20 a.p.f.u., corresponding to the variation in $X_{Mg}$. The Na content of amphibole does not exceed ca. 1 wt%. Within the mafic lenses in this anorthosite sample Kfs crystals occur (Fig. 3c). The $X_{Or}$ of this Kfs is 0.93. The Ba content ranges between 3.20 and 4.00 wt%. Accessory minerals are Ap, Sph and intergrowths of Rt and Ilm as well as Scp with $Eq_{An} = 0.68–0.70$ and $X_{Co} = 0.85–0.86$, $X_{Cl} = 0.11$ and $X_{SO} = 0.02–0.05$.

Some samples (e.g. TV02/10) have a varying Pl chemistry between $X_{An} = 0.50$ and 0.61 (Fig. 4a; Table 3). Feldspar composition with $X_{An} = 0.50$ and 0.61 (Fig. 4a; Table 3). Highly flattened mafic components contain porphyroclasts of Di with $X_{Fe} = 0.20$ (rim) and $X_{Fe} = 0.30$ (core) and surrounding Grt (Fig. 4b). The Di crystals are overgrown by dynamically recrystallized brown Prg with $XMg = 0.58–0.62$ and $Si = 5.82–6.04$ a.p.f.u. (lowest in the deformation lamellae). Some Prg crystals are in equilibrium with the idiomorphic Grt aggregates that occur in the mafic shear bands. These amphiboles show a somewhat higher $X_{Or}$ of 0.64–0.71 and Si of 5.92 a.p.f.u. The Grt is single phase, idiomorphic and shows flat chemical profiles with $X_{Alm} = 0.42$, $X_{Prp} = 0.29$, $X_{Grs} = 0.27$ and $X_{Sp} = 0.02$. Scapolite is a common mineral in the felsic matrix as well as in the mafic layers or occurs as single grains interstitially between the Grt–Cpx–Am aggregates. The mineral chemistry of the Scp is $Eq_{An} = 0.68–0.70$ and $X_{Or} = 0.80–0.81$, with $X_{Co} = 0.86–0.89$, $XMg = 0.01$ and $X_{SO} = 0.09–0.13$. Very fine-grained accessory minerals are Ilm, minor Rt and Ap. Zrn grains could not be found in thin sections.

2.2.2. Spatially associated felsic rocks (granitoids) within the Uluguru anorthosite

Spatially associated granitoids (TV02/46a) occur in the central portions of the anorthosite suite (Figs. 2d and 3e). Extreme ductile deformation has led to grain size reduction by dynamic recrystallization, and produced an equigranular fabric with straight 120° grain boundaries. The Kfs crystals show microscale deformation lamellae of Pl and the Pl crystals show Kfs lamellae. Large Grt porphyroblasts (cm size) also show dynamically recrystallized subgrains and internal corona structures of Spl and Mag with a reaction rim of Pl. Accessory minerals are Ap, Ilm and Zrn crystals of mm size, which are elongated and oriented in the foliation planes of the deformed matrix. Some grains show asymmetric deformation features with opaque phases (Ilm, Hem) in pressure shadows around Grt and Zrn. The ductility of Ilm at high temperature in anorthositic rocks has been described by Duchesne (1999b). The $X_{An}$ of Pl ranges between 0.20 and 0.40 (Fig. 4a). K-feldspar shows a variation in $X_{Si}$ between 0.71 and 0.83 (in the matrix, deformation lamellae or in the corona structures around
Ilm in Grt). The Grt crystals are Fe-rich with flat profiles and $X_{\text{An}} = 0.60$, $X_{\text{Fe}} = 0.23$, $X_{\text{Na}} = 0.17$ and $X_{\text{Al}} = 0.02$. Spinel has an $X_{\text{Mg}} = 0.31 - 0.51$ and ZnO of up to 20 wt% (gahnite).

2.2.2.1. Mafic enclaves in the granitoid. The granitoid contains abundant ultramafic enclaves (TV02/46c) (Figs. 2d and 4b) consisting of Grt–Opx–Cpx (±Am ± Bt) aggregates, that are dynamically recrystallized into equidimensional grains. The Grt grains ($X_{\text{An}} = 0.34 - 0.38$, $X_{\text{Fe}} = 0.46 - 0.51$, $X_{\text{Co}} = 0.15$ and $X_{\text{Si}} = 0.01$) are distributed as bands intercalated with Cpx ($X_{\text{Si}} = 0.01 - 0.13$). Am ($X_{\text{An}} = 0.85 - 0.90$, $Si = 6.00 - 6.20$ a.p.f.u.) and Bt ($X_{\text{An}} = 0.87$). The matrix consists of aggregates of Cpx, Opx and Grt (Fig. 2d, Table 3). The dynamically recrystallized rock has the mineral assemblage Grt–Cpx–Prg–Crn–Hc–An. The composition is $X_{\text{An}} = 0.54 - 0.57$ and $X_{\text{Me}} = 0.66 - 0.67$, with $X_{\text{Co}} = 0.26 - 0.35$, $X_{\text{Cr}} = 0.00 - 0.03$ and $X_{\text{Zn}} = 0.63 - 0.71$. Amphiboles are pargasitic with $X_{\text{Prp}} = 0.64 - 0.68$ and Si = 6.10 - 6.40 a.p.f.u. The $X_{\text{Fe}}$ of Cpx is 0.18 - 0.26.

2.2.3. Spatially associated mafic rocks in the Uluguru anorthosite

Mafic rocks (TV02/14, TV02/15, TV02/22b, TV02/18), associated with the anorthosite are typically highly deformed bands of Grt, Px and Prp (Fig. 2c). The mafic mineralogy is comparable to the highly sheared mafic enclaves in the anorthositic part discussed above. Two other rock types that are different from the mafic lenses occur mainly at the boundary of the anorthosite body. (1) Grt-garnetiferites have been already documented by Sampson and Wright (1964) (TV02/22b). The typical mafic mineral assemblage is Grt–Cpx–Am ± Pl. The composition of the Cpx is characterized by $X_{\text{Si}} = 0.18 - 0.27$, the $X_{\text{Mg}} = 0.10 - 0.15$ wt%. The Grt composition is represented by $X_{\text{An}} = 0.39 - 0.44$, $X_{\text{Fe}} = 0.32 - 0.27$, $X_{\text{Co}} = 0.28$, $X_{\text{Si}} = 0.01$. Plagioclase occurs in symplectites and has an An-rich composition ($X_{\text{An}} = 0.81 - 0.83$). (2) Corundum-bearing Grt-Cpx-Hbl-An fels is a rock type at the transition between anorthosite and basement (TV02/18). The dynamically recrystallized rock has the mafic mineral assemblage Grt–Cpx–Prg–Crn–Hc–An. The Crn-bearing Grt–Am–An fels is enriched in Mg, Al and Ca. The matrix consists of equigranular amphibole (pargasitic: $X_{\text{Si}} = 0.90 - 1.00$, $Si = 6.00 - 6.20$ a.p.f.u.) and Cpx ($X_{\text{Si}} = 0.06 - 0.08$). Around the abundant Crn crystals, which form mm to cm sized grains, multi-layered corona textures developed, with green hercynitic Sph at the immediate contact to the Crn. The Sph reacts to form a fine-grained symplectite of Am and Hc with the same composition as those in the matrix and this symplectite separates the central parts of the corona texture from the outer rim formed by polygonal plagioclase crystals of almost pure An.

2.2.4. Meta-igneous basement rocks of the Uluguru Mountains

The meta-igneous suite (TV02/45) surrounding the Uluguru anorthosite is composed of orthogneisses with the assemblage Grt–Cpx–Hbl and Scp porphyroclasts in a matrix consisting of Qtz and Pl (Fig. 3f). The composition of Pl is $X_{\text{An}} = 0.32 - 0.36$. Garnet crystals are characterized by a relatively flat profile with a retrograde rim. The following mole fractions are typical: $X_{\text{An}} = 0.49$ (rim up to 0.56), $X_{\text{Fe}} = 0.32 - 0.25$ (decreasing towards the rim to 0.25), $X_{\text{Co}} = 0.07 - 0.18$ and $X_{\text{Si}} = 0.01 - 0.02$ (from core to rim). In contrast to the Scp in anorthositic rocks, these are sulphur-rich. The composition is $X_{\text{An}} = 0.54 - 0.57$ and $X_{\text{Si}} = 0.66 - 0.67$, with $X_{\text{Co}} = 0.26 - 0.35$, $X_{\text{Cr}} = 0.00 - 0.03$ and $X_{\text{Zn}} = 0.63 - 0.71$. Amphiboles are pargasitic with $X_{\text{Prp}} = 0.64 - 0.68$ and Si = 6.10 - 6.40 a.p.f.u. The $X_{\text{Fe}}$ of Cpx is 0.18 - 0.26.

2.3. Mahenge Mountains

2.3.1. Spatially associated felsic rocks (tonalites and granitoids) of the Mahenge anorthosite

From the small anorthositic gabbro at ‘Ironve Hill’ in the NW Mahenge Mountains only the spatially associated felsic rocks could be sampled due to bad outcrop situation (T93, T94/2, T96). The representative tonalitic mineral assemblage is Grt–Cpx–Pl–Qtz ± Prg (Fig. 3g). The matrix consists of Pl and Qtz in a ratio of ca. 1:1. The chemistry of Pl is characterized by $X_{\text{An}} = 0.43 - 0.44$ (Fig. 4a, Table 3). Along the grain boundaries of Pl some late sericite can be observed. Garnet and Cpx have both been equilibrium in the metamorphic peak, but both minerals are retrogressed at the rim (Fig. 4b). The $X_{\text{Fe}}$ of Cpx is 0.33 - 0.38. Garnet is partly consumed on the rim by retrograde reactions and has therefore an irregular shape. The chemistry of Cpx is $X_{\text{An}} = 0.55 - 0.57$ (flat throughout the profile), $X_{\text{Fe}} = 0.25 - 0.22$ (decreasing towards the rim to 0.10), $X_{\text{Co}} = 0.19 - 0.20$ (increasing to the rim to 0.32) and $X_{\text{Si}} = 0.02 - 0.03$ (flat profile). Small inclusions of Qtz, Ap, Ilm, Pl and Zrn are commonly observed in Grt cores. Small retrograde amphiboles occur around Cpx; these are pargasitic with $X_{\text{Prp}} = 0.60 - 0.63$ and Si = 6.20 - 6.40 a.p.f.u. Accessory minerals are Mag, Ap, zoned Ilm, Ep and Zrn.

Granitoids from this suite are represented by sample T96 and contain the mineral assemblage Grt–Kfs–Pl–Qtz. The composition of Pl is defined by $X_{\text{An}}$ of 0.22 - 0.24. A typical flat Grt profile is characterized by $X_{\text{An}} = 0.61 - 0.63$, $X_{\text{Fe}} = 0.10 - 0.12$, $X_{\text{Zn}} = 0.20 - 0.23$. 

2.4. Other xenoliths
and $X_{Sp}=0.06$. Similar to the granitoid of the Uluguru anorthosite suite this rock is rich in Zrn, which is aligned in the ductile matrix with pressure shadows of Ilm and Hem (Fig. 5g).

2.3.2. Meta-igneous basement rocks of the Mahenge Mountains

Sample C135 represents the basement of the Mahenge Mountains. It is a mafic Ca-rich granulite (Fig. 3h) with

Fig. 5. (a–h) Optical photographs of the dated Zrn fractions for the dated samples.
the mineral assemblage Grt–Cpx–Am–Bt–Cc ± Kfs in a matrix of nearly pure Pl with a patchy pattern due to variation in An and Ab content. Accessories are Cc, Tit, Ap, Ilm, Monazite, Zrn and Qtz. The irregular 0.5 mm Grt crystals show inclusions of Prg, Cc, Pl and Fsp (Pl and Kfs). The chemistry of Grt is $X_{\text{Al}} = 0.54$–0.66, $X_{\text{Fe}} = 0.09$–0.17, $X_{\text{Gr}} = 0.22$–0.33 and $X_{\text{Si}} = 0.01$–0.03. Within the Grt crystals inclusion trails filled with carbonate phases, such as Cc and Mgs occur. The larger Cpx I porphyroblasts have values of $X_{\text{Fe}} = 0.33$–0.38 and are replaced at the rim by a symplectic pattern of secondary Cpx II–Qtz–Cc. The Cpx II generation has $X_{\text{Fe}} = 0.32$–0.33. A second type of symplectite contains the assemblage Ilm–Qtz–Cc. In some cases along the rim of the first Cpx generation retrograde Bt and amphibole growth can be observed. Biotite has $X_{\text{Mg}} = 0.48$–0.57 and actinolitic Am shows $X_{\text{Mg}} = 0.56$–0.70 with Si content of 6.7–7.6 a.p.f.u.

3. Geochronology

Zircons are extremely sparse in anorthosites and therefore hard to extract (average amount of ca. 20–30 ppm Zr in the rocks). Nevertheless, in this study direct age dating was possible on single Zrn grains directly from the anorthosites. In some parts of the study area morphologies and textures from Zrn grains suggest recrystallization processes probably by high-grade metamorphism. So, the in situ relationship of magmatic and metamorphic textures of Zrn and surrounding metamorphic minerals is sometimes not easy to derive.

3.1. Analytical procedure for age dating

The first step in zircon separation was crushing and sieving to obtain mineral fractions between 63 and 180 μm. Further separation of the heavy minerals was performed with the heavy liquid Na-polytungstate (SOMETU) and the magnetic heavy minerals were removed by the FRANTZ magnetic separator. Hand-picking of the zircon grains was done under the binocular microscope (five to six samples of ca. 50 grains each on one mount). The mounts were put into epoxy resin together with a reference zircon crystal (Geostandards 91500 zircon), dried and polished to get a section through the centre of the grains. For information about the internal growth structure of the zircons, CL and BSE images were taken using the JEOL 6310 SEM in Graz at conditions of 5 nA and 15 kV and 5 nm gold coating. The U–Pb dating was performed using a CAMECA IMS1270 ion microprobe at the Swedish Museum of Natural History, Stockholm. Analyses and data reduction procedures follow those outlined by Whitehouse et al. (1999) and references therein. In brief, a ca. 5 nA O2+–primary beam was used in aperture illumination mode to sample nominal ca. 25 μm elliptical areas. The mass spectrometer was operated in monocollector mode at a mass resolution of ca. 5000, with secondary ions detected using an ion counting electron multiplier. Pb/U ratios, calibrated to 91500 include an error component propagated from the standard analyses in a particular session while Pb-isotope ratio errors are counting statistic based or observed errors. Correction for common lead uses measured $^{206}$Pb and assumes a present day composition from the Stacey and Kramers (1975) model. In many of the samples analysed with very low Pb concentrations, $^{206}$Pb count rates were statistically indistinguishable from background and, in these cases, we have not made a correction. All data reduction was carried out using the routines of Isoplot (Ludwig, 2001).

3.2. Description of zircons characteristics and age dating results

3.2.1. Pare Mountains

The zircon grains of sample MBS1 from the Pare Mountains anorthosite are light yellow to brownish in colour and contain circular inclusions of Ilm as well as tubular fluid inclusions (Fig. 5a). The grains range in size from 50 to 200 μm and are elongate to slightly rounded, mostly subhedral or multifaceted. Internal zoning is revealed in CL images and ranges from fine scale oscillatory zoning, which is typical for magmatic growth, to patchy textures and sector zoning (Fig. 6a). The grains are surrounded by a narrow (10 μm) rim indicating a metamorphic growth phase that was, however, too narrow to date. Nine spots were dated that give magmatic $^{207}$Pb/$^{206}$Pb ages between 2618 ± 15 and 2675 ± 15 Ma as shown in the Tera–Wasserburg diagram (Tera and Wasserburg, 1972) in Fig. 6b and Table 4. From these spots a mean $^{207}$Pb/$^{206}$Pb age was calculated at 2643 ± 16 Ma. The average Th/U ratio of the nine analyses is 1. The deviance of magmatic ages from the concordia is explained by later Pb-loss most likely during Neoproterozoic granulite facies metamorphism.

3.2.2. Uluguru Mountains

The zircons of the Uluguru anorthosite TV02/05 are colourless to light pink, very clear, predominantly oval-shaped and poor in inclusions (Fig. 3b). Some grains are slightly elongated with rare inclusions that are probably of igneous origin. Other grains are rounded and clearly metamorphic. An important characteristic is that many zircon grains are broken, irregularly grown or fractured.
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<td>g2-1</td>
<td>13 4.2</td>
<td>2.9</td>
<td>2580</td>
<td>0.73</td>
<td>n/a</td>
<td>0.0796 ± 0.0023</td>
<td>n/a</td>
<td>0.1695 ± 0.0031</td>
<td>1010 ± 17</td>
<td>1001 ± 18</td>
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### Table 4 (Continued)

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<tr>
<th>Grain/spot</th>
<th>Concentration (ppm)</th>
<th>Isotope ratio ( \pm \sigma )</th>
<th>Age ( \pm \sigma )</th>
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</tr>
<tr>
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<td>21 2.5 0.27 1240 [1.51]</td>
<td>n/a</td>
<td>0.0632 ± 0.0020</td>
</tr>
</tbody>
</table>

T942—Tonalite from the Mahengeanorthosithenorthosite suite

\( f_{206} \) is the amount of common 206Pb, estimated from measured 204Pb. Values in parentheses indicate that the 204Pb count was not significantly higher than detector background and in these cases, where common Pb correction has been applied, we do not report corrected 206Pb/238U and 207Pb/206Pb ratios (indicated by n/a in these columns). Th/U ratios are calculated from measured Th/U ratios relative to the standard. 207-Corrected ages are presented when <1200 Ma and use uncorrected ratios and assume common Pb 206Pb/238U = 0.703 ± 0.001. 206Pb/238U ages are presented for ages >1200 Ma and for common Pb corrected data for ages <1000 Ma.
Fig. 6. (a–f) CL images of the dated Zrn fractions with the corresponding age data of the considered spots for the anorthosites of the Pare and Uluguru Mountains. The concordia ages and the individual age data are shown on Tera–Wasserburg diagrams (207Pb/206Pb vs. 238U/206Pb). All data point error crosses are shown at the 2σ level. Calculation results are present at 95% confidence. Points that are taken for concordant or weighted mean ages are shaded.

(Fig. 5b). The internal zoning as seen from CL images is complex and shows a two phase growth with a bright core and a darker rim (Fig. 6c). The core is characterized by sector zoning and can reach sizes up to 300 μm. The rim is mostly very narrow and in rare cases reaches a width of more than 10 μm. Grain 9 in Fig. 6c has a specific habit. It resembles the type of zircon described by Bingen et al. (2001) as 'hat'-shaped, which forms a flat base in contact with Ilm from the Rogaland AMCG (anorthosite-mangerite-charnockite-granite) suite in the Western Caledonides in Norway. Although an in situ relation between Zrn and Ilm could not be observed in our sections, due to the rare occurrence of Zrn, the similarity to the description of the granulite facies zircons from Bingen et al. (2001) is striking. The sector zoning is often overprinted by a diffuse pattern and cracks, which seem to be healed again during metamorphism. Sector zoning indicates rapid changes in the growth
environment during the crystal development (kinetic factors, lattice diffusivity and degree of solution of the medium) and strong fluctuations of growth rates. The broad overgrowth is related to granulite facies metamorphism. The Th/U ratio of the cores has an average of 0.7 (nine spots) and the rims have Th/U values of 0.36.

The age data yield two different U–Pb age domains in the time span between 890 and 680 Ma (Table 4). Along the concordia (Fig. 6d) the variation in 207Pb/206Pb is little in this time range and the uncertainties are relatively large. Therefore, the age determinations of this sample are based on the 206Pb/207Pb ratio. The oldest age found in Zrn cores is 890 ± 20 Ma. A concordant age of 870 ± 10 Ma is calculated from three spots from magmatic Zrn cores. The subsequent metamorphic overprint is well-constrained from two spots at the metamorphic rim of grain 9 in Fig. 6c at ca. 680 Ma. The distinct age population that correlates with these analyses produces a concordant age of 693 ± 10 Ma (Fig. 6d). This is interpreted as Neoproterozoic metamorphism and 40–50 Ma earlier than the age of 640 Ma reported by Muhongo et al. (2001). The relation between core and rim in this sample is clear, so mixing can be excluded for the interpretation of the 693 ± 10 Ma age. More likely is a partial reset of the isotopic system during Neoproterozoic metamorphism at 640 Ma. The three other ages of 797 ± 5, 763 ± 5 and 737 ± 5 Ma are derived from cores with inclusions and cracks and are interpreted to have experienced Pb-loss or some mixing between the two age domains by metamorphic diffusive processes.

The second sample from the Uluguru anorthosite is TV02/10. Two different types of Zrn grains are found. One type is similar to those found of sample TV02/05. They are colourless, oval- to irregular-shaped and the complex patchy zoning is overprinted by cracks and inclusion tails (Fig. 5c, grain 1). A subordinate second type is idiomorphic and elongated, yellowish colour with opaque appearance and magmatic oscillatory zoning (Fig. 6e) as well as abundant tubular inclusions (Fig. 5c, grain 2). The size of both types ranges between 150 and 300 μm. The two types of zircons show two significant age domains (Fig. 6f; Table 4). The early Neoproterozoic crystallization is defined by the concordia age of 887 ± 20 Ma (clear grains; Th/U = 0.17–0.19) and the late Neoproterozoic metamorphic overprint is dated at 651 ± 15 Ma (opaque grains; Th/U = 0.1–0.4). Grain 2 has possibly grown from a melt during Neoproterozoic granulite facies metamorphism that led to the abundant migmatization seen in the rocks.

The associated granitoid from the Uluguru anorthosites suite (TV02/46a) shows zircons of mm size that display ductile deformation, elongated and oriented into the flow of the dynamically deformed matrix (Fig. 3e). Some grains are elongated and slightly rounded, but most of the zircons are irregularly shaped (Fig. 5d). The zircons are clear with a slightly dotted surface and light pink in colour. The internal structure is defined by magmatic oscillatory zonation (Fig. 7a), which is overprinted by complex diffusive processes by later metamorphism. Some grains show recrystallized rims and pressure shadows of Fe–Ti oxides (Fig. 3e). This indicates mobility of Zr during high-grade deformation and metamorphism leading to irregular diffusive rims. The original igneous oscillatory texture however is usually conserved in the cores of the large grains and could be dated. The in situ relationship of magmatic Zrn grains with diffusive overgrowths and metamorphic Fe–Ti oxides is easy to establish in this sample in thin sections.

The Zrn cores from the granitoids yield similar ages as found in the anorthosites (Fig. 7b; Table 4). The concordant ages cluster around the concordant age of 854 ± 33 Ma. Two measured spots are exceptions. One 238U/206Pb age is older (968 ± 28 Ma) but discordant from a magmatic core, possibly indicating an older period of Zrn growth in this rock. The metamorphic overprint could be dated from one grain at 632 ± 19 Ma. The Th/U ratios are generally high (Th/U = 0.7–1.5).

The orthogneiss from the Ulugurus basement TV02/45 contains zircons with idiomorphic magmatic cores, showing tubular inclusions and a dotted surface as well as a clear overgrowth (Fig. 5e). The CL images reveal complex internal structures with dark cores with oscillatory zoning patterns. In some grains the small core is followed by a dark rim that is homogeneous to slightly oscillatory. A luminescent metamorphic rim is mostly observed, sometimes with an irregular boundary to the core due to recrystallization (Fig. 7c).

The isotopic ratios of the Zrn cores in this orthogneiss give two different age domains (Fig. 7d). A first domain of spots yields ages around 986 ± 16 Ma and a second domain shows ages around 830–850 Ma. The oldest age group is interpreted as crystallization age from zircon cores. This group is defined by regression of uncorrected data points through common Pb composition. One age from a well-defined metamorphic rim shows the metamorphic overprint at 647 ± 11 Ma. A sequence of data subparallel to the concordia around 850 Ma is probably explained by Pb-loss to the Neoproterozoic event. A second age at 683 ± 12 Ma is from an older zircon that is partially reset by Neoproterozoic metamorphism (Table 4). Ages lying above the concordia may be explained by recent Pb-loss (as indicated in Fig. 7d).
Fig. 7. (a–h) CL images and corresponding age data of the rocks that are associated with the anorthosite suites of the Uluguru and Mahenge Mountains as well as the basement gneisses of both complexes. The concordia ages and the individual age data are shown on Tera–Wasserburg diagrams ($^{207}\text{Pb}/^{206}\text{Pb}$ vs. $^{238}\text{U}/^{206}\text{Pb}$). All data point error crosses are shown at the 2σ level. Calculation results are present at 95% confidence.
3.2.3. Mahenge Mountains

The associated tonalitic sample T94/2 from the Mahenge Mountains has zircons, which are between 50 and 250 μm in size, colourless and show a two phase growth with a small idiomorphic core that contains tiny inclusions and a distinct metamorphic rim (Fig. 5f). The grains are rounded, subhedral to slightly elongated or irregular. The small magmatic core is characterized in CL images by either broad oscillatory zoning or sector zoning, the metamorphic rims (30–250 μm) in contrast are homogeneous and bright (Fig. 7e). Some grains have a patchy core. There are grains similar to zircons within the granitoids of the Uluguru Mountains anorthosite suite, with pressure shadows of Fe–Ti oxides (Hem, Ilm) that can be found in a ductilely deformed matrix in associated granites (Fig. 5g). The Th/U ratio is around 0.3–0.6 in the cores and slightly lower in the rims (0.2–0.4).

Four Zrn cores from sample T94/2 show a weighted mean age of 792 ± 29 Ma. A Neoproterozoic metamorphic overprint was constrained at 658 ± 16 Ma as weighted mean age from zircon rims (Table 4).

The zircons of sample C135 from the basement of the Mahenge Mountains are light pink, rounded to oval with slightly elongated shape and very clear (Fig. 5h). Some grains are irregular or slightly faceted. Minor cracks and inclusions occur. The internal structure is defined by broader oscillatory to sector zoning of the core and a metamorphic rim, that is dark in contrast to the rim in the other samples (Fig. 7g). The grain size ranges between 50 and 200 μm.

The broad oscillatory zoning of C135 may not be exclusively magmatic, but can also be produced by migmatitic processes during high-grade metamorphism. The age of 730 ± 9 Ma was retrieved from a zircon core, which appears very bright and homogeneous without visible zoning (Fig. 7h). Most of the analysed points reflect the Neoproterozoic metamorphic age summarized as a concordia age of 652 ± 7 Ma (Table 4).

4. Geochemistry

4.1. Analytical methods

A first set of samples was analysed in the Activation Laboratories, Ontario, Canada. Major and trace elements of the samples together with international reference materials were analysed on a Thermo Jarrell Ash ENVIRO II simultaneous and sequential ICP and a Perkin-Elmer Optima 5000 ICP. For trace element analyses the sample solution was spiked with internal standards and further diluted and introduced into a Perkin-Elmer SCIEX ELAN 6000 ICP-MS using a proprietary sample introduction methodology. A second set of samples was analysed at the Institute of Earths Sciences at the University of Graz using a Bruker Pioneer S4 for the determination of major and trace elements. Selected trace elements and rare earth elements were analysed with a HP Agilent 4500 ICP-MS after 0.1 g of sample was digested with 0.1–0.2 ml HF and 5 ml HNO₃ in Teflon bombs at 220 °C for 24 h. International reference materials were routinely analysed with the unknown samples. The rock compositions are presented in Table 5.
Table 5
Representative geochemical analyses of the samples from the discussed anorthosites and associated rocks (major elements in %, trace elements in ppm)

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<th>Sample</th>
<th>Major Elements</th>
<th>Trace Elements</th>
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</thead>
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<tr>
<td>Sample 1</td>
<td>SiO₂: 51.63, Al₂O₃: 18.56, Fe₂O₃: 0.45</td>
<td>Ca: 56.4, Mg: 1.2</td>
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<td>Sample 2</td>
<td>SiO₂: 51.23, Al₂O₃: 18.29, Fe₂O₃: 0.39</td>
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<td>Sample 3</td>
<td>SiO₂: 51.38, Al₂O₃: 18.62, Fe₂O₃: 0.43</td>
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(Continued...)
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<td>0.1</td>
<td>1.5</td>
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</table>

Detection limits are according to the methods; n.d., not detected. Nos. 1–35 as in Fig. 9.
4.2. Geochemical composition

4.2.1. Pare Mountains

4.2.1.1. Major elements. The AFM diagram (Fig. 8a) shows a linear fractionation trend from metagabbroic rocks (A157, A164) to true anorthosites (A158b, A165, A160, A163, MB81) in the samples from the Pare anorthosite. The mafic dykes are also plotted and have a higher FeO content (A161, A162). These do not lie on a fractionation trend and are therefore not co-magmatic. Major elements are plotted against Al2O3 and are displayed in Fig. 9 (with abbreviated sample numbers as in Table 5). The CaO and Na2O distribution shows that the Pare anorthosites are richer in CaO and relatively poor in Na2O compared to the Uluguru samples (Fig. 9b and c). The linear trend from gabbroic to anorthositic rocks is best represented by MgO and FeO distribution versus Al2O3 (Fig. 9d and e), with gaps between the rock types. P2O5 is a minor constituent of the anorthosite and source of Fe–Ti–P-rich rocks, however, is widely debated in the literature (e.g. Owens and Dymek, 1992). A broader spectrum of element distribution is also obvious in Fig. 9. The Ca content of the Uluguru samples ranges from 9 to 11 wt% and the Na2O content from 4 to 5.5 wt% (Fig. 9b and c). The FeO and MgO contents show a linear increase of FeO with decreasing Al2O3 values displaying an increase in mafic components of the rock similar to the Pare samples (Fig. 9d and e). The P2O5–Al2O3 diagram exposes the presence of Ap in the samples TV02/47b and TV02/06 (Fig. 9f) that are typically associated with anorthosite bodies. The meaning and source of Fe–Ti–P-rich rocks, however, is widely debated in the literature (e.g. Owens and Dymek, 1992). A significant variation of the trace elements Ba and Sr is shown in Fig. 9g and h. Both elements occur in a higher concentration compared to the Pare anorthosites and show a wide scatter between the individual samples.

4.2.1.2. REE and trace element patterns. The REE concentrations of the rocks are normalized against chondritic values after Nakamura (1974). The anorthosites show continuously decreasing LREE values from La to Sm followed by a large positive Eu-anomaly (Fig. 10a). The HREE concentration follows a slight zig-zag pattern with positive anomalies in Tb and Tm and Lu with an upward trend at the end of the pattern. Two samples, MB81 and A157 exhibit a slightly negative anomaly in Tb. The REE distribution in the mafic dykes is about 10 times enriched relative to chondrites with a normalized La/Lu ratio of ca. 1.7–2.5 (Fig. 10b). Eu shows a small positive Eu-anomaly caused by the abundance of plagioclase. Trace element concentrations are normalized to primordial mantle after Taylor and McLennan (1985) and shown in Fig. 11a and b. The anorthositic samples exhibit a relatively uniform pattern characterized by positive spikes on a logarithmic scale (Fig. 11a and b). Typical are elevations of Ba relative to K and Sr in both the anorthosites and the mafic dykes. The anorthosites show marked spikes at the incompatible elements Ba, K, Sr (spikes in these elements correlate with LREE-enrichment) and Hf and elevation in La–Ce relative to Nb. The elements Th and U in contrast show troughs. The opposing distribution of Y at the end of the trace element diagram indicates the presence of accessory phases like Grt, Sp or Ap (in the samples MB81 and A157) that readily concentrate Y.

4.2.2. Uluguru Mountains

4.2.2.1. Major elements. The Uluguru anorthosite suite displays a wider range of rock types and is less homogeneous than the Pare Mountains suite. This is indicated in the AFM diagram (Fig. 8b), where a trend from anorthositic to more metagabbroic samples is seen. In contrast to the Pare anorthosite, the plotted samples scatter from an idealized linear trend (as in Fig. 8a) towards FeO and MgO. The range in composition is even wider for the mafic samples on the FeO–MgO-rich side of the diagram. The spatially related granitoid (TV02/46a) with mafic segregations (TV02/46c) is indicated separately. A broader spectrum of element distribution is also obvious in Fig. 9. The Ca content of the Uluguru samples ranges from 9 to 11 wt% and the Na2O content from 4 to 5.5 wt% (Fig. 9b and c). The FeO and MgO contents show an increase of FeO with decreasing Al2O3 values displaying an increase in mafic components of the rock. Two samples, MB81 and A157 exhibit a slightly negative anomaly in Tb. The REE distribution in the mafic dykes is about 10 times enriched relative to chondrites with a normalized La/Lu ratio of ca. 1.7–2.5 (Fig. 10b). Eu shows a small positive Eu-anomaly caused by the abundance of plagioclase. Trace element concentrations are normalized to primordial mantle after Taylor and McLennan (1985) and shown in Fig. 11a and b. The anorthositic samples exhibit a relatively uniform pattern characterized by positive spikes on a logarithmic scale (Fig. 11a and b). Typical are elevations of Ba relative to K and Sr in both the anorthosites and the mafic dykes. The anorthosites show marked spikes at the incompatible elements Ba, K, Sr (spikes in these elements correlate with LREE-enrichment) and Hf and elevation in La–Ce relative to Nb. The elements Th and U in contrast show troughs. The opposing distribution of Y at the end of the trace element diagram indicates the presence of accessory phases like Grt, Sp or Ap (in the samples MB81 and A157) that readily concentrate Y.

4.2.2.2. REE and trace element patterns. The anorthositic samples show the typical plagioclase pattern with LREE-enrichment continuously decreasing to HREE depletion with the positive Eu spike similarly to the Pare anorthosites. However, the total abundance of REE is generally about one order of magnitude higher in the Uluguru anorthosites than in the Pare ones (with the exception TV02/10). In contrast to the Pare samples, the Uluguru anorthosites do not exhibit spiky patterns of the HREE and no upward trend of Lu can be observed. The metagabbroic samples are generally more enriched in REE than the anorthosites and the positive Eu-anomaly is smaller. Sample TV02/10b is an exception with a steeper REE pattern and a less pronounced Eu-anomaly. This is caused by the marginal position of the sample in the shear zone and it is therefore most likely contaminated by the surrounding migmatitic basement gneisses.
Fig. 9. (a–h) Major elements are plotted against the Al₂O₃ content for the rocks from the anorthosite suites and the Uluguru basement. The Uluguru samples are indicated as diamonds and the Pare samples are represented by half-filled circles. The mafic rocks from the Mahenge Mountains are represented as boxes (filled and half-filled). The long sample names have been changed to shorter reference numbers (as shown in Table 5). For comparison, an average geochemistry of the Adirondacks anorthosites is added as asterisk and abbreviated with ‘A’ (taken from Sampson and Wright, 1964).
The pattern of the basement gneiss (TV02/45) shows continuously decreasing REE contents from light to heavy without any Eu-anomaly (Fig. 10d). Other samples on this diagram are the associated mafic and felsic rocks from the anorthosite suite itself and contain slight to more pronounced positive Eu-anomalies depending on the amount of plagioclase. The REE-enrichment is variable from about 2 to 100× chondrites.

The variation in rock types is also documented in the trace element patterns (Fig. 11c). Similar to the Pare anorthosites the patterns are very spiky with troughs in Th–U and Nb–Ta. Positive spikes occur at Ba, K, Sr. The two Ap-rich samples (TV02/47b, TV02/06) show a positive spike in P and high values in Sm and Ti. Sample TV02/10 with the lowest trace element concentrations compared to primordial mantle has a positive spike in Hf similar to the Pare samples. The other samples, however, show a small negative spike of Hf. La–Ce values are generally elevated with higher abundances in the metagabbroic samples than in the anorthosites. The associated rocks are plotted in Fig. 11d. The patterns with the highest trace element concentrations are from the granitoid sample TV02/46a (up to 1000× primordial mantle in Rb), the lowest con-
mantle (after Taylor and McLennan, 1985). The samples numbers given in the diagrams are in the order of the trace element patterns. (a) Spider Fig. 11. Trace element spider diagrams for the discussed samples of the anorthosites and associated rocks. Normalization is against a primordial evolution during the magmatic history. In Fig. 11e, a comparison between an average lower crustal rock (Taylor and McLennan, 1985) and the basement gneiss TV02/45 is shown. Both patterns correlate well and therefore it can be concluded that the migmatitic basement gneisses are of lower crustal origin.

4.2.3. Mahenge Mountains

4.2.3.1. Major elements. Three meta-tonalitic samples of the Mahenge Mountains have Al₂O₃ contents that are lower than the two other anorthosites and range between 13 and 15 wt% and much higher in SiO₂ ranging between 64 and 76 wt% (Fig. 9a). The exception is sample T96, which is a Kfs-bearing monzogranitic rock, here referred to as meta-granitoid. The chemical difference of this rock type on the diagrams of Fig. 9 is mainly seen on the distribution of SiO₂, CaO and Ba (Fig. 9a, b and g). The Mahenge samples do not fall along the linear differentiation trend of the anorthosites and related mafic rocks, but plot apart on the Al₂O₃ poorer part of the diagrams.

4.2.3.2. REE and trace element patterns. The REE distribution of the Mahenge samples demonstrates the marked compositional difference of these rocks from those of the anorthosites of the northern outcrops (Fig. 10e). The meta-tonalitic samples generally show a flat three to six times enriched REE pattern with the exception of the positive Eu-anomaly. The chemical difference of this rock type on the diagrams of Fig. 9 is mainly seen on the distribution of SiO₂, CaO and Ba (Fig. 9a, b and g). The Mahenge samples do not fall along the linear differentiation trend of the anorthosites and related mafic rocks, but plot apart on the Al₂O₃ poorer part of the diagrams.

4.3. Summary of geochemical data

Proterozoic and Archean anorthosites show a very similar pattern of trace and REE elements; however, the REE concentrations are generally one order of magnitude higher than the Archean anorthosites. All three considered complexes of the Eastern Granulites show different geochemical signatures governed by a complex history of differentiation processes. The geochemistry of the Pare Mountains fits into the system of the Archean anorthosites with its low REE abundances and higher Ca content than the Uluguru Mountains. The Pare anorthosites show uniform chemical systematics with a rise in Lu and Hf. The Uluguru anorthosites are less homogeneous with generally higher but variable concentrations in REE and trace elements and do not show this Hf peak with one exception (TV02/10). A comparable geochemical feature of all three occurrences (also the Mahenge Mountains) is, that the associated mafic and felsic rocks experienced a different degree of differentiation assuming a similar metamorphic grade of high-pressure granulite facies overprinting the whole study area of the Eastern Granulites.

Enrichments in normalized trace element patterns are mostly a consequence of crustal components added to a mantle source. This is visible in almost all samples in the elements Ba, Sr (occasionally K, P, Nd, Sm). Crustal
contamination is a common process in destructive plate margin settings (continental or island arc magmatism). The tectonic setting of the anorthosites however cannot clearly be inferred from geochemistry and remains a matter of debate (Ashwal, 1993).

5. Interpretation

The anorthosite from the Pare Mountains preserves an Archean age of 2643±16 Ma. This is consistent with the geochemical and mineral chemical analyses of several samples in spite of many of the magmatic features of the rocks being obliterated by Neoproterozoic granulite facies metamorphism. It cannot be excluded that those Archean ages are from xenocrysts although there is no evidence for another magmatic zircon type in the dated sample representing a second crystallization event in the Proterozoic. In contrast zircons from the Uluguru anorthosite suite do not preserve any evidence for Archean formation ages. By comparing the internal structure of the zircon grains with the age spectra of zircons from the anorthosites, associated granitoids and basement gneisses, we infer two crystallization events prior to Neoproterozoic metamorphism. Ages around 986±16 Ma are found in zircon cores of the basement gneiss. A subsequent magmatic event between 880 and 820 Ma produced the anorthosites. Similar age clusters and Zrn morphologies can be found in the magmatic rocks of the Mahenge Mountains. Broad metamorphic rims around the igneous zircons (most pronounced in the Mahenge Mountains) allowed a determination of the Neoproterozoic metamorphic overprint. Ages around 850–880 Ma have been occasionally found in the East African Orogen but these ages are not very well-known yet in Tanzania and seemingly restricted to the Eastern Granulites (Table 2).

A magmatic event around 820 Ma has already been widely discussed in the recent literature in other localities, e.g. Madagascar (Kröner et al., 2000; Kröner, 2001). Some authors (Hoffman, 1991, 1999) argued that a variety of processes can be responsible for the formation of vast melt generations during this time span that is associated with the dispersal of Rodinia. Meert’s (2003) synopsis of the events interpreted the time span between 800 and 700 Ma in Kenya–Tanzania as a time of long-lasting arc magmatism accretion. Extensive Andean-type arc magmatism at ca. 790 Ma is reported from Madagascar (Handke et al., 1999; Kröner et al., 2000). The Nd isotope systematics as well as abundant zircon xenocrysts attest to extensive remelting of Archean and Paleoproterozoic crust.

Appel (1996) inferred from comparable basement gneisses an active continental margin as most likely tectonic scenario for this melt formation. Subsequently, a considerable amount of lower crust must have crystalized in the time span around 880–820 Ma in the Eastern Granulites represented at present as the anorthosites. Collins and Pisarevsky (2005) interpreted that Azania (Central Madagascar) rifted off East Africa at this time. This may have resulted in the anorthosite magmatism described here.

6. Conclusions

(1) The anorthosites of the Pare Mountains and the Uluguru Mountains are structurally similar, oval-shaped to pear-shaped bodies that are tectonically incorporated in meta-magmatic high-grade granulites. Both occurrences have been intensely affected by Neoproterozoic metamorphism and deformation. The boundary of the anorthosites to the country rock is in both places defined by major sinistral strike-slip zones that developed at granulites facies conditions.

(2) Characteristic metamorphic mineral assemblages are elongated lenses of Cpx and Prg surrounded by a Grt corona within the Pl fabric in both bodies. The Pl chemistry of the Pare anorthosite is An-dominated with $X_{An} = 0.6–0.8$ and the Uluguru anorthosite has an intermediate chemistry of $X_{An} = 0.4–0.6$.

(3) A remarkable difference in the formation ages of the anorthosites is prevalent in the Uluguru and the Pare Mountains. The anorthosite from the Pare Mountains has Archean formation ages around 2.64 Ga and the Uluguru occurrence yield igneous ages of 880–820 Ma with a well-defined overprint due to Neoproterozoic granulite facies metamorphism. The migmatitic basement gneiss has formed around 986 Ma most likely in an island arc tectonic setting, but age dating of metamorphic rims around magmatic Zrn cores shows Neoproterozoic metamorphism at 640–650 Ma.

(4) The geochemistry reflects a difference between the more calcic Pare anorthosite and the less calcic Uluguru anorthosite. The abundance levels of REE compared to a chondrite is about one order of magnitude higher in the Uluguru Mountains anorthosite suite than in the Pare anorthosite. Another significant feature is the elevated amount of crustal contamination that affected the Uluguru samples compared to the Pare samples.

(5) The tectonic setting in which the magmatic bodies intruded within the period from 880 Ma onwards as
represented by Zrn formation in the Uluguru samples is most likely to be magmatic. An island-arc collisional setting is realistic as the geochemical signature of the surrounding granulites which formed at 986 Ma points to an active continental margin. The time difference between 986 and 880 Ma could imply an extensive phase following arc accretion either in a back-arc basin or in a continental rift, e.g. when Madagascar rifted off East Africa. However, there are no other rocks dated yet in Tanzania that might be related to rift associated magmatism in that time span.

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