Zircon ages and Nd–Hf isotopes in UHT granulites of the Ider Complex: A cratonic terrane within the Central Asian Orogenic Belt in NW Mongolia

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1. Introduction

The broad terrane of southern Siberia and Mongolia south of the Siberian craton is part of the Central Asian Orogen Belt (CAOB). It consists of Neoproterozoic to early Paleozoic island arc and ophiolite complexes as well as fragments of continental massifs that were tectonically juxtaposed during accretionary orogenic events in the Paleo- zoic (Mossakovsky et al., 1993; Kovach et al., 2004; Kovalenko et al., 2005; Kozakov et al., 2008, 2011; Yarmolyuk et al., 2011; Kröner et al., 2014). The oldest of these continental massifs consists of crystalline rocks dating back to the Archaean and may be rifting-off fragments of ancient cratons. Single zircon ages as well as whole-rock Sm–Nd and Pb–Pb isotopic data indicate that Archaean to Paleoproterozoic rocks only occur in the Baidarik Block, the Ider Complex of the Tarbagatai Block all in northwestern Mongolia, (Kozakov et al., 2007, 2011), and in the Gargan Block of East Sayan in southern Siberia (Kovach et al., 2004; Anisimova et al., 2009a) (Fig. 1).

The zircon age of tonalitic gneisses in the Gargan Block is 2727 ± 6 Ma (Anisimova et al., 2009a) and the age of high-temperature metamorphism is 2664 ± 15 Ma (Kovach et al., 2004), but the P–T conditions of metamorphism are not well defined. In contrast, the oldest zircon age of a granite-gneiss of the Tarbagatai Block is 2219 ± 25 Ma (Kozakov et al., 2011).

The most detailed geochronological information is available for the Baidarik Block where the oldest zircon age is 2646 ± 45 Ma for tonalitic gneisses of the Baidrag Complex (Mitrofanov et al., 1981). Detailed SHRIMP dating has shown that the oldest components are 207Pb/206Pb core ages at around 2.8 Ga, and the main group of zircons yielded 207Pb/206Pb ages of 2.50–2.65 Ga, including grains as typically found in granulite-facies rocks (Kozakov et al., 2007). Mapping and zircon dating of granitoid gneisses have also documented a major tec tono-metamorphic event at 1854–1825 Ma and an earlier granulite-facies metamorphism in the Bumberg Complex of the Baidarik Block between 2364 and 2308 Ma (Kozakov et al., 2007; Demoux et al., 2009).

Thus, only two localities of early Precambrian basement are currently known in the northern CAOB. Here we present new age data and Hf-in-zircon isotopic compositions for granulite-facies rocks in the early Precambrian Ider Complex of the Tarbagatai Block and discuss their sources and tectonic setting.

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2. Geology of the Tarbagatai Block

Crystalline rocks of the Tarbagatai Block were first considered to be the part of the Dzabkhan microcontinent (Moskakovsky et al., 1993) and were later defined as the Tarbagatai cratonic terrane (Badarch et al., 2002). However, they constitute several large but separate blocks enclosed in the Permian Hangay batholith and are structurally isolated from the Dzabkhan microcontinent and the Baidarik block (Fig. 2). Amongst the Tarbagatai crystalline rocks an early Precambrian and a Neoproterozoic domain were recognized and named the Ider and Jargalant Complexes, respectively (Fig. 3). Originally, high-temperature regional metamorphism was the criterion for considering the Ider Complex to be early Precambrian in age.

The Ider Complex consists of strongly migmatized biotite and hornblende orthogneisses with rare inclusions of amphibolite and granite-gneiss bodies; sillimanite-bearing paragneisses occur in subordinate quantities. An early high-grade metamorphism produced granulite-facies assemblages as indicated by inclusions of two-pyroxene-bearing gneiss, enderbite and charnockite (Table 1). The central parts of such inclusions remained unaffected by amphibolite facies retrogressive (regional) metamorphism, and relict parageneses are: clinopyroxene ± brown hornblende + plagioclase ± quartz and clinopyroxene + plagioclase + orthoclase + quartz. Charnockite formation is documented by the paragenesis orthopyroxene + plagioclase + orthoclase + quartz, and enderbites contain the paragenesis orthopyroxene + plagioclase with antiperthite + quartz + biotite. Pyroxene is replaced by green hornblende in the marginal parts of such granulite inclusions. In general, as a result of amphibolite-facies retrograde regional metamorphism, many pyroxene-bearing gneisses were downgraded into amphibolite gneisses or schists, and dark colored minerals such as biotite and/or green hornblende are present in most migmatites.

Retrograde metamorphism of the granulite-facies assemblages was accompanied by widespread migmatization, culminating in the injection of numerous granitoid veins with a width of a few tens of centimeters to 1.5–2 m. These granitoids cut across the migmatite layering, and in areas of high strain they are folded together with the migmatites and these domains become well foliated, gneissose granite. The zircon age of one of these granite-gneisses is 2219 ± 25 Ma (Kozakov et al., 2011). Later, more massive, granitoids also occur in the migmatites and make up irregularly-shaped bodies with indistinct contacts over an area of more than several hundred square meters. The established zircon age range for these granitoids is about 1870–1860 Ma (Kozakov et al., 2011).

Generation of the main foliation and structures of the Ider Complex and amphibolite-facies regional metamorphism occurred prior to
3. Field relations and metamorphic conditions of granulite sample 7340

Samples 7340 and 7011 are from the same outcrop of migmatized biotite gneiss and green hornblende gneiss of the Ider Complex which contain rare relics of granulite-facies rocks such as clinopyroxene and orthopyroxene gneisses, enderbite, and charnockite in amphibolite-facies gneiss and migmatite.

<table>
<thead>
<tr>
<th>No.</th>
<th>Sample no.</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
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<tbody>
<tr>
<td>1</td>
<td>6657</td>
<td>6658</td>
<td>7340</td>
<td>7011</td>
<td>6151</td>
<td>6152</td>
<td>6153</td>
<td></td>
</tr>
</tbody>
</table>

### Table 1: Major and trace element composition of gneisses from the Ider Complex.

<table>
<thead>
<tr>
<th>Element</th>
<th>SiO₂</th>
<th>MgO</th>
<th>Al₂O₃</th>
<th>Fe₂O₃</th>
<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sample 1</td>
<td>64.30</td>
<td>16.38</td>
<td>24.62</td>
<td>25.87</td>
<td>1.99</td>
<td>0.11</td>
<td>0.03</td>
</tr>
<tr>
<td>Sample 2</td>
<td>65.94</td>
<td>17.89</td>
<td>24.62</td>
<td>25.87</td>
<td>1.99</td>
<td>0.11</td>
<td>0.03</td>
</tr>
</tbody>
</table>

Enderbite sample 7340 represents one of these relics and was collected from a small granulite-facies inclusion with gradational contact within an amphibolite-facies gneiss. Granulite-facies metamorphic mineral assemblages are found in the central parts of these bodies, whereas in the marginal parts orthopyroxene is replaced by hornblende. Charnockite sample 7011 was collected from a thin (25–30 cm) layered body with a visible length of about 3 m and is concordant with the enclosing rocks.

Enderbite sample 7340 mainly consists of plagioclase, quartz and orthoclase rims and orthoclase-bronzite and orthoclase-magnetite. Enderbite sample 7340 represents one of these relicts and was collected from a small granulite-facies inclusion with gradational contact within an amphibolite-facies gneiss. Granulite-facies metamorphic mineral assemblages are found in the central parts of these bodies, whereas in the marginal parts orthopyroxene is replaced by hornblende. Charnockite sample 7011 was collected from a thin (25–30 cm) layered body with a visible length of about 3 m and is concordant with the amphibolite-facies regional foliation (defined by green hornblende) in the enclosing rocks.

Enderbite sample 7340 mainly consists of plagioclase, quartz and orthopyroxene but also contains small biotite flakes and orthoclase rims around biotite and orthopyroxene (Fig. 4). Plagioclase contains orthoclase inclusions (antiperthite), and some biotite flakes are found as ingrowths in orthopyroxene.

The temperature of granulite-facies metamorphism was established from the exchange (Fe/Mg) reaction between orthopyroxene and...
4. Geochemical characteristic of high-grade rocks in the Ider Complex

Granulites of Ider Complex can be subdivided into two groups according to their composition and chemical features. The first group (samples 6657, 6658 and 7340 in Table 1) correspond to metaluminous quartz diorite and granodiorite with Na2O/K2O = 1.9–3.5. They are characterized by fractionated trace-elements distribution patterns with negative anomalies in Th, U, Nb, Ta, P and Ti (Fig. 5a), low Sr/Y ratios (15–45), moderately fractionated REE patterns (La/Yb = 7.9–17.8) with gentle slopes of HREE (Gd/Yb = 2.1–2.8), and weak negative and positive Eu anomalies (Eu/Eu* = 0.74–1.12) (Fig. 5b). The geochemical data are compatible with the interpretation that the enderbite protoliths may have formed by partial melting of basaltic source in an island arc or active continental margin setting at a pressure of ~3–7 kbar with Pl + Cpx + Opx in the residue.

In contrast to the enderbites, charnockite sample 7011 is characterized by high silica and chemically corresponds to leucogranite or quartz monzonite (Table 1). It has a trace element pattern with pronounced negative anomalies in Th, U, Nb-Ta, P and Ti, and positive anomalies in Sr, Zr–Hf and Eu (Fig. 5a, b), a highly fractionated REE pattern (La/Yb = 110) with low HREE contents (Yb = 0.036 ppm), and a strong positive Eu anomaly (Eu/Eu* = 11.0). These features suggest derivation of the charnockite protolith through anatetic melting of country rocks during high-grade metamorphism.

The migmatized biotite gneisses (samples 6151, 6152 in Table 1) resemble rhyodacite and rhyolite (granodiorite and granite) in major element composition (Table 1). They have trace-element patterns enriched in Th and strongly depleted in HREE relative to the enderbites (Fig. 5a, b) as well as strongly fractionated REE patterns (La/Yb = 81–106) with weak Eu anomalies (Eu/Eu* = 1.06–1.14). The chemical characteristics of the migmatized biotite gneisses require hornblende in the residue during derivation of their protoliths, implying higher pressures and, consequently, different sources than suggested for protoliths of the enderbites.

The biotite granite-gneiss (sample 6153), cutting the migmatites, has a granodioritic composition and geochemical characteristics similar to those of the migmatites but differs from these in its positive Sr anomaly, an absence of a Zr–Hf anomaly, and lower trace elements content (Fig. 5a, b).

5. Previous geochronology

U–Pb ID-TIMS dating of zircons from sample 6153 was performed in the Vernadsky Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences for three multigrain zircon aliquots, with two fractions subjected to selective dissolution for 4 and 8 h, respectively. Analyses corresponding to the isotopic composition of the dated zircon fractions define a discordia line with an upper Concordia intercept at 2219 ± 25 Ma and a lower intercept at 535 ± 570 Ma, MSWD = 2.8. This age was considered to reflect the time of zircon crystallization and was interpreted as the upper age boundary for the Ider Complex (Kozakov et al., 2011). However, because of the multigrain dating method and no CL control, it is possible that the above discordia line reflects mixing between magmatic (or inherited) and metamorphic zircons. Therefore, the age of amphibolite-facies retrograde metamorphism in the Ider Complex remains uncertain.

6. Results and discussion

6.1. Zircon morphology, cathodoluminescence images, and U–Pb ages

Zircons from charnockite and enderbite samples 7011 and 7340 were analyzed on SHRIMP II of the Beijing SHRIMP Centre, and the analytical procedures are detailed in Appendix A. The analytical data are listed in Table 2.

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Zircons from both samples are colorless and transparent, as is common in granulite-facies terrains (e.g., Kröner et al., 1994; Corfu et al., 2003). Both samples contain two morphologically distinct zircon generations. The more numerous grains are predominantly long-prismatic in granulite-facies terrains (e.g., Kröner et al., 1994; Corfu et al., 2003). Both samples contain two morphologically distinct zircon generations. The more numerous grains are predominantly long-prismatic, and have been described from many metamorphic terranes. The predominantly morphism, and has been described from many metamorphic terranes.

Partial zircon dissolution and recrystallization during high-grade metamorphism with similar zircons from other high-grade terranes, we interpret as the result of partial recrystallization (e.g., Fig. 6d), metamictization and other processes related to modifications as a result of interaction with a fluid phase (e.g., Wan et al., 2009). There are also rare low-U cores of magmatic origin and occasionally there occurs a complex intergrowth of low- and high-U phases (Fig. 6f). We avoided such zones during analysis and selected domains on the magmatic grains that indicated undisturbed growth.

The magmatic zircons in sample 7340 are generally less rounded at their terminations than those in 7011 and have simpler-looking internal structures (Fig. 6f), but many have very high thinly luminescent rims, too narrow to be analyzed on SHRIMP and most likely of metamorphic origin.

The second group of zircons in both samples is oval-shaped to ballround, often multifaceted and very distinct because of their light gray highly luminescent CL images indicating low U-contents (Fig. 6g–i). These grains are undoubtedly of metamorphic origin. They often display typical fine-tree or sector zoning (Fig. 6g–h) due to slow diffusive growth during high-grade metamorphism under anhydrous conditions (e.g., Kröner et al., 2000; Wan et al., 2009; Dong et al., 2014). Some grains have older igneous-looking cores, in some instances just tiny domains (Fig. 7i), and some well-rounded grains, originally considered to be metamorphic turned out to be igneous in origin under CL (Fig. 6j). In general, the metamorphic zircons appear to have simpler internal structures than the igneous grains.

The heterogeneity of most zircons of igneous origin, as revealed by their CL images, clearly indicates that whole-grain isotopic analysis will not produce meaningful ages because several components are completely intergrown in almost each grain and Pb-loss and/or recrystallization have most likely disturbed the original U–Pb isotopic ratios. A good example of such isotopic disturbance due to high-grade metamorphism is shown by Wan et al. (2011). In the ideal case of only two components, one magmatic and one metamorphic, the U–Pb isotopic ratios should be in antithetic Pb; 8 — lamellae in antithetic Pb.

### Table 2

Major element composition of minerals from enderbite sample 7340 of the Ider Complex, northwestern Mongolia.

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Opx</th>
<th>Opx</th>
<th>Opx</th>
<th>Opx</th>
<th>Opx</th>
<th>Opx</th>
<th>Bt</th>
<th>Bt</th>
<th>Bt</th>
<th>Bt</th>
<th>Bt</th>
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</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>82.86</td>
<td>63.25</td>
<td>63.13</td>
<td>62.42</td>
<td>62.67</td>
<td>63.10</td>
<td>58.43</td>
<td>58.41</td>
<td>58.41</td>
<td>58.31</td>
<td>58.66</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.66</td>
<td>0.59</td>
<td>0.50</td>
<td>0.65</td>
<td>0.44</td>
<td>0.43</td>
<td>0.75</td>
<td>0.11</td>
<td>0.08</td>
<td>0.10</td>
<td>0.03</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>0.21</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
</tr>
<tr>
<td>MgO</td>
<td>0.21</td>
<td>0.14</td>
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<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
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<td>0.14</td>
<td>0.14</td>
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</tr>
<tr>
<td>CaO</td>
<td>0.21</td>
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<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.21</td>
<td>0.14</td>
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<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.21</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
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<tr>
<td>Total</td>
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<td>99.99</td>
<td>99.99</td>
<td>100.01</td>
<td>100.00</td>
<td>100.00</td>
<td>99.99</td>
<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
</tr>
</tbody>
</table>

Notes: 5 — mantled with Or; 2 — mica flake; 3 — ingrowth in Opx; 4 — secondary, in Opx grain; 5 — rim around a Bt grain; 6 — rim around an Opx grain; 7 — inclusions in antithetic Pl; 8 — lamellae in antithetic Pl.

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concordant results and two are significantly discordant (Fig. 7a), but all have similar $^{207}\text{Pb}/^{206}\text{Pb}$ isotopic ratios (Table 2) resulting in a mean age of $1855 \pm 2$ Ma (Fig. 7a). Three cores in the igneous zircons are much older but are strongly discordant and precise ages can therefore not be calculated. However, they are well aligned in the Concordia diagram and, as in their younger hosts, have near-identical $^{207}\text{Pb}/^{206}\text{Pb}$ ratios corresponding to a mean age of $2522 \pm 2$ Ma (Fig. 7a, inset). These cores most likely reflect the crustal source from which the charnockite protolith of sample 7011 was derived.

Five metamorphic zircons of sample 7011 produced concordant and near-concordant results (Table 2) with a mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of $1857 \pm 3$ Ma (Fig. 7b) that we interpret to reflect the time of peak granulite-facies metamorphism. This age is identical, within error, to the age obtained on the magmatic grains, and there are two possibilities to interpret these results. First, and most likely, the granitic charnockite protolith crystallized at about $1855$ Ma ago and resulted from melting of a late Archaean crustal source. High-grade metamorphism almost immediately followed on pluton emplacement and transformed the granite into a charnockite. Alternatively, all magmatic-looking zircons in this sample recrystallized completely during high-grade metamorphism, and their original isotopic system was totally reset. We consider the second option unlikely because some of the dated grains clearly preserve igneous oscillatory zoning and show no evidence in their CL-images of recrystallization.

Ten magmatic-looking zircons from sample 7340 yielded variably discordant but well aligned data with one concordant analysis, and the resulting mean $^{207}\text{Pb}/^{206}\text{Pb}$ age is $2542 \pm 1$ Ma (Table 2, Fig. 8a). We interpret this to reflect late Archaean crystallization of the enderbite protolith. Nine metamorphic zircons from the same sample have concordant, slightly discordant, and significantly discordant results but, as in sample 7011, their $^{207}\text{Pb}/^{206}\text{Pb}$ ratios are almost identical (Table 2) and result in a mean age of $1855 \pm 3$ Ma (Fig. 8b) which most likely reflects the peak of high-grade metamorphism. Two igneous cores in these zircons yielded discordant results suggesting a late Archaean minimum age of $2546 \pm 1$ Ma. This is identical to the magmatic age of sample 7340, as would be expected.

Comparing the two results it appears that sample 7340 represents a late Archaean basement that underwent late Paleoproterozoic granulite-facies metamorphism, whereas sample 7011 may reflect crustal melting of a late Archaean protolith in the late Paleoproterozoic, possibly the same basement as represented by sample 7340, immediately followed by high-grade metamorphism. Both samples experienced the same granulite-facies metamorphic event as recorded in their respective metamorphic zircons.

**Fig. 5.** (a) Primitive mantle-normalized trace-element variation diagram for rocks of the Ider Complex listed in Table 1. (b) Chondrite-normalized REE distribution patterns for rocks listed in Table 1. Normalizing values are from Sun and McDonough (1989) and Taylor and McLennan (1985), respectively.
6.2. Nd whole rock isotopic systematics

The Sm–Nd isotopic systematics of felsic igneous rocks are widely used to constrain their sources and the approximate time of crust formation from the depleted mantle (DePaolo, 1988). A sample is characterized by the \( \epsilon_{\text{Nd}}(t) \) parameter which denotes the initial Nd ratio at the time of rock formation. It is usually accepted that a positive \( \epsilon_{\text{Nd}}(t) \) value indicates a short-lived juvenile source, whereas a negative \( \epsilon_{\text{Nd}}(t) \) value is indicative of rock formation due to a reworking of a long-lived crustal source. The time of formation of a rock from a mantle source is calculated as the Nd model age \( t_{\text{Nd(DM)}} \) that corresponds to the time when the Nd isotopic composition of this rock was identical to its mantle source. However, since felsic igneous rocks are often formed from sources with different isotopic compositions and ages, a Nd model age is usually interpreted as average crustal residence age and reflects the average age of the material involved in the formation of the rock.

Fig. 6. Cathodoluminescence images of zircons analyzed in this study. (a–e) Magmatic grains from charnockite sample 7011. (f) Magmatic grain from charnockite sample 7340. (g–h) Metamorphic zircons from sample 7011. (i) Metamorphic grain from sample 7340. (j) Well-rounded, translucent grain of seemingly metamorphic origin from sample 7340 but, in fact, igneous as shown by CL image and age. For explanation of the various features see text. SHRIMP analytical spots and numbers refer to Table 3.
In spite of the generally immobile behavior of Sm and Nd during intracrustal processes (Taylor and McLennan, 1985), there are cases where anatexis of crustal rocks or fractionation of accessory minerals significantly changes the Sm/Nd ratio and thus leads to unrealistic tNd(DM) estimates (e.g., Harris et al., 1996; Kovalenko et al., 1999). In order to account for such possible Sm–Nd fractionation, a so-called two-stage (or "crustal" by analogy with Hf-in-zircon isotope systematics; see below) Nd model age tNd(c) (Keto and Jacobsen, 1987) can be calculated. This crustal Nd model age assumes that the average crustal Sm/Nd ratio for a potential source evolved from the time of depleted mantle melting to the age of rock crystallization. We use such crustal model ages below.

Nd whole-rock isotopic data for whole-rock samples of the Ider Complex are summarized in Table 4 and compared with Nd isotopic data for gneisses and Paleoproterozoic granitoids of the Baidarik Block in Fig. 9. Enderbites of the Ider Complex have slightly negative and close to zero εNd(t) values varying from −1.8 to +0.4 and corresponding Mesoarchean Nd depleted mantle model ages tNd(DM) of 3.1–2.9 Ga. The Nd isotopic and geochemical data suggest derivation of the enderbite precursors from melting of late Archaean juvenile as well as older crustal sources, probably in an active continental margin setting.

The migmatized biotite gneisses yielded higher positive εNd(t) values of +1.5 to +1.6, assuming a rock formation age of 2540 Ma. The remaining migmatized gneisses as well as a granite-gneiss and charnockite have low 147Sm/144Nd ratios of 0.0561–0.0733, much lower than average crustal values of 0.09–0.13, and this may have been caused by migmatization and anatexis. The corresponding Nd crustal model ages tNd(c) of 2.8–2.5 Ga are strongly dependent on the assumed crystallization age. The Nd crustal model ages tNd(c) of 2.8 Ga for the migmatized gneisses are close to depleted mantle model ages tNd(DM) for the enderbites, suggesting that both crustal reworking and input of some juvenile material occurred during generation of the migmatite precursors.

The granite-gneiss and charnockite are characterized by negative εNd(t) values of −4.5 and −7.6, corresponding to Mesoarchean crustal
Nd model ages $t_{Nd(c)}$ of 3.0 Ga. These rocks are similar in their Nd isotopic evolution to enderbites and migmatized gneisses of the Ider Complex (Fig. 9). The Nd isotope and geochemical data imply derivation of the charnockite and granite-gneiss precursors through anatexis of older crustal rocks.

Rocks of the Ider Complex have Nd model ages that partly overlap with, but are generally younger than, 3.3–2.9 Ga model ages for metamorphic rocks of the Baidarik Block (Fig. 9). This is probably related to the addition of juvenile material at ca. 2540 Ma.

6.3. Hf isotopic systematics

Hf isotopic ratios were obtained on previously dated SHRIMP spots and additional domains on zircons from samples 7011 and 7340, using a Nu Plasma HR MC-ICP-MS equipped with a laser ablation system at the University of Hong Kong (for analytical details see Appendix A). The results are summarized in Table 5.

The isotopic ratios in all zircons are remarkably similar, attesting to the robustness of the Lu–Hf system, even in cases of significant Pb-loss in the zircons during metamorphism. It is possible to calculate an approximate age for the source rock from which the zircons were derived, namely a Hf crustal model age ($t_{Hf(c)}$), by forcing a growth curve for a system with a Lu/Hf isotopic ratio corresponding to the average crust through the zircon initial ratio. However, there is some uncertainty in the calculation of this model age because the whole-rock $^{176}$Lu/$^{177}$Hf ratio of the protolith is not known, and therefore most authors use an assumed value for the continental crust, varying between 0.009 and 0.015 (e.g., Vervoort and Patchett, 1996; Belousova et al., 2006). Whole-rock Lu–Hf isotopic data from a wide variety of high-grade granitoid gneisses suggest that a ratio of 0.010 appears to be most suitable and also best matches whole-rock Nd model ages (Kröner et al., 2013a, 2014). However, we caution against taking these values too literally because the real Lu/Hf protolith ratios are not known, and there is also some uncertainty on the degree of depletion of the average.

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Archaean mantle. Thus the error in the Hf crustal model ages cited below may be considerable.

The Hf initial ratios for six of the 1855 Ma zircons from sample 7011, expressed in εHf(t)-units, are between −9.0 and −5.7 (Table 5, Fig. 10a). Four analyses were made on zircon cores of which three had a measured mean age of 2522 Ma, and we assume that the remaining core has approximately the same age. The εHf(t)-values for these old cores range between −3.6 and −0.3 (Table 4). These data are graphically shown in Fig. 10a where the Hf initial ratios at the time of rock formation (1855 Ma for the magmatic grains, 2522 Ma for the inherited cores, see Table 5) are plotted against time. The red broken lines denote the evolution of the Hf isotopic ratios for the highest and lowest εHf(t)-values, assuming a 176Lu/177Hf ratio of 0.01. For the magmatic zircons of sample 7011 the resulting tHf, crustal model ages vary between 2.7 and 2.9 Ga, whereas the Neoarchean cores have model ages of 2.9–3.1 Ga (Table 5, Fig. 10a). The latter are just slightly older than those for the Paleoproterozoic magmatic grains and support the interpretation that both the charnockite precursor of sample 7011 and its possible igneous source represented by the zircon cores, were derived from Neo- to Mesoproterozoic crustal material.

Nine Lu–Hf analyses of metamorphic zircons from sample 7011, calculated for an age of 1857 Ma, yielded strongly negative εHf(t)-values ranging between −19.2 and −11.2 and corresponding tHf model ages of ca. 3.0–3.4 Ga (Table 5, Fig. 10b). These old model ages are close to those of the igneous zircons and suggest that the Lu–Hf isotopic system has been inherited from the magmatic source and was not significantly disturbed by the high-grade metamorphic event.

The Hf initial ratios for ten magmatic zircons from sample 7340, calculated for a crystallization age of 2542 Ma, vary widely between −5.0 and +2.7 and resulting in Hf crustal model ages of 2.8 to 3.2 Ga (Table 5, Table 6).
Fig. 9. εNd(t) versus age diagram for gneisses of the Ider Complex in comparison with gneisses and granitoids of the Baidarik Block.

Data from Kozakov et al. (1997) and Kozakov et al. (2011) and Table 3.

Fig. 10c), very similar to the zircon cores of charnockite sample 7011. This suggests that these cores probably represent the same crustal material as represented by sample 7340. Ten Lu–Hf analyses of metamorphic zircons from sample 7340 yielded extremely variable εHf(t)-values ranging from ~8.5 to ~5.4 if calculated for a metamorphic age of 1855 Ma. Two of these have very low values of ~15.5 and ~18.5, and we suspect that these grains contain inherited material of possible Archaean age not visible in CL images but shown in some of the metamorphic zircons of this population (see Fig. 7f). It is therefore possible that the laser beam cut deeply into these zircons and evaporated an Archaean or mixed phase. If we calculate crustal model ages for these two samples on the basis of a crystallization at 2542 Ma (Table 5) and combine these with the remaining εHf(t)-values for the metamorphic grains, the range is between ~10.0 and +0.1 (Fig. 10d), corresponding to tBLT model ages of 2.7 to 3.1 Ga which are again similar to those in the igneous zircons of the same sample and reiterate the suggestion that the Lu–Hf isotopic system of the metamorphic zircons is inherited from the magmatic enderbite protolith.

We suggest that this occurred in the following manner. Igneous zircon becomes “corroded” through the action of a fluid phase and partly dissolves during prograde high-temperature metamorphism. This can be seen in many zircons from high-grade terranes where the original pyramidal terminations of magmatic zircons are lost and become well rounded and “healed” as evidenced by multifaceted surfaces seen under high microscopic magnification (e.g., Kröner et al., 2013b). Such dissolution may also result in unusual zircon shapes (e.g., Fig. 7d), including ell-, or even ball-shaped grains (e.g., Fig. 7e) that are easily mistaken for detrital zircon. The dissolved material bearing the Hf isotopic composition of the “parent” igneous zircon is transported in the

Table 5

<table>
<thead>
<tr>
<th>Sample, spot</th>
<th>t (Ma)</th>
<th>ε176Yb/177Hf ±1</th>
<th>ε176Lu/177Hf ±1</th>
<th>ε176Hf/177Hf ±1</th>
<th>εHf(t) ±1</th>
<th>εNd ±1</th>
</tr>
</thead>
<tbody>
<tr>
<td>KOZ7011-1</td>
<td>1855</td>
<td>0.003403 ±0.000058</td>
<td>0.003157 ±0.000003</td>
<td>0.281238 ±0.000012</td>
<td>0.281363</td>
<td>-8.55</td>
</tr>
<tr>
<td>KOZ7011-2</td>
<td>2522</td>
<td>0.017506 ±0.000022</td>
<td>0.003855 ±0.000012</td>
<td>0.281173 ±0.000014</td>
<td>0.281132</td>
<td>-1.47</td>
</tr>
<tr>
<td>KOZ7011-3</td>
<td>1855</td>
<td>0.009297 ±0.000020</td>
<td>0.003159 ±0.000001</td>
<td>0.281355 ±0.000012</td>
<td>0.281350</td>
<td>-9.08</td>
</tr>
<tr>
<td>KOZ7011-4</td>
<td>1855</td>
<td>0.009308 ±0.000163</td>
<td>0.004665 ±0.000009</td>
<td>0.281397 ±0.000018</td>
<td>0.281381</td>
<td>-7.90</td>
</tr>
<tr>
<td>KOZ7011-5</td>
<td>1855</td>
<td>0.003215 ±0.000025</td>
<td>0.001500 ±0.000002</td>
<td>0.281449 ±0.000010</td>
<td>0.281444</td>
<td>-5.66</td>
</tr>
<tr>
<td>KOZ7011-6</td>
<td>1855</td>
<td>0.003151 ±0.000012</td>
<td>0.001486 ±0.000001</td>
<td>0.281406 ±0.000011</td>
<td>0.281401</td>
<td>-7.19</td>
</tr>
<tr>
<td>KOZ7011-6-2</td>
<td>2522</td>
<td>0.030158 ±0.000949</td>
<td>0.015645 ±0.000605</td>
<td>0.281150 ±0.000013</td>
<td>0.281075</td>
<td>-3.50</td>
</tr>
</tbody>
</table>

KoZ7011-6-1 is grain 6, spot 1, KoZ7011-6-2 is grain 6, spot 2, etc.; m denotes metamorphic zircon.
metamorphic fluid that, under suitable conditions, leads to the crystallization of new metamorphic zircon in the rock. This newly-formed zircon, characterized by a typical ball-round, multifaceted shape, broad zoning as revealed under CL, and a low Th/U ratio, as seen in our samples, then has a new U–Pb isotopic system reflecting the time of metamorphic crystallization, but it inherited the Hf isotopic composition from the fluid from which it nucleated. A similar process was envisaged by Zeh et al. (2010) for Hf isotopes in metamorphic zircon from amphibolite-facies schists in Antarctica and in eclogite-facies rocks from the Dabie orogen, China. Zeh et al. (2010) suggested that the dissolution/transportation/re-precipitation process resulted from a complex interplay between metamorphic dehydration reactions and symmetamorphic deformation processes.

7. Conclusions

The results of our study on high-grade rocks from the Ider Complex of northwestern Mongolia are summarized as follows:

1. The Ider Complex of the Tarbagatai Block in northwestern Mongolia contains high-grade granitoid gneisses with protolith ages of ca. 2550 Ma.
2. These late Archaean rocks (enderbites and two-pyroxene gneisses) have undergone UHT granulite-facies metamorphism at ca. 1850 Ma, followed by a retrogressive and amphibolite-facies overprint. Similar ages have been obtained for granulites of the Baidaragin Complex (protolith emplacement ages of 2.65–2.44 Ga, Kozakov et al., 2007) and intrusion of a Grt ± Opx bearing syntectonic granodiorite at 1854 ± 5 Ma (Kozakov, 1993).
3. Nd whole-rock isotopic systematics with Nd model ages ranging between ca. 2.5 and 3.1 Ga suggest that both crustal reworking and input of some juvenile material were involved in the generation of the Ider Complex gneiss protoliths. Hf-in-zircon isotopic data provide a similar pattern, and magmatic zircon also yielded Archaean Hf crustal model ages. The metamorphic zircon population seems to have inherited its Hf isotopic composition from the igneous grains, suggesting a complex process of dissolution, transportation, and re-precipitation involving a fluid phase during high-grade metamorphism.
4. Current data show that only three relatively small crustal blocks containing early Precambrian high-grade rocks occur in the eastern segment of the CAOB, namely the Baidrag, Ider and Gargan complexes. Other supposed microcontinents in the region are Neoproterozoic in age.
5. The zircon age patterns for the two samples of the Ider Complex do not make it possible to unambiguously assign the Tarbagatai Block to any of the cratons bordering the Central Asian Orogenic Belt, since age peaks at ca. 2520–2550 and ca. 1860 Ma are common in the Siberian, North China and Tarim cratons (Rojas-Agramonte et al., 2011).

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L.B. Terent’eva, Jianfeng Gao of the Geological Survey of Canada provided an unpublished Excel-based program to construct the Hf evolution diagrams of Fig. 10, and Figs. 1–3 were drawn by G.P. Pleskach. This research was supported by the Russian Foundation for Basic Research (RFBR grants 11-05-92003 and 14-05-00208) and is also part of a collaborative study between Mainz University and The University of Hong Kong, funded jointly by the Hong Kong Research Grants Council (RGC 704712P) and the DAAO/RGC Germany/Hong Kong Joint Research Scheme (G HK033/12 and PPP).

**Appendix A. Sample preparation and analytical procedures**

**Major element abundances** were determined using XRF methods in the Institute of the Earth’s Crust, Siberian Branch, Russian Academy of Sciences (RAS), Irkutsk. The detailed analytical procedures are described in Afonin et al. (1992). **Trace elements and REE concentrations** were determined on a VG Elemental PlasmaQuad PQ2 + “Turbo” inductively-coupled plasma mass spectrometer (ICP-MS) after lithium metaborate fusion. The analytical procedures were described by Panteeva et al. (2003). Standard reference materials were BHVO-2 (basalt), RGM-1 (ryolite), W-2 (diabase), JG-2 and G-2 (granites). The precision is better than 10%.

**Mineral analyses** from endebite sample 7340 were performed on a JEOL JSM-6510A electron microscope equipped with a JED-2200 analyzer at the Institute of Precambrian Geology and Geochronology, St. Petersburg, using a 20 kV accelerating voltage. Counting times for peak and background intensities were 30 s. Calibration was carried out with a set of artificial and natural mineral standards. Data were regressed using an oxide-ZAF correction program supplied by JEOL.

**A.1. SHRIMP zircon dating**

Heavy minerals were separated from whole-rock samples by standard procedures using a jaw crusher, a steel rolling mill and heavy liquids. Zircons were then hand-picked and mounted in epoxy resin together with chips of the standard M257 (Nasdala et al., 2008) for SHRIMP analyses. The mount was ground and polished so that the zircon cores were exposed, and zircons were photographed under cathodoluminescence (CL) to enable an easy and best location on the mount during SHRIMP analyses. CL imaging was done on a 305 Hitachi SEM S-3000N equipped with Gatan ChromaCL detector and a DigiSan II data306 recorder in the Beijing SHRIMP Center. Operating conditions were 9 kV, 99 μA.

Isotopic analyses were performed on the Beijing SHRIMP II ion microprobe whose instrumental characteristics are identical to that installed in Perth and described by De Laeter and Kennedy (1998). The analytical procedures are outlined in Compston et al. (1992), Claué-Long et al. (1995), Nelson (1997) and Williams (1998). Prior to each analysis, the surface of the analysis site was cleaned by the rastering of the primary beam for 3 min, to reduce or eliminate surface common Pb. The reduced 208Pb/206U ratios were normalized to 0.90101, which is equivalent to the adopted age of 561.3 Ma for standard M257 (Nasdala et al., 2008). Pb/U ratios in the unknown samples were corrected using the ln(Pb/U)/ln(UO/U) relationship as measured in M257 and as outlined in Compston et al. (1992) and Nelson (1997). The 1σ error in the ratio 208Pb/206U during analysis of all standard zircons during this study was 1.6%. Primary beam intensity was 9 nA, and a Köhler aperture of 100 μm diameter was used, giving a slightly elliptical spot size of about 30 μm. Peak resolution was 4324 (at 1% peak height) on mass 235U of the standard, enabling clear separation of the 208Pb-peak from the HfO peak. Analyses of samples and standards were alternated to allow assessment of Pb/U discrimination.

**Raw data reduction and error assessment** followed the method described by Nelson (1997). Common Pb corrections have been applied using the 204Pb-correction method and assuming the isotopic composition of Broken Hill, because common Pb is thought to be surface-related (Kinny, 1986). The analytical data are presented in Table 2. Errors given on individual analyses are based on counting statistics and are at the 1σ level and include the uncertainty of the standard added in quadrature. Errors for pooled analyses are at 2σ. The ages and errors were calculated for mean 207Pb/206Pb ratios, including the calibration error on 34 standards measured during the analytical session.

**A.2. Nd isotopic analysis**

Sm–Nd isotopic analyses were performed at the Institute of Precambrian Geology and Geochronology, St. Petersburg. About 100 mg of whole-rock powder was dissolved in a mixture of HF, HNO3 and HClO4. A 140Sm–150Nd spike solution was added to all samples prior to dissolution. REE were separated on BioRad AGW50-X8 200–400 mesh resin using conventional cation-exchange techniques. Sm and Nd were separated by extraction chromatography with a LN-Spec (100–150 mesh) resin. The total laboratory blanks were 0.1–0.2 ng for Sm and 0.1–0.5 ng for Nd. Isotopic compositions of Sm and Nd were determined on a TRITON Ti multicollector mass-spectrometer in static mode. The precision (2σ) of Sm and Nd contents and 143Sm/144Nd ratios was ca. 0.5% and ca. 0.005% for 143Nd/144Nd ratios.

143Nd/144Nd ratios were adjusted relative to a value of 0.518860 +0.000010 for the La Jolla standard. During the period of analysis the weighted average of 9 La Jolla Nd standard runs yielded 0.518852 ± 0.000012 (2σ) for 143Nd/144Nd, normalized against 146Nd/144Nd = 0.7219. The ɛNd(t) values were calculated using the present-day values for a chondritic uniform reservoir (CHUR) 143Nd/144Nd = 0.512638 and 147Sm/144Nd = 0.1967 (Jacobsen and Wasserburg, 1984). Whole-rock Nd εNd(DM) model ages were calculated using the models of Goldstein and Jacobsen (1988) according to which the Nd isotopic composition of the depleted mantle evolved linearly since 4.56 Ga ago and has a present-day value εNd(0) of +10 (143Nd/144Nd = 0.513151 and 147Sm/144Nd = 0.21365).

Two-stage or crustal Nd model ages εNd(t)(c) (Kato and Jacobson, 1987) were calculated using a mean crustal 143Sm/144Nd ratio of 0.12 (Taylor and McLennan, 1985).

**A.3. Hf isotopic analysis**

Zircon spots previously analyzed on SHRIMP II were selected for Hf isotopic analyses on a Nu Plasma HR MC-ICP-MS (Nu Instruments, UK), coupled to a 193 nm excimer laser ablation system (RESOLUTION M-50, Resonetics LLC, USA), installed in the Dept. of Earth Sciences, The University of Hong Kong. The instrumental settings are detailed in Xia et al. (2011). Data were acquired by ablatiing 55 μm (diameter) laser spots, and a 6 Hz repetition rate was used. The analyses consisted of a 30 s blank measurement prior to the start of ablation and 40 s of ablation. The measured isotopic ratios of 176Hf/177Hf were normalized to 179Hf/177Hf = 0.7325, using exponential correction for mass bias. Isobaric interference of 176Yb and 176Lu on 176Hf was corrected by monitoring 177Yb and 176Lu respectively. The in-situ measured 177Yb/176Yb ratio was used for mass bias correction for both Yb and Lu because of their similar physicochemical properties. Ratios used for the corrections were 0.5887 for 177Yb/176Yb and 0.02655 for 176Lu/176Yb (Vervoort et al., 2004). External corrections were applied to all unknowns, and standard zircons 91500 and GJ-1 were used as external standards and were analyzed twice before and after every 10 analyses. The measured 176Lu/177Hf ratios and a 176Lu decay constant of 1.865 × 10−11 y−1 as reported by Scherer et al. (2001) were used to calculate initial 176Hf/177Hf ratios. Calculation of εHf values is based on the chondritic values of 176Hf/177Hf and 176Lu/177Hf as reported by Blücher-Toft and Albarède (1997). The mantle extraction model age εHf(DM) was calculated using the measured 176Lu/177Hf of the zircon, but this only provides a minimum age for the source material of the magma from which the zircon crystalized. Therefore, we also have calculated a crustal model age εHf(c) (Table 4), which assumes that the parental magma of the zircons was produced from an average lower continental crust (176Lu/177Hf = 0.010, Kröner et al., 2013a, 2014) but was ultimately derived from the
References


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