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Paleoproterozoic S-type granites in the Helanshan Complex, Khondalite Belt, North China Craton: Implications for rapid sediment recycling during slab break-off

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ABSTRACT

S-type granites, typically derived from the rapid recycling of sedimentary rocks, are sometimes accompanied by contemporary mafic magmatism and granulite metamorphism. However, the geodynamic context for such rock suites is often highly disputed, with various model proposed, including back-arc basin opening, lithospheric delamination, mantle plume and continental rifting. The Paleoproterozoic Khondalite Belt in the North China Craton provides an example of synchronous mafic and felsic magmatism that was accompanied by granulite-facies metamorphic events for which the tectonic affinities of these rocks remains unclear. This study integrates in situ zircon Hf–O isotope analyses, whole-rock geochemistry and Nd isotope results for the earliest two-mica granites (ca. 1.95 Ga) in order to provide constraints on the above issues. The granites are strongly peraluminous (A/CNK value >1.1), and characterized by high zircon $\delta^{18}O$ values of 7.3–10.6‰, corresponding to calculated magmatic $\delta^{18}O$ values of 9.1–12.3‰, similar to those of typical S-type granites. They have relatively high and homogeneous $\varepsilon_{Nd}(t)$ values of −1.1 to +0.9 and highly variable zircon $\varepsilon_{Hf}(t)$ values ranging from −1.0 to +8.3. In situ zircon Hf–O isotopic compositions indicate that the S-type granites may contain some mantle or juvenile crustal components in addition to a sediment component. Based on the new results and published data, a slab break-off model is proposed to explain the rapid recycling of sedimentary precursors and the generation of the ca. 1.95 Ga S-type granites.

1. Introduction

Accretionary and collisional orogens are important sites for the partial melting of recycled sedimentary rocks to form S-type granites (e.g., Sylvester, 1998; Barbarin, 1999; Collins and Richards, 2008; Kemp et al., 2009). The time interval between the deposition of sedimentary rocks and formation of magmatic rocks can be less than 50 m.y. (e.g., Kemp et al., 2007; Zheng et al., 2008). Several mechanisms have been proposed to explain the rapid formation of S-type granites. For subduction systems or accretionary orogens, the rapid formation of S-type granites and related rocks has been accounted for by models of back-arc basin opening (Cenozoic Hidaka metamorphic belt, Japan; Kemp et al., 2007) and lithospheric delamination (Ordovician Lachlan Fold Belt, Australia; Foster and Goscombe, 2013) or by models that combine both features (Collins and Richards, 2008). The rapid formation of S-type granites in collisional orogenic belts, however, has commonly been attributed to mantle plumes (Jiangnan Orogen, South China Block, Li et al., 2003) or continental rifting (Jiangnan Orogen, Zheng et al., 2008). In some cases, coeval granulites and mafic magmatism accompanied sediment recycling, as in the Hidaka metamorphic belt, Japan (accretionary orogen, Kemp et al., 2007) and the Silurian Tongbai Orogen, Central China (collisional orogen, Wang et al., 2011a). Such examples of coeval mafic magmatism, felsic magmatism and granulite metamorphism provide multiple constraints on contemporary geodynamics, which can help to resolve the origin of S-type granites.

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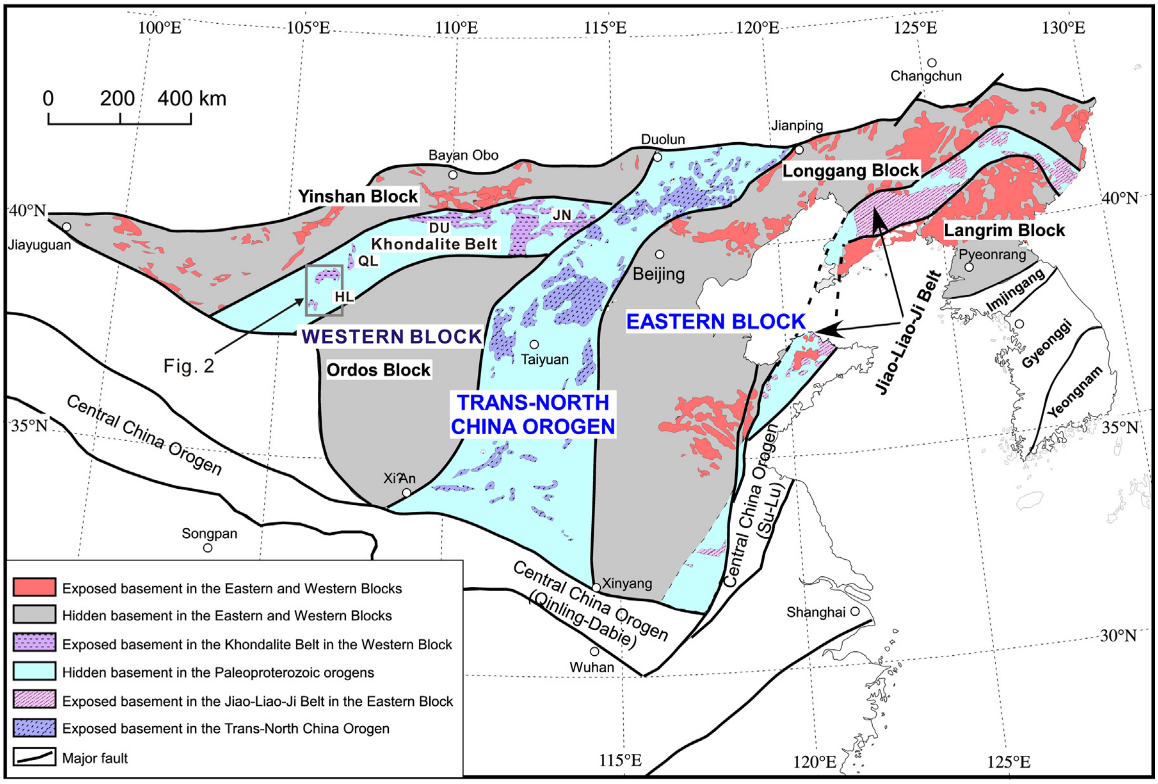

Fig. 1. Tectonic subdivision of the North China Craton (Zhao et al., 2005, 2012). Abbreviations for metamorphic complexes in the Khondalite Belt: DU, Daqingshan-Ulashan; HL, Helanshan; JN, Jining; QL, Qianlishan.

The Khondalite Belt in the North China Craton is characterized by widely distributed granulite facies metasedimentary rocks, along with dolerite dykes, S-type granites and minor exposures of ultra-high temperature (UHT) rocks (e.g., Lu et al., 1992, 1996; Guo et al., 2006; Santosh et al., 2007a, 2007b; Liu et al., 2012a; Zhang et al., 2012a; Zhao, 2014). Recent high-precision geochronological data have revealed magmatic episodes at ca. 2.56–2.50, 2.45–2.37, 2.30–2.00, 1.97–1.90 and 1.85–1.82 Ga, and metamorphic overprints at ca. 1.95–1.90 and 1.87–1.82 Ga (e.g., Jian et al., 2005; Xia et al., 2006a,b, 2008, 2009; Wan et al., 2006, 2009, 2013a; Yin et al., 2009, 2011; Zhao et al., 2010; Li et al., 2011a; Dan et al., 2012; Ma et al., 2012; Dong et al., 2013, 2014; Liu et al., 2013, 2014; Jiao et al., 2013). The ca. 1.95–1.90 Ga metamorphic event is commonly interpreted to have resulted from collisional orogeny to form the Khondalite Belt collisional orogen (e.g., Zhao et al., 2005, 2010; Yin et al., 2009, 2011, 2014; Zhao and Zhai, 2013). However, numerous ca. 1.97–1.95 Ga magmatic rocks have recently been discovered, including gabbro, carbonatite, gabbronorite, charnockite and S-type granite (Wan et al., 2008, 2013a; Geng et al., 2009; Song et al., 2010; Peng et al., 2010; Dan et al., 2012; Ma et al., 2012; Liu et al., 2013). Accordingly, a revised model is required to account for the formation of these magmatic rocks and the associated metamorphism. In this contribution, we report an integrated study of in situ zircon Hf–O isotopes as well as whole-rock geochemistry and Nd isotopes for the earliest S-type granites (ca. 1.95 Ga) (Dan et al., 2012). These new results, combined with available published data, are used to constrain the tectonic setting of the ca. 1.97–1.90 Ga magmatism and the rapidly recycled sedimentary rocks.

2. Geological background

The North China Craton (NCC) has been divided into four small Archean-Paleoproterozoic continental blocks (i.e., the Longgang, Langrim, Yinshan and Ordos blocks) and three Paleoproterozoic tectonic belts, named the Khondalite Belt, Trans-North China Orogen and Jiao-Liao-Ji Belt (Fig. 1), of which the Khondalite Belt formed by amalgamation of the Yinshan and Ordos blocks to form the Western Block at ~1.95 Ga (Zhao et al., 2005; Yin et al., 2009, 2011), whereas the Jiao-Liao-Ji Belt formed by amalgamation of the Longgang and Langrim blocks to form the Eastern Block at ~1.90 Ga (Li et al., 2004a, 2005a, 2006, 2011b, 2012; Li and Zhao, 2007; Luo et al., 2004, 2008; Zhou et al., 2008; Tam et al., 2011, 2012a,b,c). Finally, the NCC formed by amalgamation of the Eastern and Western blocks along the Trans-North China Orogen (TNCO) at ca. 1.85 Ga (e.g., Guo et al., 2005; Kröner et al., 2005; Zhang et al., 2007, 2009, 2012b; Zhao et al., 2001, 2005, 2010; Lu et al., 2008; Li et al., 2010a; Wang et al., 2010a; Liu et al., 2012b,c,d), though some alternative models have also been proposed for the formation and evolution of the craton (e.g., Kusky and Li, 2003; Kusky et al., 2007; Trap et al., 2007; Wang et al., 2010b), and a detailed overview of these models was given by Zhao and Cawood (2012).

The Khondalite Belt extends northwesterly over 1000 km, consisting of, from east to west, the Jining, Daqingshan-Ulashan, Qianlishan and Helanshan complexes (Fig. 1). Granulite facies metasedimentary rocks and S-type granites dominate the belt (e.g., Lu et al., 1996). The supracrustal metasedimentary sequences consist of graphite-garnet-sillimanite gneiss, garnet quartzite, felsic paragneiss, calc-silicate rock and marble, referred as the "Khondalite series" in the Chinese literature (Lu et al., 1992, 1996), whose protoliths are traditionally considered to have been deposited on a stable continental margin (Condie et al., 1992; Lu et al., 1992, 1996) and underwent three stages of metamorphism and deformation (Zhao et al., 1999). Recently, based on U–Pb isotopic detrital zircon dating results, Dan et al. (2012) proposed that the protoliths of the Khondalites were sourced mainly from ca. 2.18 to 2.00 Ga magmatic arcs, deposited on an active continental margin between ca. 2.00 and 1.95 Ga, and then subsequently metamorphosed at ca. 1.95–1.85 Ga (Xia et al., 2006a,b; Wan et al., 2006, 2009; Yin et al.,

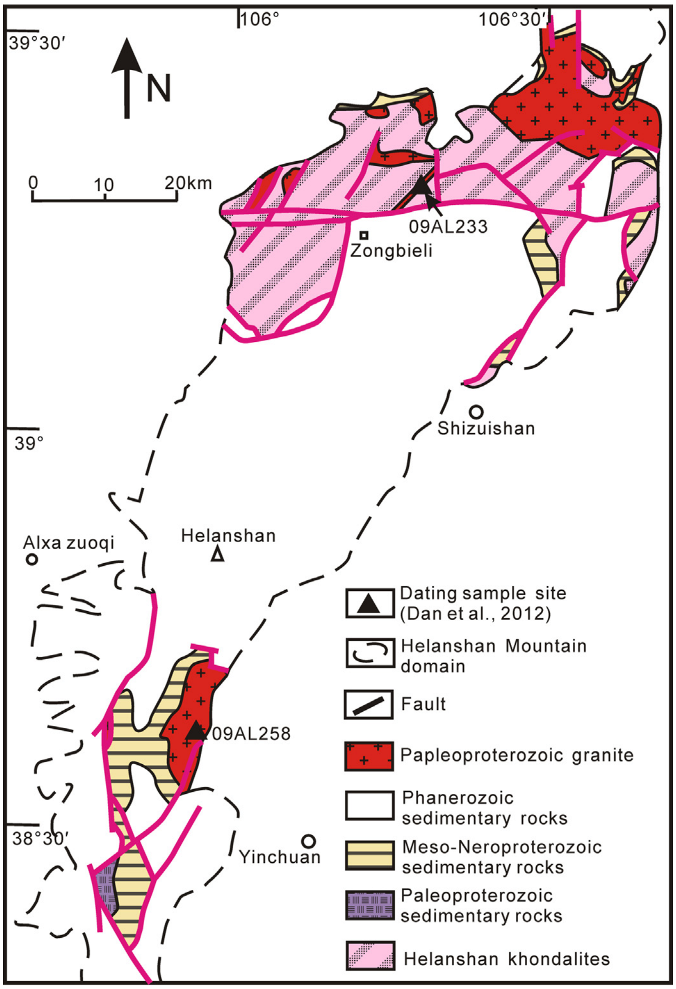

Fig. 2. Simplified geological map of the Helanshan Complex.

2009, 2011; Zhao et al., 2010; Guo et al., 2012; Santosh et al., 2013). The age of ca. 1.95 Ga is interpreted to reflect the timing of regional metamorphism linked to the collision and amalgamation of the Yinshan and Ordos blocks (e.g., Yin et al., 2009, 2011, 2014; Zhao et al., 2010; Guo et al., 2012; Santosh et al., 2013). Ultrahigh-temperature (UHT) assemblages have been recognized at several localities in recent years, mainly situated in the eastern Daqingshan-Ulashan and central Jining terranes (e.g., Guo et al., 2006; Santosh et al., 2007a,b; Liu et al., 2010, 2012a; Guo et al., 2012; Zhang et al., 2012a).

The Helanshan Complex is located in the westernmost part of the Khondalite Belt (Fig. 1) and consists mainly of Khondalite metasedimentary rocks and garnet-bearing S-type granites (Fig. 2). The metasedimentary rocks initially underwent high-grade metamorphism at ca. 1.95 Ga during collision of the Yinshan and Ordos blocks. A second high-grade event at ca. 1.87 Ga was synchronous with the formation of post-orogenic S-type granites (Yin et al., 2011), which have been attributed to anatectic melting of the metasedimentary units (e.g., Hu, 1994; Yin et al., 2011). The known minor granulite occurrences include both high-pressure and high-temperature/low pressure types (Zhou and Geng, 2009; Zhou et al., 2010). Dolerite dykes intruding the Khondalite units have been dated at ca. 1.96 Ga (recalculated from Song et al., 2010). New geochemical analyses were undertaken on the 1947 ± 6 Ma porphyritic K-feldspar granites and 1956 ± 19 Ma two-mica granites, both of which were dated in our previous study (Dan et al., 2012). The porphyritic K-feldspar granite occurs as a dyke, about 15 m wide, that intruded into the metasedimentary rocks (Fig. 3a), and the two-mica granite occurs as a pluton with an outcrop area of 90 km^2. Neither granite exhibits a deformational fabric. The granites have similar mineral assemblages, consisting of K-feldspar (20–50 vol.%) + plagioclase (10–35 vol.%) + quartz (20–40 vol.%) + biotite (3–8 vol.%) + muscovite (1–3 vol.%) (Fig. 3), and trace amounts of magnetite, apatite and zircon.

3. Analytical procedures

3.1. Major and trace elements

Ten rock samples powdered to ~200-mesh size were used for chemical analyses. Major element oxides were analyzed on fused glass beads using a Rigaku RIX 2000 X-ray fluorescence spectrometer at the State Key Laboratory of Isotope Geochemistry, the Guangzhou Institute of Geochemistry, Chinese Academy of Sciences (SKLaBIG-GIG-CAS). Calibration lines used in quantification were produced by bivariate regression of data from 36 reference materials encompassing a wide range of silicate compositions (Li et al., 2005b). Analytical uncertainties are between 1% and 5%. Trace elements were analyzed using an Agilent 7500a ICP-MS at GIG-CAS. Analytical procedures were similar to those described by Li et al. (2000). A set of USGS and Chinese national rock standards, including BHVO-2, GSR-1, GSR-2, GSR-3, AGV-2, W-2 and SARM-4 were chosen for calibration. Analytical precision typically is better than 5%. Geochemical results are listed in Table 1.

3.2. Nd isotopic compositions

Nd isotopic compositions were determined using a Micromass Isoprobe multi-collector ICP-MS at SKLaBIG-GIG-CAS, and analytical procedures described by Li et al. (2004b). Sr and Nd were separated using cation columns, and Nd fractions were further separated by HDEHP-coated Kef columns. The measured $^{143}Nd/^{144}Nd$ ratio of the JNdi-1 standard were 0.512093 ± 11 (n = 11, 2σ). All measured Nd isotope ratios were normalized to $^{146}Nd/^{144}Nd = 0.7219$. The Nd isotope results are listed in Table 2.

3.3. Zircon oxygen isotopes

Zircon oxygen isotopes were measured using the same Cameca IMS-1280 SIMS at IGG-CAS. The detailed analytical procedures were similar to those described by Li et al. (2010b). The measured oxygen isotopic data were corrected for instrumental mass fractionation (IMF) using the Penglai zircon standard ($\delta^{18}O_{VSMOW} = 5.3‰$) (Li et al., 2010c). The internal precision of a single analysis generally was better than 0.2‰ (1σ standard error) for the $^{18}O/^{16}O$ ratio. The external precision, measured by the reproducibility of repeated analyses of Penglai standard, is 0.38‰ (2SD, n = 16). Five measurements of the 91500 zircon standard during the course of this study yielded a weighted mean of $\delta^{18}O = 10.2 \pm 0.5‰$ (2SD), which is consistent within errors with the reported value of 9.9 ± 0.3‰ (Wiedenbeck et al., 2004). Zircon oxygen isotopic data are listed in Table 3.

3.4. Zircon Lu–Hf isotopes

In situ zircon Lu–Hf isotopic analyses were carried out on a Neptune multi-collector ICP-MS equipped with a Geolas-193 laser-ablation system at IGG-CAS. Lu–Hf isotopic analyses were conducted on the same zircon grains that were previously analyzed for U–Pb and O isotopes, with ablation pits of 60 or 44 μm in diameter, ablation time of 26 s, repetition rate of 8 Hz, and laser

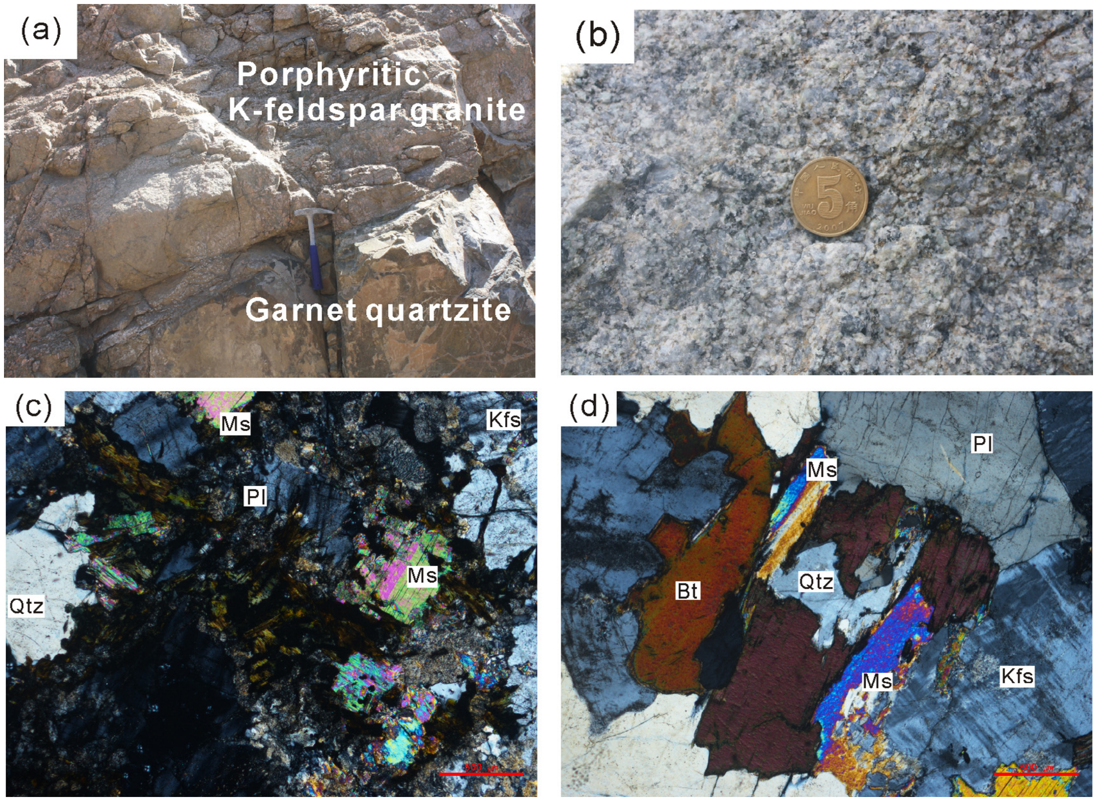

Fig. 3. Field photograph and photomicrographs for the Helanshan granites: (a, c) porphyritic K-feldspar granite and (b, d) two-mica granite. The mineral abbreviations are after Kretz (1983).
The (a) and (c) are reproduced from Dan et al. (2012).

beam energy density of 10 J/cm^2. Detailed analytical procedures were similar to those described by Wu et al. (2006). Measured $^{176}Hf/^{177}Hf$ ratios were normalized to $^{179}Hf/^{177}Hf = 0.7325$. Further external adjustment was not applied for the unknowns because our determined $^{176}Hf/^{177}Hf$ ratios for zircon standards 91500 (0.282309 ± 0.000004) and GJ-1 (0.282000 ± 0.000008) were in good agreement within errors with the reported values (Griffin et al., 2006; Wu et al., 2006). Zircon Hf isotopic data are listed in Table 3.

4. Results

4.1. Whole-rock major and trace element compositions

The porphyritic K-feldspar granites and two-mica granites from the Helanshan Complex plot in or adjacent to the granite field on a SiO_2 vs. $K_2O + Na_2O$ diagram (Fig. 4a), and have highly variable abundances of SiO_2, ranging from 65.9 to 74.6 wt.% (volatile free, Table 1). On a SiO_2 vs. K_2O diagram (Fig. 4b), they plot from the middle of the calc alkaline field to well within the field of shoshonitic magmatic rocks. These granites have variable and high A/CNK (molar $Al_2O_3/(CaO + Na_2O + K_2O)$) values of 1.08–1.52, and plot into the strongly peraluminous field (Fig. 4c). They are characterized by strong but variable light rare earth element (REE) enrichments ($La_N/Yb_N = 6.9–140$) (Fig. 5a), negative Eu anomalies (one sample exhibits a positive Eu anomaly caused by plagioclase accumulation), and high field strength element (HFSE, e.g., Nb, Ta, Ti), Sr and P depletions compared with neighboring elements on multi-element "spidergrams" (Fig. 5b). The porphyritic K-feldspar granites have higher LREE contents but both types display variable MREE–HREE contents.

4.2. Whole rock Sm–Nd isotopic compositions

Eight samples were selected for whole rock Sm–Nd isotopic analyses (Table 2). The porphyritic K-feldspar granites exhibit homogeneous $\varepsilon_{Nd}(t)$ values (−0.6 to +0.2) (Fig. 4d), corresponding to two-stage Nd mode ages (T_{2DM}) of 2.47–2.40 Ga, whereas the two-mica granites have slightly variable $\varepsilon_{Nd}(t)$ values (−1.1 to +0.9) (Fig. 4d), corresponding to T_{2DM} of 2.51–2.35 Ga.

4.3. Zircon Hf-O isotopic compositions

Hafnium and oxygen isotope analyses were conducted on those zircon grains that were previous analyzed for U–Pb dating (Dan et al., 2012). The ca. 1.95 Ga magmatic zircon grains from samples 09AL233 and 09AL258 have $\varepsilon_{Hf}(t)$ values of −1.0 to +3.0 (averaged at 1.1 ± 0.8 (1SD)) and +2.1 to +8.3 (averaged at +4.9 ± 1.7 (1SD)) (Fig. 6), corresponding to two-stage zircon Hf model ages (T_{DM}^{C}) of 2.66–2.41 Ga (averaged at 2.53 ± 0.05 Ga (1SD)) and 2.47–2.08 Ga (averaged at 2.29 ± 0.11 Ga (1SD)), respectively (Table 3).

The measured $\delta^{18}O$ values for ca. 1.95 Ga zircon grains from sample 09AL233 and 09AL258 show a wide range of 8.7–10.6‰ and 7.3–10.2‰, respectively (Fig. 7). One grain with age of 2058 Ma from 09AL233 has a $\delta^{18}O$ value of 7.2‰ (Table 3).

5. Discussion

5.1. 2.0–1.8 Ga events in the Khondalite Belt

The Khondalite Belt is commonly considered to have formed at ca. 1.95 Ga (e.g., Zhao et al., 2005; Yin et al., 2009, 2011), following the termination of a ca. 2.18–2.00 Ga magmatic arc between

Table 1
Geochemical data for the ca. 1.95 Ga granites, Helanshan Complex.

Sample	09AL225	09AL227	09AL228	09AL232	09AL233	09AL258	09AL259	09AL260	09AL261	09AL262
Rock type	Porphyritic K-feldspar granite					Two-mica granite				
Latitude (N)	39°17′01″	39°17′22″		39°18′17″		38°36′34″				
Longitude (E)	106°16′25″	106°17′03″		106°17′33″		105°56′05″				
Major element (%)										
SiO_2	65.89	67.76	69.26	74.65	68.64	71.42	69.56	68.51	69.68	73.13
TiO_2	0.42	0.61	0.80	0.38	0.35	0.20	0.23	0.24	0.70	0.49
Al_2O_3	15.37	16.00	13.93	13.66	16.64	16.02	16.59	18.05	14.23	14.32
$Fe_2O_3{}^T$	7.89	3.57	4.90	2.24	1.76	1.58	1.87	1.59	5.57	4.03
MnO	0.03	0.02	0.03	0.03	0.01	0.01	0.01	0.01	0.04	0.03
MgO	1.45	1.00	1.14	1.13	0.71	0.68	0.83	0.64	2.22	1.44
CaO	1.47	0.93	1.80	1.37	0.85	1.10	1.29	0.87	1.77	0.75
Na_2O	0.17	2.67	2.82	3.15	2.83	3.12	3.35	2.90	3.11	3.35
K_2O	6.95	7.05	4.56	3.40	8.21	5.89	6.32	7.20	2.54	2.37
P_2O_5	0.12	0.42	0.48	0.03	0.12	0.08	0.09	0.15	0.06	0.11
Total	99.74	100.03	99.73	100.04	100.13	100.09	100.13	100.16	99.92	100.01
L.O.I.	3.30	1.75	1.54	2.27	1.38	0.96	0.78	1.03	1.77	1.65
C	5.2	3.3	2.3	2.4	1.8	2.7	2.1	4.3	3.3	5.2
Mg#	28.8	38.1	33.9	52.7	47.1	48.5	49.4	46.9	46.7	44.1
A/CNK	1.47	1.17	1.08	1.20	1.10	1.18	1.13	1.27	1.28	1.52
T (°C)	825	860	870	860	778	798	807	754	807	803
Trace element (ppm)										
Sc	5.23	7.19	10.1	3.54	3.00	4.76	6.09	4.52	19.3	10.5
V	18.6	27.9	38.9	20.6	16.3	17.1	17.6	33.7	91.6	72.6
Cr	5.66	8.23	8.23	15.9	8.07	7.76	11.5	28.6	210	106
Co	6.22	4.66	7.37	3.49	3.22	2.95	3.59	4.14	12.6	10.0
Ni	5.02	4.39	4.92	7.80	3.85	5.15	7.19	9.08	37.1	17.4
Ga	22.7	22.7	23.1	18.1	18.8	14.4	15.5	15.7	19.6	16.2
Rb	262	313	248	167	258	161	183	197	154	123
Sr	99.4	376	386	279	229	262	268	216	142	84.7
Y	57.3	30.3	38.6	8.90	9.19	8.21	4.46	57.1	38.9	16.4
Zr	191	349	404	300	154	168	205	97.6	172	135
Nb	26.7	23.1	29.9	9.78	7.80	6.69	8.47	6.76	13.5	7.93
Cs	2.07	4.21	5.18	1.73	1.45	2.80	4.26	3.88	7.29	5.27
Ba	1283	1699	1221	634	754	1139	1215	921	161	159
La	90.7	152	176	135	92.0	36.7	17.4	59.0	34.4	25.2
Ce	192	294	352	253	191	73.1	31.7	122	70.9	53.5
Pr	23.2	33.7	40.2	27.4	21.9	7.98	3.57	13.9	7.88	5.93
Nd	87.1	117	140	90.8	76.8	29.8	13.5	51.8	30.6	23.0
Sm	17.0	17.6	20.3	10.9	11.7	5.32	2.44	11.0	5.83	4.67
Eu	1.31	2.86	2.00	1.45	1.67	1.30	1.34	1.35	0.63	1.00
Gd	16.4	21.2	25.2	18.1	7.18	5.59	2.51	10.1	5.71	3.53
Tb	2.53	2.73	3.13	1.93	0.67	0.76	0.33	1.75	0.95	0.54
Dy	11.6	11.0	12.5	6.97	2.69	2.97	1.35	8.83	5.08	2.85
Ho	1.79	1.47	1.82	0.78	0.34	0.29	0.14	1.80	1.20	0.59
Er	5.38	2.77	3.25	1.12	0.72	0.59	0.36	4.70	3.12	1.43
Tm	0.75	0.32	0.40	0.12	0.08	0.10	0.06	0.72	0.56	0.22
Yb	4.76	2.31	2.83	0.80	0.47	0.65	0.44	4.19	3.56	1.37
Lu	0.71	0.33	0.40	0.11	0.08	0.10	0.07	0.63	0.57	0.22
Hf	5.56	8.86	10.4	8.15	4.28	4.44	5.04	2.94	4.60	3.65
Ta	1.53	1.69	1.17	0.32	0.40	0.42	0.74	0.37	0.82	1.47
Pb	25.9	30.1	24.7	15.5	34.9	27.0	24.3	35.6	10.5	27.1
Th	43.3	35.8	47.3	47.8	35.1	10.7	5.03	20.4	8.37	4.87
U	12.2	4.58	4.90	2.19	1.58	2.71	1.68	6.18	4.27	1.68

Mg# = 100 × molar $Mg^{2+}/(Mg^{2+} + Fe^{2+})$, assuming FeO/(FeO + Fe_2O_3) = 0.9; A/CNK = molar Al_2O_3/(CaO + Na_2O + K_2O); T (°C), zircon saturation temperature, calculated based on Watson and Harrison (1983). C = normative corundum.

the Yinshan and Ordos blocks (Dan et al., 2012). In order to better understand the petrogenesis and tectonic significance of the ca. 1.95 Ga event, we have summarized age data for 2.0–1.8 Ga magmatism and metamorphism in the Khondalite Belt. Owing to the rapidly growing dating database for the belt, only high-resolution ages (LA-ICP-MS and SIMS) were considered. Fig. 8 shows the frequency distribution of magmatic and metamorphic ages from the region using ISOPLOT (Ludwig, 2003). The age distribution pattern has a strong peak at 1.95 Ga and two minor peaks at 1.92 and 1.84 Ga. Magmatic activities extend from 2.0 to 1.8 Ga and can be divided into two main periods: 1.97–1.90 and 1.89–1.82 Ga. The latter event is mainly represented by un-deformed granites and associated metamorphism and is commonly attributed to regional extension, probably during orogenic collapse (e.g., Yin et al., 2009; Jiao et al., 2013; Liu et al., 2013).

The ca. 1.97–1.90 Ga period is represented by development of diverse magmatic rocks and associated metamorphism corresponding to the earliest major igneous activity following deposition of the sedimentary protoliths of the Khondalites (ca. 2.00–1.97 Ga). This period can be roughly divided into two events, i.e., the ca. 1.97–1.95 Ga and ca. 1.93–1.90 Ga, with corresponding peak ages of ca. 1.95 Ga and ca. 1.92 Ga, respectively (Fig. 8). The ca. 1.97–1.95 Ga event is recorded by various magmatic rocks, including metagabbros (Wan et al., 2013a), dolerite dykes (Song et al., 2010), charnockitic gneisses (Liu et al., 2013), carbonatitic dykes (Wan et al., 2008), syenogranites (Liu et al., 2013) and S-type granites (Dan et al., 2012). The related ca. 1.95 Ga metamorphism was recorded by almost all rocks types in all complexes of the Khondalite Belt, including meta-igneous rocks (Wan et al., 2013a), meta-sedimentary rocks (Yin et al., 2009, 2011), as well

Table 2
Whole rock Nd data for the ca. 1.95 Ga granites, Helanshan Complex.

Sample	Sm	Nd	^{147}Sm/^{144}Nd	^{143}Nd/^{144}Nd	±2σ	$\varepsilon_{Nd}(t)$	T_{DM} (Ma)	T_{2DM} (Ma)
Porphyritic K-feldspar granite								
09AL225	17.0	87.1	0.1181	0.511640	0.000007	0.19	2404	2404
09AL227	17.6	117	0.0909	0.511249	0.000006	−0.62	2355	2470
09AL232	10.9	90.8	0.0725	0.511039	0.000006	−0.10	2274	2428
09AL233	11.7	76.8	0.0919	0.511287	0.000008	−0.11	2326	2429
Two-mica granite								
09AL258	5.32	29.8	0.1078	0.511532	0.000006	0.69	2324	2364
09AL259	2.44	13.5	0.1088	0.511557	0.000008	0.91	2312	2346
09AL260	11.0	51.8	0.1284	0.511728	0.000005	−0.65	2536	2472
09AL262	4.67	23.0	0.1225	0.511629	0.000007	−1.12	2538	2510

$\varepsilon_{Nd}(t) = 10\,000 \times \{[(^{143}Nd/^{144}Nd)_s - (^{147}Sm/^{144}Nd)_s \times (e^{\lambda t} - 1)]/[(^{143}Nd/^{144}Nd)_{CHUR,0} - (^{147}Sm/^{144}Nd)_{CHUR} \times (e^{\lambda t} - 1)] - 1\}$; $T_{DM} = 1/\lambda \times \ln\{1 + [(^{143}Nd/^{144}Nd)_s - (^{143}Nd/^{144}Nd)_{DM}]/[(^{147}Sm/^{144}Nd)_s - (^{147}Sm/^{144}Nd)_{DM}]\}$; $T_{2DM} = T_{DM} - (T_{DM} - t)((f_c - f_s)/(f_c - f_{DM}))$; $f_{Sm/Nd} = (^{147}Sm/^{144}Nd)_s/(^{147}Sm/^{144}Nd)_{CHUR} - 1$; where f_c, f_s and f_{DM} are the $f_{Sm/Nd}$ values of the continental crust, sample and the depleted mantle; $f_c = -0.4$, $f_{DM} = 0.08592$; t = crystallization time; $(^{147}Sm/^{144}Nd)_s$ and $(^{143}Nd/^{144}Nd)_s$ are values of analysed sample; $(^{147}Sm/^{144}Nd)_{CHUR} = 0.1967$ and $(^{143}Nd/^{144}Nd)_{CHUR,0} = 0.512638$; $(^{147}Sm/^{144}Nd)_{DM} = 0.2135$ and $(^{143}Nd/^{144}Nd)_{DM} = 0.51315$; $(^{147}Sm/^{144}Nd)_c = 0.118$; $\lambda_{Sm-Nd} = 0.00654\ Ga^{-1}$.

as mafic and felsic granulites (Zhou and Geng, 2009; Zhou et al., 2010). The ca. 1.95 Ga metamorphic event has commonly been attributed to collision of the Yinshan and Ordos blocks (e.g., Zhao et al., 2005, 2010; Yin et al., 2009, 2011, 2014; Wang et al., 2011b; Zhao and Guo, 2012), but it is noted that the reported ca. 1.97–1.95 Ga magmatic rocks are almost entirely confined to the west-central segment of the Khondalite Belt (i.e., Helanshan and Daqingshan complexes). One possible explanation is that there is a tectonic boundary between the west-central and east segments of the Khondalite Belt (Peng et al., 2011). Another explanation is that the ca. 1.97–1.95 Ga magmatic rocks were less extensively developed toward the present-day northeast and may be covered by the extensive ca. 1.93–1.88 Ga volcanico-sedimentary sequences of the Jining Complex, whose generation has been attributed to a ridge subduction event (Peng et al., 2010, 2011, 2012).

Examples of the ca. 1.93–1.90 Ga magmatic and metamorphic events are also common features across the whole Khondalite Belt (Zhong et al., 2007; Geng et al., 2009; Wan et al., 2009; Yin et al., 2009; Peng et al., 2010, 2011; Santosh et al., 2007a,b, 2009, 2013). Most ca. 1.93–1.92 Ga ages metamorphic ages were obtained from UHT rocks in the Jining Complex by Santosh et al. (2007a,b, 2009, 2013). The tectonic setting responsible for generation of the UHT rocks has been highly disputed and various models invoke a post-collisonal setting (Zhao, 2009; Zhao et al., 2012), slab break-off (Zhai and Santosh, 2011), a mantle plume (Santosh et al., 2010), and ridge subduction (Peng et al., 2010, 2011, 2012). This ca. 1.93–1.90 Ga event is sometimes considered to be the second stage of the ca.

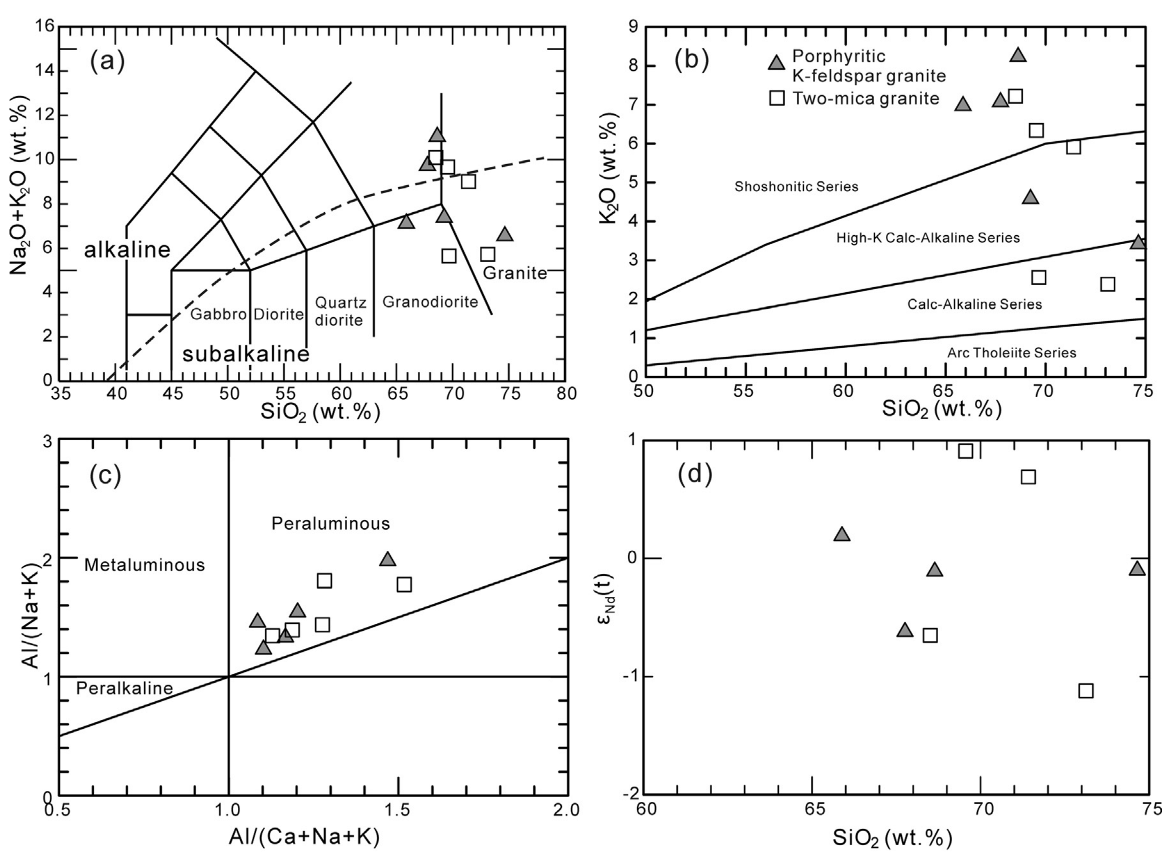

Fig. 4. Diagrams of (a) SiO_2 vs. $K_2O + Na_2O$, (b) SiO_2 vs. K_2O, (c) A/CNK vs. A/NK, and (d) SiO_2 vs. $\varepsilon_{Nd}(t)$ values.

Table 3
In situ zircon Hf–O isotopic results for the ca. 1.95 Ga granites in Helanshan Complex.

Sample spot	Age (Ma)	$^{176}Lu/^{177}Hf$	$^{176}Hf/^{177}Hf$	2σ	$\varepsilon_{Hf}(t)$	2σ	T_{DM} (Ma)	T^{C}_{DM} (Ma)	$\delta^{18}O$ (‰)
09AL233									
09AL233@1	1947	0.000766	0.281658	0.000035	3.0	1.2	2222	2405	9.64
09AL233@2	1947	0.000891	0.281619	0.000032	1.5	1.1	2282	2502	10.24
09AL233@3	1947	0.000855	0.281590	0.000025	0.5	0.9	2320	2567	10.18
09AL233@4	1947	0.000747	0.281606	0.000022	1.2	0.8	2292	2521	10.06
09AL233@5	1947	0.000916	0.281586	0.000022	0.3	0.8	2329	2579	9.82
09AL233@6	1947	0.001152	0.281558	0.000026	-1.0	0.9	2381	2662	10.46
09AL233@7	2058	0.000789	0.281581	0.000028	2.7	1.0	2328	2508	7.23
09AL233@8	1947	0.000843	0.281594	0.000027	0.7	1.0	2313	2555	9.43
09AL233@9	1947	0.000874	0.281595	0.000024	0.7	0.9	2314	2555	9.84
09AL233@10	1947	0.001003	0.281613	0.000024	1.1	0.8	2297	2526	10.64
09AL233@11	1947	0.001679	0.281625	0.000024	0.7	0.8	2321	2555	10.00
09AL233@12	1947	0.000909	0.281599	0.000024	0.8	0.8	2310	2549	9.62
09AL233@13	1947	0.001317	0.281650	0.000023	2.0	0.8	2266	2470	10.36
09AL233@14	1947	0.000834	0.281618	0.000021	1.5	0.7	2280	2500	8.74
09AL233@15	1947	0.000892	0.281605	0.000021	1.0	0.7	2301	2535	10.59
09AL233@16	1947	0.000871	0.281620	0.000023	1.6	0.8	2279	2498	10.10
09AL233@17	1947	0.000875	0.281593	0.000022	0.6	0.8	2317	2562	10.54
09AL233@18	1947	0.000606	0.281626	0.000017	2.1	0.6	2256	2464	10.49
09AL233@19	1947	0.000915	0.281619	0.000025	1.4	0.9	2284	2506	10.46
09AL233@20	1947	0.000890	0.281595	0.000022	0.6	0.8	2315	2557	n.d.
09AL233@21	1947	0.000920	0.281602	0.000020	0.8	0.7	2307	2544	n.d.
09AL258									
09AL258@1	1956	0.001768	0.281704	0.000026	3.5	0.9	2217	2379	8.24
09AL258@2	1956	0.001665	0.281743	0.000017	5.1	0.6	2155	2280	8.04
09AL258@3	1956	0.002140	0.281700	0.000018	2.9	0.6	2245	2419	9.00
09AL258@4	1956	0.002339	0.281821	0.000017	6.9	0.6	2084	2161	8.20
09AL258@5	1956	0.001078	0.281720	0.000014	5.0	0.5	2155	2285	9.85
09AL258@6	1956	0.001575	0.281735	0.000029	4.9	1.0	2162	2292	10.16
09AL258@7	1956	0.001115	0.281639	0.000028	2.1	1.0	2269	2472	8.54
09AL258@8	1956	0.001515	0.281713	0.000027	4.2	1.0	2189	2336	9.18
09AL258@9	1956	0.001449	0.281714	0.000022	4.3	0.8	2184	2328	9.54
09AL258@10	1956	0.003042	0.281884	0.000029	8.3	1.0	2033	2076	7.27
09AL258@11	1956	0.002117	0.281817	0.000021	7.1	0.7	2077	2150	8.43
09AL258@12	1956	0.002791	0.281727	0.000021	3.0	0.7	2245	2412	8.21
09AL258@13	1956	0.001350	0.281649	0.000021	2.1	0.7	2269	2469	8.94
09AL258@14	1956	0.002042	0.281701	0.000029	3.1	1.0	2237	2408	10.12
09AL258@15	1956	0.002179	0.281811	0.000021	6.8	0.8	2089	2169	7.65
09AL258@16	1956	0.001718	0.281770	0.000025	6.0	0.9	2121	2224	8.88
09AL258@17	1956	0.002361	0.281743	0.000019	4.1	0.7	2197	2340	8.79
09AL258@18	1956	0.001507	0.281709	0.000021	4.1	0.7	2194	2345	7.94
09AL258@19	1956	0.002090	0.281772	0.000026	5.5	0.9	2139	2251	7.81
09AL258@20	1956	0.001504	0.281705	0.000021	3.9	0.7	2199	2353	n.d.
09AL258@21	1956	0.001330	0.281678	0.000028	3.2	1.0	2226	2400	n.d.
09AL258@22	1956	0.001607	0.281781	0.000024	6.5	0.9	2099	2190	n.d.
09AL258@23	1956	0.001129	0.281765	0.000024	6.5	0.9	2095	2186	n.d.
09AL258@24	1956	0.002408	0.281808	0.000031	6.4	1.1	2107	2196	n.d.
09AL258@25	1956	0.000919	0.281772	0.000035	7.1	1.2	2073	2151	n.d.
09AL258@26	1956	0.001955	0.281769	0.000017	5.6	0.6	2136	2246	n.d.
09AL258@27	1956	0.001793	0.281667	0.000022	2.2	0.8	2270	2465	n.d.
09AL258@28	1956	0.001365	0.281759	0.000018	6.0	0.6	2117	2220	n.d.

n.d., not determined.
$\varepsilon_{Hf}(t) = 10\,000\{[(^{176}Hf/^{177}Hf)_s - (^{176}Lu/^{177}Hf)_s \times (e^{\lambda t} - 1)]/[(^{176}Hf/^{177}Hf)_{CHUR(0)} - (^{176}Lu/^{177}Hf)_{CHUR} \times (e^{\lambda t} - 1)] - 1$; $T_{DM} = 1/\lambda \times \ln\{1 + [(^{176}Hf/^{177}Hf)_s - (^{176}Hf/^{177}Hf)_{DM}]/[(^{176}Lu/^{177}Hf)_s - (^{176}Lu/^{177}Hf)_{DM}]\}$; $T_{DM}{}^{C} = T_{DM} - (T_{DM} - t) \times [f_{cc} - f_s/(f_{cc} - f_{DM})]$; $f_{Lu/Hf} = (^{176}Lu/^{177}Hf)_s/(^{176}Lu/^{177}Hf)_{CHUR} - 1$; where, f_{CC}, f_s and f_{DM} are the $f_{Lu/Hf}$ values of the continental crust, zircon sample and the depleted mantle; subscript s = analyzed zircon sample, CHUR = chondritic uniform reservoir; DM = depleted mantle; t = crystallization time or metamorphic time of zircon; $\lambda = 1.867 \times 10^{-11}$ year^{-1}; $^{176}Hf/^{177}Hf_{DM} = 0.28325$; $^{176}Lu/^{177}Hf_{DM} = 0.0384$; present-day $^{176}Hf/^{177}Hf_{CHUR(0)} = 0.282772$; $^{176}Lu/^{177}Hf_{CHUR} = 0.0332$; $^{176}Hf/^{177}Hf_{CC} = 0.015$.

1.97–1.95 Ga event, rather than a discrete episode (Liu et al., 2013), as application of the Ti-in-zircon geothermometer suggests that the peak UHT metamorphic temperature might be slightly higher than the closure temperature of zircon (Liu et al., 2010).

5.2. Source components for S-type granites

The ca. 1.95 Ga granites are one of the major rock types of the Helanshan Complex. The granites contain peraluminous minerals (i.e., muscovite), have >1% normative corundum (Table 1), and are characterized as strongly peraluminous with A/CNK values of 1.1–1.5. Samples 09AL233 and 09AL258 have similar wide ranges of zircon $\delta^{18}O$ values, 8.7–10.6 ‰ and 7.3–10.2 ‰, corresponding to calculated magmatic $\delta^{18}O$ values of 10.4–12.3 ‰ and 9.1–12.0 ‰, respectively (Lackey et al., 2008). The wide range of oxygen isotope compositions and high $\delta^{18}O$ values are typical of S-type granites (e.g., Kemp et al., 2008; Appleby et al., 2010; Dan et al., 2014). All of these lines of evidence indicate that the Helanshan granites can be classified as S-type.

S-type magmas are commonly produced by partial melting of metapelites and metagreywackes under water-undersaturated conditions (e.g., Vielzeuf and Holloway, 1988; Patiño Douce and Harris, 1998; Zhang et al., 2004; Guo and Wilson, 2012; Wang et al., 2012). A series of experiments show that metasediments tend to yield melts with low FeO abundances (generally <3 wt.% FeO) despite variations in Al_2O_3 content (Patiño Douce and

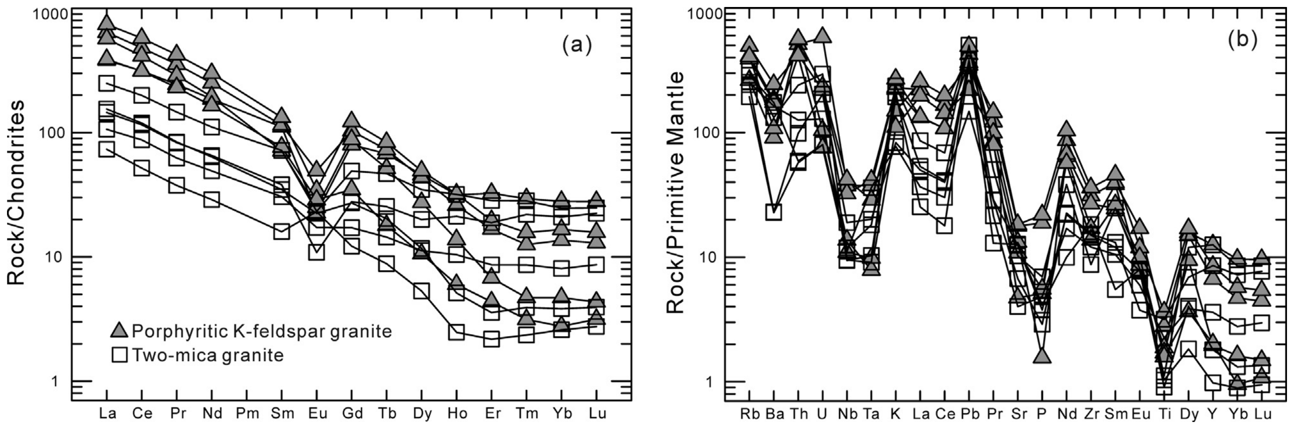

Fig. 5. Chondrite-normalized REE diagrams (a) and (b) primitive mantle-normalized incompatible trace element spidergrams for the granites. The normalization values are from Sun and McDonough (1989).

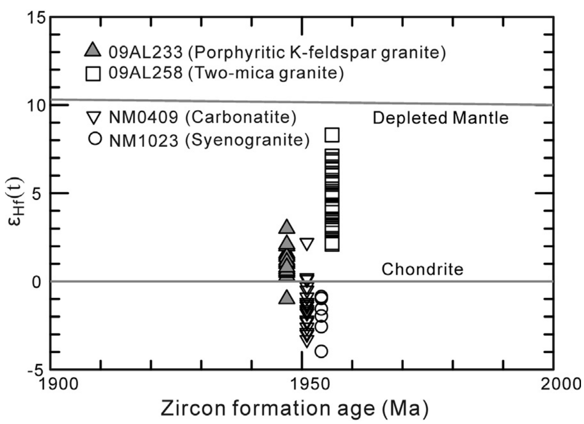

Fig. 6. Plot of $\varepsilon_{Hf}(t)$ values vs. zircon ages for the Paleoproterozoic rocks in Khondalite Belt.

Data sources: carbonatite Wan et al. (2008); syenogranite, Ma et al. (2012).

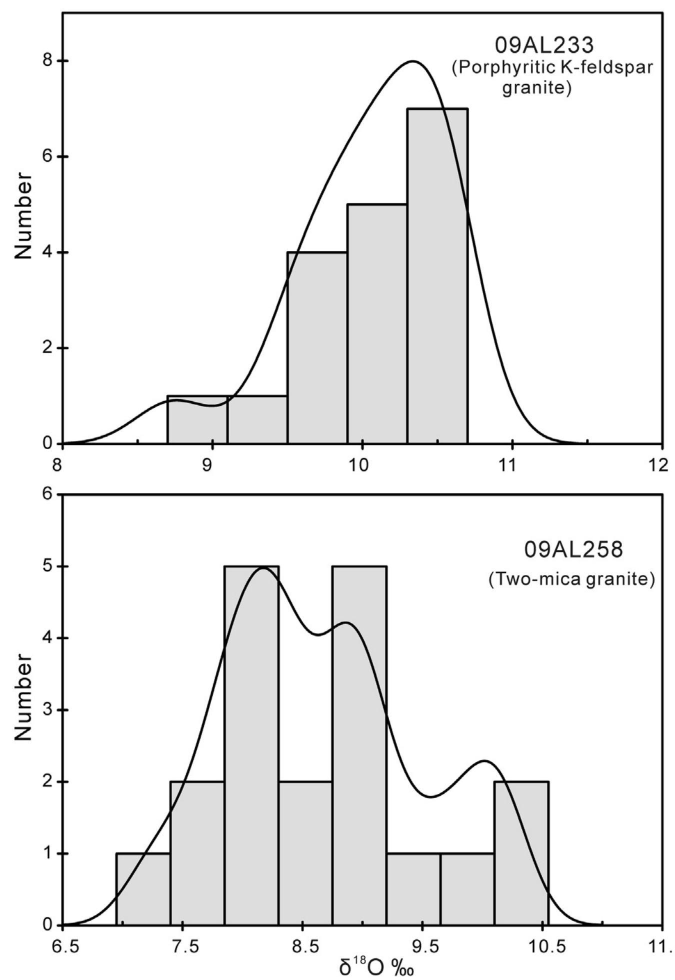

Fig. 7. Probability plots of zircon δ^{18}O values for the Paleoproterozoic granites.

Johnston, 1991; Montel and Vielzeuf, 1997; Patiño Douce and Harris, 1998;Fig. 9a). The exception is melts produced at high temperatures (1050–1250 °C) from metapelitic sources that started with a high FeO content, as shown by Vielzeuf and Holloway (1988). This appears to be due to the breakdown of garnet to spinel and quartz, liberating FeO into the melt (in addition to Al_2O_3, giving a relatively high abundance of both elements) (Turner and Rushmer, 2009). Compared to the typically metapelite-derived Himalayan leucogranites (Patiño Douce and Harris, 1998), the Helanshan S-type granites have higher FeO content (Fig. 9a). Thus, it might be argued that the Helanshan S-type granites were derived from high temperature metapelite-derived melts, which underwent plagioclase crystallization (Fig. 9a). However, their low MgO contents compared to experimental high-temperature melts do not support this scenario (Fig. 9b). Another possibility is that another component was added to the metasediments-derived melts. Mantle-derived mafic magmas generally have high FeO contents, satisfying this requirement. The occurrence of contemporaneous mafic dykes in the Helanshan Complex (Song et al., 2010) further supports this possibility.

This suggestion is also supported by the in situ isotopic data. In previous studies, the biotite and/or garnet-bearing granites were attributed solely to ananetic melting of Khondalite metasediments during regional metamorphism (Lu et al., 1992, 1996). The $\varepsilon_{Nd}(t)$ values of −1.1 to +0.9 obtained from the ca. 1.95 granites was considered to be evidence that they were generated by partial melting of a single juvenile sedimentary source, consistent with other S-type granites in the Khondalite Belt (e.g., Peng et al., 2012). Metasedimentary rocks of the Helanshan Complex, however, have significantly lower $\varepsilon_{Nd}(t)$ values of −7.7 to −1.2 (recalculated at 1950 Ma from Wan et al., 2000), indicating that other mantle-derived or juvenile crustal materials were added to the granites. These granites also have zircon $\varepsilon_{Hf}(t)$ values (−1.0 to +8.3) that are higher than those of ca. 1.95 Ga meta-syenogranite veins (−4.0 to +2.3) and crustal carbonatites (−3.2 to +0.7) in the Daqingshan Complex (Fig. 6) (Wan et al., 2008; Ma et al., 2012). In situ zircon Hf and O isotope data further suggest that the mantle-derived or juvenile crust-derived melts contributed to the ca. 1.95 Ga

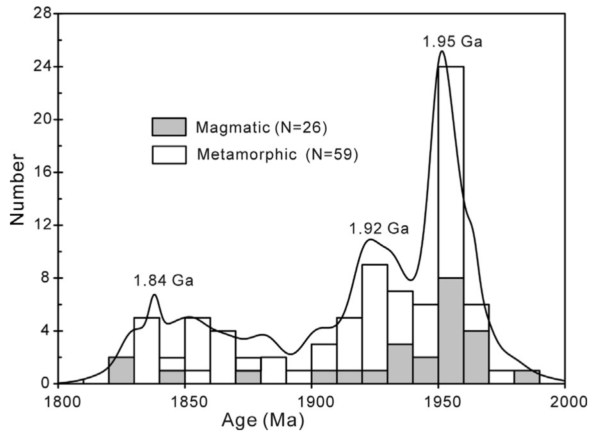

Fig. 8. Cumulative age spectra for the ca. 2.0–1.8 Ga ages from the Khondalite Belt. Data sources are in Supplementary Table A.1.

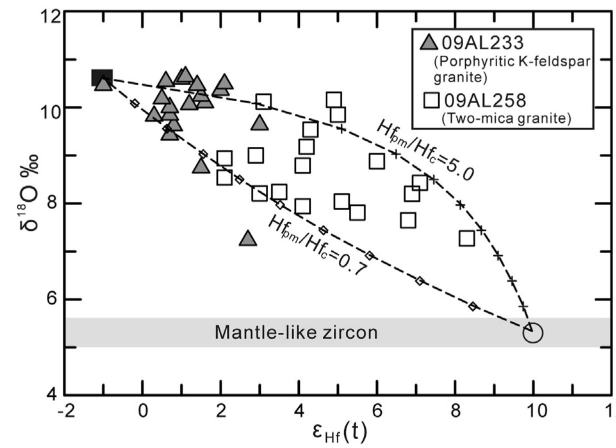

Fig. 10. Plot of zircon $\varepsilon_{Hf}(t)$ values vs. $\delta^{18}O$ values. The dashed lines denote the two-component mixing trends between the mantle- and supracrust-derived magmas. Hf_{pm}/Hf_c is the ratio of Hf concentration in the parental mantle magma (pm) over crustal (c) melt indicated for each curves, and the small rhombuses and ticks on the curves represent 10% mixing increments by arbitrary assuming the mantle zircon has ε_{Hf} = 10 and $\delta^{18}O$ = 5.3‰, and crustal zircon = −1 and $\delta^{18}O$ = 10.6‰, respectively.

S-type granites (Fig. 10), although the contribution to the porphyritic K-feldspar granite (sample 09AL233) was minor.

It is noted that there are many examples of cryptic magma mixing known from younger granites in the NCC (e.g., Sun et al., 2010), but no mafic enclaves have been discovered in those rocks. This can be explained by the fact that, like the granites of the present study, they were highly evolved (SiO_2 = 66–71 wt.%) rather than diorites or granodiorites. In such rocks, magma evolution can erase the macro-scale evidence of magma mixing, which is now only revealed by zircon Hf-O isotope studies (e.g., Appleby et al., 2008; Sun et al., 2010). Differences in terms of MgO, and SiO_2 abundances or $\varepsilon_{Nd}(t)$ and $\varepsilon_{Hf}(t)$) compositions between the porphyritic K-feldspar and two-mica granites (Figs. 4, 9 and 10) do not require that they were generated by different types of processes. Their geochemical differences can be explained in at least two ways. First, the two-mica granites may have undergone more crustal-level evolution than the porphyritic K-feldspar granites, as the former occurs as a pluton and latter as a dyke. Second, the two types of granites may have slightly different metasedimentary protolith components (e.g., Wan et al., 2000).

5.3. Geodynamics: slab break-off during ca. 1.97–1.90 Ga?

The ca. 1.95 Ga S-type granites were mainly sourced from the Khondalite metasedimentary rocks, whose protoliths are constrained to be deposition between ca. 2.00 and 1.95 Ga (Yin et al., 2011; Dan et al., 2012). As previously noted, the requirements for such rapid sedimentary recycling are only met in a few distinctive tectonic settings that involve processes such as lithospheric delamination, back-arc basin opening, mantle plume ascent or continental rifting (Li et al., 2003; Kemp et al., 2007; Zheng et al., 2008; Collins and Richards, 2008; Foster and Goscombe, 2013). A common feature of these tectonic settings is the presence of asthenosphere upwelling, consistent with the generation of high-temperature S-type granites (Sylvester, 1998).

The Helanshan Complex S-type granites can also be classified as high-temperature type. They have low Al_2O_3/TiO_2 (17–81), and plot into the "high-temperature" field (Fig. 11a), consistent with their calculated zircon saturation temperatures, which extend to ∼870 °C (Fig. 11b, Table 1). Although several inherited zircon grains were identified during geochronological studies (Dan et al., 2012), they can only have slightly raised the calculated zircon saturation temperatures. For example, the calculated Ti-in zircon temperatures are similar to the calculated zircon saturation temperatures in the Mid-Miocene to Quaternary strongly peraluminous rhyolites in the Southern Kunlun Range (Wang et al., 2012). The high zircon saturation temperatures were also supported by the high TiO_2 contents in these samples (Fig. 11b), which are temperature dependent under conditions of uniform TiO_2 activity (Wark et al., 2007). Furthermore, the Helanshan S-type granites

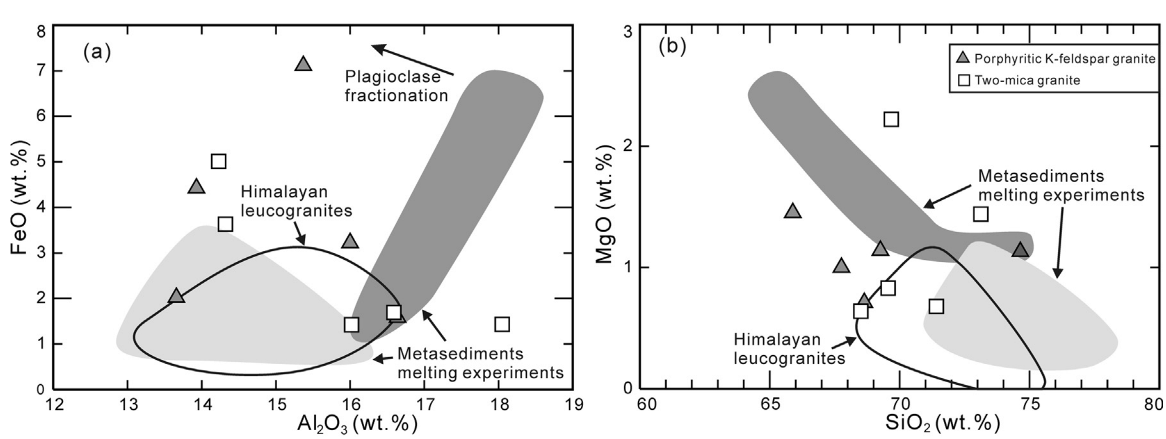

Fig. 9. Plots of (a) Al_2O_3 vs. FeO and (b) MgO vs. SiO_2. Metasediments experimental data: light shaded area: Patiño Douce and Johnston, 1991; Montel and Vielzeuf, 1997; Patiño Douce and Harris, 1998; dark shaded area: Vielzeuf and Holloway, 1988.
Data for the Himalayan leucogranites: Visonà and Lombardo (2002), Zhang et al. (2004), King et al. (2011), Guo and Wilson (2012), and references therein.

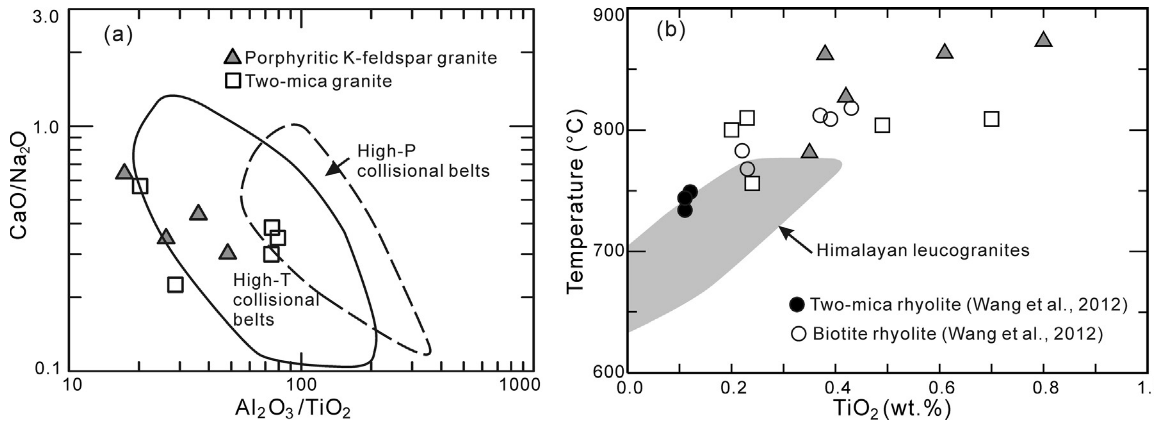

Fig. 11. (a) CaO/Na_2O vs. Al_2O_3/TiO_2 diagram for the granites. The data for fields of high-temperature and high-pressure collisional orogens are compiled by Sylvester (1998). (b) SiO_2 vs. temperature for the granites.
Data sources for the Himalayan leucogranites are the same as in Fig. 9.

have distinctly higher zircon saturation temperatures than those of the Himalaya leucogranites (Fig. 11b), which were produced by fluid-present melting (e.g., Patiño Douce and Harris, 1998) or decompression melting (e.g., Guo and Wilson, 2012). Given that both the Himalaya leucogranites and Helanshan S-type granites have inherited zircons, the effects of inheritance on the zircon saturation temperatures are offset in any comparison of the two suites.

High-temperature S-type granites, formed in a high-temperature collisional orogen, are generally cordierite-bearing peraluminous granites that are considered to be crust-mantle hybrids (e.g., Barbarin, 1996; Sylvester, 1998; Healy et al., 2004). However, this observation does not exclude the possibility that other S-type granites can be produced by mixing between mantle and crust-derived magmas. One example is the Permian peraluminous granites of the Lhasa terrane (Zhu et al., 2009). The inferred petrogenesis of the high temperature Helanshan S-type granites, involving mixing between mantle and crust-derived magmas, further demonstrates that asthenospheric upwelling is probably required to form these S-type granites (Fig. 10). More direct evidence may be provided by the ca. 1.96 Ga dolerite dykes (Song et al., 2010) of the Khondalite Belt, although additional geochemical data is required to confirm this speculation.

The tectonic settings typically applied to high-temperature S-type granites present problems if applied to the Helanshan Complex or the Khondalite Belt as a whole. The Khondalite Belt is considered to have been formed at ca. 1.95 Ga (e.g., Zhao et al., 2005, 2010), which argues against back-arc extension for generating the ca. 1.95 Ga magmatic rocks. Furthermore, the ca. 1.95 Ga granites in this study show significant geochemical variations from mid-calc-alkaline to shoshonitic compositions, implying involvement of heterogeneous mantle and/or crustal source regions in the petrogenesis that is not characteristic of the main, “normal”, stage of subduction (Lee et al., 2009). Similarly, lithospheric delamination models based on an active subduction system, such as those found in Mesozoic North China (e.g., Wu et al., 2008) and Late Paleozoic Iberia (e.g., Gutiérrez-Alonso et al., 2011), are also not viable. The lack of plume-related rocks (high-temperature lavas, flood basalts), the linear distribution of magmatic events, and 1.97–1.90 Ga collisional setting (Jiao et al., 2013) are also inconsistent with a plume model.

Taking into account the distribution of ca. 1.97–1.90 Ga magmatic rocks in the entire Khondalite Belt, a model of slab break-off is proposed, which predicts a relatively narrow, linear zone of magmatism located along a suture zone (e.g., von Blanckenburg and Davis, 1995; Atherton and Ghani, 2002; Chung et al., 2005; Keskin et al., 2008; Xu et al., 2008; Whalen et al., 2006, 2010; Yuan et al., 2010). Several lines of evidence support this scenario. The ca. 1.95 Ga event consists of simultaneous felsic and mafic magmatism, such as the ca. 1.95 Ga S-type granites and dolerite dykes in Helanshan Complex (Song et al., 2010; Dan et al., 2012) and the ca. 1.97–1.95 Ga Gabbros, syenogranites and carbonatites in Daqingshan Complex (Wan et al., 2008, 2013a; Ma et al., 2012; Liu et al., 2013). The carbonatites have characteristic enrichments in Sr, Ba and REE combined with depletions in HFSE, similar to other carbonatites emplaced in post-collisional settings (Hou et al., 2006; Chakmouradian et al., 2008). This period of magmatism was accompanied by regional ca. 1.95 Ga granulite facies metamorphism recorded by felsic and mafic granulites throughout the entire Khondalite Belt (e.g., Zhao et al., 2005, 2010; Yin et al., 2009, 2011). Finally, the tectonic setting during the ca. 1.97–1.90 Ga period was collisional or post-collisional (e.g., Jiao et al., 2013; Liu et al., 2013). The combination of contemporaneous mafic and felsic magmatism accompanied by coeval metamorphism in a post-collisional orogen is similar to other Phanerozoic orogens, such as the Alps (von Blanckenburg and Davis, 1995) and Tibet (Chung et al., 2005; Lee et al., 2009; Jiang et al., 2014; Ma et al., 2014).

The Paleoproterozoic Khondalite Belt is probably wider than presently mapped, given that Late Paleoproterozoic sedimentary rocks and metamorphism have been revealed in the Ordos Block from drill core samples in recent years (Hu et al., 2012; Wan et al., 2013b). Irrespective of the width that is ultimately defined, however, the Khondalite Belt must represent a suture zone. Moreover, the southwest segment of the Khondalite Belt must have extended into other, now removed, parts of the Columbia Supercontinent (e.g., Zhao et al., 2004; Hou et al., 2008; Wang et al., 2014). Accordingly, the total length of the Khondalite Belt must have originally been greater than the present width of the North China Craton. The long and narrow features of the proposed slab break-off model can readily explain the syn- and/or post-collisional magmatism and metamorphism along the Khondalite Belt suture zone.

As previously noted, a model of ridge subduction has been proposed to account for the generation of the ca. 1.93–1.92 Ga magmatic rocks and UHT metamorphism in Jining Complex (Peng et al., 2010, 2011, 2012). Irrespective of the merits of this model for the Jining Complex, it does not account for the similar magmatism and metamorphism in the middle and southwestern parts of the Khondalite Belt, such as the ca. 1.92 Ga magmatic rocks in Helanshan Complex (Geng et al., 2009), and contemporaneous metamorphism in the Qianlishan and Daqingshan complexes (Yin et al., 2009; Ma et al., 2012). Moreover, any ridge subduction at ca. 1.93–1.90 Ga should be considered as a geodynamically distinct event from those processes associated with the generation of ca. 1.97–1.95 Ga magmatism and metamorphism along the whole 1000 km Khondalite Belt.

6. Conclusions

Based on our new data on the Helanshan Complex and previously published data for the Khondalite Belt of the North China Craton, we draw the following major conclusions:

(1) The ca. 1.95 Ga porphyritic K-feldspar granites and two-mica granites in the Helanshan Complex, North China Craton are the earliest S-type granites in the Khondalite Belt. They are peraluminous (A/CNK value >1.0), and characterized by high zircon $\delta^{18}O$ values of 7.3–10.6 ‰.
(2) In situ zircon Hf–O isotopic compositions indicate that the mantle not only supplied heat but also a magmatic component to the generation the S-type granites.
(3) A slab break-off model best explains the documented succession of events, i.e., ca. 1.97–1.76 Ga mafic magmatism, ca. 1.95–1.94 Ga mafic, carbonatitic, and S-type granitic magmatism and granulite metamorphism.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.precamres.2014.07.024.

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